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Palaeogeography, Palaeoclimatology, Palaeoecology 240 (2006) 318 – 372
PALAEO
www.elsevier.com/locate/palaeo

Global time scale and regional stratigraphic reference scales of Central and West Europe, East Europe, Tethys, South China, and North America as used in the Devonian–Carboniferous–Permian Correlation Chart 2003 (DCP 2003)

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Received 4 April 2005; accepted 24 March 2006

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doi:10.1016/j.palaeo.2006.03.058

Abstract

The boundaries of the Devonian, Carboniferous, and Permian stages of the *Global Stratigraphic Reference Scale* (abbreviated to *Global Stratigraphic Scale—GSS*) are described in relation to the biostratigraphic and/or lithostratigraphic units of the *Regional Stratigraphic Reference Scales* (abbreviated to *Regional Stratigraphic Scales—RSS)* of Central and West Europe, East Europe, Tethys, South China (eastern Tethys), and North America. In their type regions the boundaries of GSS units rarely coincide with those of homonymous RSS units. Moreover, the definitions of some RSS units have changed several times over the last decades, and subsequent misunderstanding of the stratigraphical significance of these changes has often introduced errors into proposed global correlation charts. The stratigraphic framework proposed in our global Devonian–Carboniferous–Permian Correlation Chart 2003 [DCP 2003 (Devonian–Carboniferous–Permian Correlation Chart 2003, Menning, M., Schneider, J. W., Alekseev, A. S., Amon, E. O., Becker, G., von Bitter, P. H., Boardman, D. R., Bogoslovskaya, M., Braun, A., Brocke, R., Chernykh, V., Chuvashov, B. I., Clayton, G., Dusar, M., Davydov, V. I., Dybova-Jachowicz, S., Forke, H. C., Gibling, M., Gilmour, E. H., Goretzki, J., Grunt, T. A., Hance, L., Heckel, P. H., Izokh, N. G., Jansen, U., Jin Y.-G., Jones, P., Käding, K.-Ch., Kerp, H., Kiersnowski, H., Klets, A., Klug, Ch., Korn, D., Kossovaya, O., Kotlyar, G. V., Kozur, H. W., Laveine, J.-P., Martens, Th., Nemyrovska, T. I., Nigmadganov, A. I., Paech, H.-J., Peryt, T. M., Rohn, R., Roscher, M., Rubidge, B., Schiappa, T. A., Schindler, E., Skompski, S., Ueno, K., Utting, J., Vdovenko, M. V., Villa, E., Voigt, S., Wahlman, G. P., Wardlaw, B. R., Warrington, G., Weddige, K., Werneburg, R., Weyer, D., Wilde, V., Winkler Prins, C. F., Work, D. M., 2004). Abschlußkolloquium DFG-Schwerpunktprogramm 1054: Evolution des Systems Erde während des jüngeren Paläozoikums im Spiegel der Sedimentgeochemie. Abstracts Univ. Erlangen, Germany, 2004, p. 43.] (herein abbreviated to DCP 2003, and cited as DCP, 2003 in references) is an attempt to reduce these errors.

The DCP 2003 is the stratigraphic base for Project 1054 of the Deutsche Forschungsgemeinschaft (DFG) ”The evolution of the Late Palaeozoic in the light of sedimentary geochemistry”. This composite time scale has been carefully balanced, as far as data allows, to remove unnecessary, artificial compression and expansion of time intervals, biozonations and depositional events. The ages selected in DCP 2003 are markedly different to those in the Geologic Time Scale 1989 [GTS 1989 (Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., Smith, D.G., 1990). A geologic time scale 1989. Cambridge Univ. Press, Cambridge.; Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., Smith, D.G., 1990. A geologic time scale 1989. Cambridge Univ. Press, Cambridge, pp. 1–263.] and in Gradstein and Ogg [Gradstein, F.M., Ogg, J., 1996. A Phanerozoic time scale. Episodes 19 (1/2), 3–4, insert.), whereas they are closer to those of the Geologic Time Scale 2004 [GTS 2004; Gradstein, F. M., Ogg, J.G., Smith, A.G., 2004. A Geologic Time Scale 2004. Cambridge Univ. Press, Cambridge, pp. 1–589.]. Mostly, the ages are rounded to the nearest 0.5 Ma in order to avoid estimates of questionable accuracy, whereas ages of 0.1 Ma in the GTS 2004 and their error bars of ±0.4 Ma to ±2.8 Ma for the Devonian to Permian stage boundaries suggest an improved accuracy. In contrast, in the DCP 2003 questionable ages and positions of stratigraphic boundaries are marked by arrows.

Keywords: Devonian; Carboniferous; Permian; Europe; Asia; North America; Time scale; Marine–terrestrial correlations

1. Introduction

At the beginning of Project 1054 of the Deutsche Forschungsgemeinschaft (DFG) “The evolution of the Late Palaeozoic in the light of sedimentary geochemistry” German geochemists were using time scales with significant age differences. Therefore, it was decided to use a uniform time scale and a stratigraphic correlation chart with sections from several continents. This was done to ensure variations in isotopic profiles of carbon, oxygen, sulfur, nitrogen, boron, and osmium would be directly comparable, and hence prove to be more reliable indicators of isochronous global events.

During 2001–2002 the “Stratigraphische Tabelle von Deutschland 2002” (herein abbreviated to STD 2002, and cited as STD, 2002 in references) was prepared by the Deutsche Stratigraphische Kommission (www.stratigraphie.de). At that time an updating of the global time scale was necessary in order to integrate isotopic age determinations that had become available over the previous decade since the publication of the previous Phanerozoic time scales (Harland et al., 1990; Odin, 1994; Gradstein and Ogg, 1996; Young and Laurie, 1996; IUGS, 2000), and the separate time scales for the Devonian (Tucker et al., 1998), Carboniferous (Menning et al., 2000), and Permian (Menning, 2001) periods.

The STD 2002 became the model for the Devonian–Carboniferous–Permian Correlation Chart 2003 (herein abbreviated to DCP 2003, and cited as DCP, 2003 in references), and in consequence, both charts have identical time scales. Version 2 and Version 4g of the DCP 2003 were presented as posters, respectively at the 15th International Congress on Carboniferous and Permian 2003 in Utrecht, and at the 32nd International Geological Congress 2004 in Florence. By August 2004, about 40 authors had contributed to more than 45 supraregional and regional columns, and to over 50 columns with marine and terrestrial biozonations from six continents. Work on several of these columns is still in progress. Both the DCP 2003 and the Explanations 2008 on the Devonian–Carboniferous–Permian Correlation Chart 2003 should be available in 2008.

As a major part of the DCP 2003 its *Global Stratigraphic Scale (GSS)*, the numerical ages of the stages, and (part of) *Regional Stratigraphic Scales* (*RSS*; supraregional composite sections) of Central and West Europe, East Europe, Tethys, South China (eastern Tethys), and North America are presented here for the first time. It is most important to emphasize there are significant differences between homonymous global and supraregional stratigraphic units which must be taken into account, when time-related geochemical data and events of the DFG-Project 1054 are compared on a global scale. For the East European Platform, the DCP 2003 presents the Resolutions (1990a,b) ("Stratigraphic Guide" of the Soviet Union 1990) for the Carboniferous and Permian periods, and additional terms where necessary. The data used in the DCP 2003 are published or will be published in near future. Arrows indicate questionable ages and positions of stratigraphic boundaries. Frequently used abbreviations are:

CPB	*C*arboniferous–*P*ermian *b*oundary
DCB	*D*evonian–*C*arboniferous *b*oundary
DCP 2003	*D*evonian–*C*arboniferous–*P*ermian Correlation Chart 2003 (will be available in 2008)
FAD	*F*irst *A*ppearance *D*atum
FOD	*F*irst *O*currence *D*atum
GSS	*G*lobal *S*tratigraphic *S*cale
GSSP	*G*lobal *S*tratotype *S*ection and *P*oint
ICS	*I*nternational *C*ommission on *S*tratigraphy
IUGS	*I*nternational *U*nion of *G*eological *S*ciences
Mts.	Mountains
PTB	*P*ermian–*T*riassic *b*oundary
RSS	*R*egional *S*tratigraphic *S*cale
SCCS	*S*ub*c*ommission on *C*arboniferous *S*tratigraphy
STD 2002	*S*tratigraphische *T*abelle von *D*eutschland 2002 (Stratigraphic Table of Germany 2002)

Uncommon stratigraphic terms and examples for the derivation of endings of stratigraphic terms are listed in Table 1.

2. Devonian–Carboniferous–Permian time scales

2.1. The global time scale of the DCP 2003/STD 2002 (M. Menning, P.J. Jones, K. Weddige, D. Weyer)

The numerical calibration of the Global Stratigraphic Scale (GSS) as used in the DCP 2003 is identical to that of the STD 2002. The DCP 2003 time scale is composed of time scales for three periods: Devonian (Weddige et al., 2005), Carboniferous (Menning et al., 2000), and Permian (Menning, 2001). These time scales are numerically calibrated to portray geological sequences and events in a linear time frame. Every endeavour has been made, as far as data allows, to remove unnecessary, artificial compression and expansion of time intervals, biozonations and depositional events. The numerical ages of the stages are mostly rounded to the nearest 0.5 Ma.

The accepted and proposed Global Stratotype Section and Points (GSSP) and their index fossils are shown for the DCP 2003. For stages yet to be defined by a GSSP, traditional index fossils are used, as far as possible, in order to establish their boundaries for global correlation. An exception is the base of the Serpukhovian Stage, which is traditionally defined by foraminifers and ammonoids, but when defined by conodonts, lies stratigraphically slightly below the traditional boundary (i.e. the use of the FAD of *Lochriea ziegleri* leads to the referral of the latest Late "Visean" to the Serpukhovian).

In the DCP 2003 time scale, as in most time scales, the Middle Devonian is considerably shorter than the Early and the Late Devonian epochs. The Emsian and Famennian are the longest of the Devonian stages. The DCP 2003 age of ca. 358 Ma for the Devonian–Carboniferous boundary (DCB) is a compromise between the ca. 354 Ma age, based on U–Pb SHRIMP dates (Claoué-Long et al., 1993), and the ca. 362 Ma age, based on U–Pb ID-TIMS dates (Tucker et al., 1998). Recently, an age of 360.7±0.7 Ma has been derived for the DCB from U–Pb ID-TIMS dating of two metabentonites (360.2±0.7 Ma for bed 70, Early *Siphonodella duplicata* Zone, and 360.5±0.8 Ma for bed 79, late part of the *Siphonodella sulcata* Zone) taken from the Hasselbach auxiliary global stratotype section in the Rhenish Slate Mts. (Rheinisches Schiefergebirge) (Trapp et al., 2004). Using the probable FAD of *S. sulcata* in bed 84 of this section, and its thickness, the age of the DCB is estimated at ca. 361.4±0.7 Ma (Menning, this work; Fig. 1, Table 2).

The 320 Ma age for the Mississippian–Pennsylvanian boundary is based on the $^{40}Ar/^{39}Ar$ date of 324.6 Ma for the Jaklovec Member in Upper Silesia (Lippolt et al., 1984), close to the Pendleian–Arnsbergian boundary (intra-Namurian A) (Menning et al., 2000: Fig. 6). The $^{40}Ar/^{39}Ar$ age of 319.5 Ma for the Poruba Member (Lippolt et al., 1984), within the Middle Arnsbergian (Menning et al., 2000: Fig. 6), is indicative of a younger age for this boundary, and a ca. 318 Ma is

Table 1
Uncommon stratigraphic terms and examples for the derivation of stratigraphic terms used in the DCP 2003

Geographical source term	English basic term	Usage in the DCP 2003	Very common synonym I	Less common synonym II
Indus (Engl.)	Indus	INDUSIAN	Induan	“Ind*ian*”
Татары (Russ.)	Tatar*y*	Tatarian R. St.	Tatar*sky*	Tatar*sk*ian
Арти (Russ.)	Art*i*	Art*insk*ian	Art*insky*	–
Сакмара (Russ.)	Sakmar*a*	SAKMARIAN	Sakmar*sky*	Sakmar*sk*ian
Стерлитамак (Russ.)	Sterlitamak	Sterlitamakian R. SSt.	Sterlitamak*sky*	Sterlitamak*sk*ian
Тастуба (Russ.)	Tastub*a*	Tastubian R. SSt.	Tastub*sky*	Tastub*skian*
Гжель (Russ.)	Gzhel	GZHELIAN	Gzhel*sky*	Gzelian
Москва (Russ.)	Mosco*w*	MOSCOVIAN	Moskov*sky*	Moskov*sk*ian
Stéfanus (Lat.)	Stefan	Stephanian R. St.	Stefanian	Stéphanien
Westfalen (Germ.)	Westphal*ia*	Westphalian R. St.	–	West*f*alian
Башкирия (Russ.)	Bashkir*ia*	BASHKIRIAN	Bashkir*sky*	Bashkir*skian*
Серпухов (Russ.)	Serpukhov	SERPUKHOVIAN	–	Serpu*chow*ian
Visé (Franz.)	Vis*e*	VIS*EAN*	–	“Vis*ian*”
Atoka (Amer.)	Atok*a*	Atok*an* R. St.	–	Atok*ian*”
Morrow (Amer.)	Morrow	Morrow*an* R. St.	–	“Morrow*ian*”
Osage (Amer.)	Osag*e*	Osag*ean* R. St.	–	“Osag*ian*”
Přídolí (Czech)	Pridol*i*	PRIDOLIAN	–	–
Yanguan (Chi.)	Yanguan	Yanguanian R. Ser.	–	Aikuan*ian*

GLOBAL STAGE in capital letters.
R. St.—Regional Stage, with ending *-ian*.
R. SSt.—Regional Substage, with ending *-ian*.
Horizon—Regional Stage or Regional Substage, with ending *-ian*.
Svita—mainly Formation, *without* ending *-ian*.

suggested (this work). The latter age is more consistent with the U–Pb ID-TIMS ages from the Visean of Trapp (pers. com.).

The Mississippian/Early Carboniferous (Tournaisian–Serpukhovian) is considerably longer than the Pennsylvanian/Late Carboniferous (Bashkirian–Gzhelian). For a long time the duration of the Pennsylvanian was overestimated, mainly because of the tremendous thicknesses of coal-bearing deposits in the United States and Europe. In the DCP 2003 time scale, the time relationship of Mississippian to Pennsylvanian is 38 my to 24 my. When the boundary ages of ca. 361 Ma (Devonian–Carboniferous), 319 Ma (Mississippian–Pennsylvanian) and 299 Ma (Carboniferous–Permian) are used, the Mississippian (ca. 42 my) is about twice as long as the Pennsylvanian (ca. 20 my).

The most significant shift of a numerical age of the Permian is at the top of the Kungurian Stage (top Cisuralian Epoch). The ages suggested for this boundary have swung from 255–256 Ma (Harland et al., 1982, 1990; Gradstein and Ogg, 1996) to 274–270 Ma (Menning, 1989, 1995a,b; Young and Laurie, 1996; Gradstein et al., 2004), despite the upward expansion of the Kungurian of the GSS at the expense of of the Ufimian Stage (cf. 5.1, 5.2.3).

The age of the Permian–Triassic boundary (PTB) has been changed from 245 Ma (Harland et al., 1982) to 251 Ma (Menning, 1995a,b; Young and Laurie, 1996; Gradstein et al., 2004) based on the $^{40}Ar/^{39}Ar$ dating of the Grenzbitumen-Zone of 233 ± 9 Ma (Hellmann and Lippolt, 1981) and the U–Pb SHRIMP age of 251.2 ± 3.4 Ma for bed 25 in the GSSP of Meishan which is only slightly older than the PTB (Claoué-Long et al., 1991). Using the U–Pb ID-TIMS age of 252.6 ± 0.2 Ma for bed 25 in Meishan (Mundil et al., 2004), an age of 252.5 Ma has been derived for the PTB (Fig. 1; Menning et al., 2005).

2.2. Time scales 1989–2006 (M. Menning)

The STD 2002 (Menning and Deutsche Stratigraphische Kommission, 2002)/DCP 2003 time scales differ only slightly from the Geologic Time Scale 2004 (herein abbreviated as GTS 2004, and cited as GTS 2004; Gradstein et al., 2004). However, they all are in marked contrast to those of Haq and Van Eysinga (1987), Geologic Time Scale 1989 (herein abbreviated to GTS 1989, and cited as GTS, 1989 in references), (Harland et al., 1990), Odin (1994), and Gradstein and Ogg (1996). Comparing the DCP 2003 and the GTS 2004, the maximum age differences of stage boundaries are 4.6 Ma in the Permian, 1.9 Ma in the Carboniferous and 7.6 Ma in the Devonian. It is difficult to assess which of the stage durations (Table 2, Fig. 1) are the

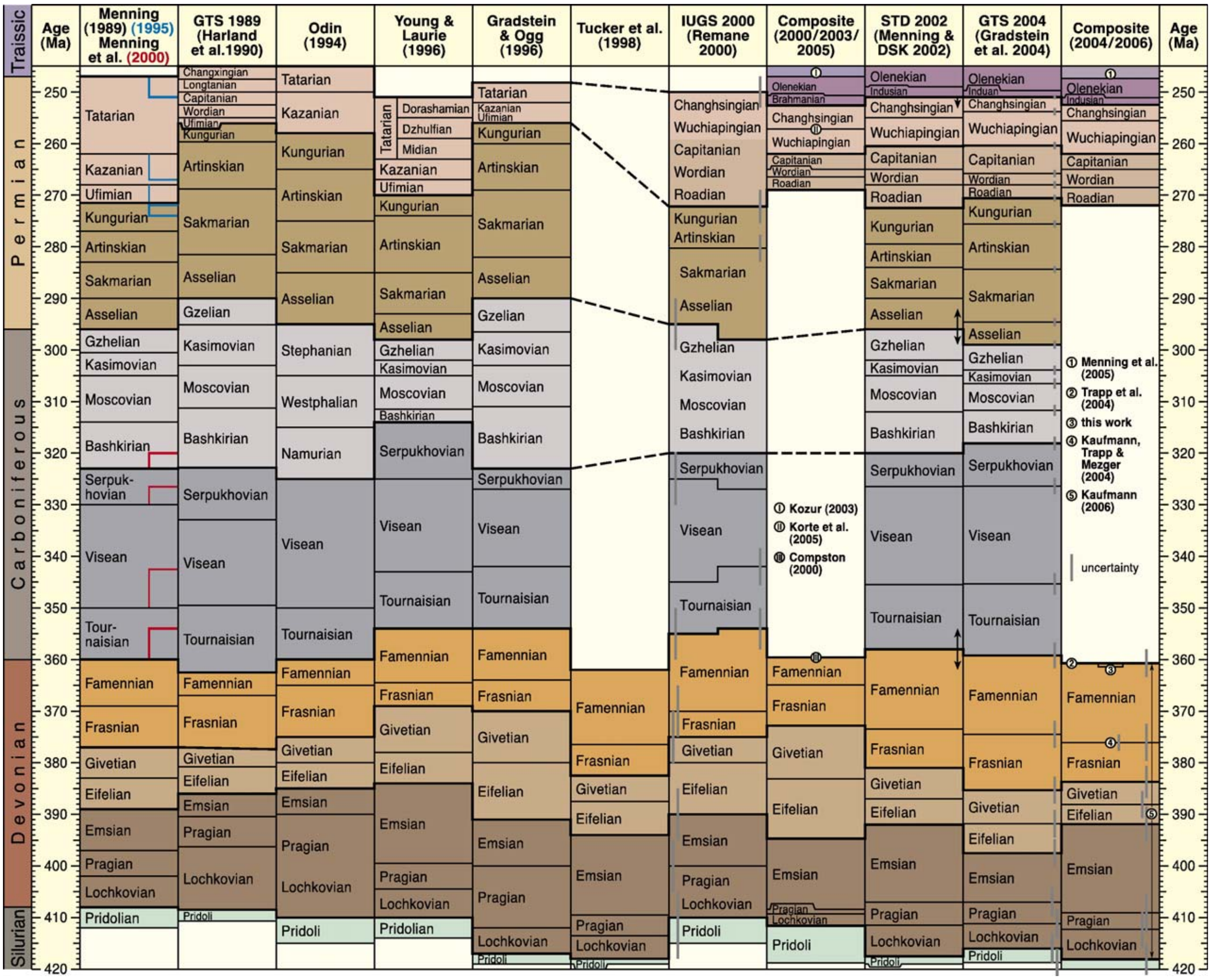

Fig. 1. The Devonian, Carboniferous, and Permian periods in time scales 1989–2006.

more reliable, e.g. for the Sakmarian (6 my in DCP 2003, or 10.2 my in GTS 2004); for the Emsian (15 my in DCP 2003, 9.5 my in GTS 2004, or 17 my in Kaufmann, 2006). Since the publication of the GTS 1989 (Harland et al., 1990) time scale, the most significant changes in age for the Devonian–Permian time interval are for: the Silurian–Devonian boundary from 408 Ma to 416–418 Ma, the Dinantian–Silesian (Visean–Namurian) boundary from 332.9 Ma to 326.5–325 Ma, the Carboniferous–Permian boundary from 290 Ma to 296–299 Ma, the Early–Late/Middle Permian boundary from 256–258 Ma to 270–274 Ma, and for the Permian–Triassic boundary from 245 Ma to 251–252.5 Ma (cf. Fig. 1, Table 2; Menning, 1989).

In the DCP 2003 questionable ages and positions of stratigraphic boundaries are marked by arrows, rather than by error bars. The ages of the DCP 2003 are mostly rounded to the nearest 0.5 Ma in order to avoid estimates of questionable accuracy, whereas ages of 0.1 Ma in the GTS 2004 and their small error bars of ±0.4 Ma to ±2.8 Ma for the ages of the Devonian to Permian stages suggest an improved accuracy, however these is not well documented (Fig. 1). For example, comparing ages for the global Emsian–Eifelian boundary (FAD of *Eucostapolygnathus partitus*) of 397.5±2.7 Ma (GTS 2004) and 391.9±3.4 Ma (Kaufmann, 2006) the error bars (±2σ) do overlap hardly (Tables 2 and 3).

3. Devonian GSSPs and Regional Stratigraphic Scales

3.1. General (K. Weddige, P.J. Jones)

The Devonian System was introduced by Sedgwick and Murchison (1839). There was, however, no clear

Table 2
Numerical ages of geologic time scales 1989–2006 (Devonian to Permian)

Stage	Menning (1989)	GTS 1989, Harland et al. (1990)	Young and Laurie (1996)	Gradstein and Ogg (1996)	Tucker et al. (1998)	Composite (2000/2003/ 2005)	STD 2002 (Menning and DSK 2002)	GTS 2004 (Gradstein et al., 2004)	Composite (2004/ 2005)	Stage
Scythian	Scythian	Scythian	Griesbach.	Induan		Brahman.	Indusian	Induan	(1)	Indusian
	247–	245–	251–	248.2 ±4.8		252.6 (I)	251+2.0	251.0 ±0.4	~252.5	
		Changhs.	Dorasham.			4.6	4	2.8	~3.1	Changhs.
Tatarian	15	247.5–	255–	3.9		257.2 (II)	255–	253.8 ±0.7	~255.6	
		Longtan.	Dzhulfian			4.8	5.5	6.6	~6.4 (6.9)	Wuchiap.
		250–	259–			262	260.5–	260.4 ±0.7	~262.0	
		Capitanian	Midian			3	4.5	5.4	~3.0 (~4.0	Capitan.
	262–	252.5–	263–	252.1–		265	265–	265.8 ±0.7	~265.0	
Kazanian	6	Wordian	Kazanian	Kazanian		1.5	3	2,2	~3.5	Wordian
	268–	255–	267–	–		266.5	268–	268.0 ±0.7	~268.5	
Ufimian	3.5	Ufimian	Ufimian	Ufimian		2.5	4.5	2.6	~3.5	Roadian
	271.5–	256.1–	270–	256–		269	272.5–	270.6 ±0.7	~272.0	
Kungur.	5.5	3.6	4	4			7	5.0		Kungur.
	277–	259.7–	274–	260–			279.5–	275.6 ±0.7		
Artinsk.	6	9.1	11	9			4.5	8.8		Artinsk.
	283–	268.8–	285–	269–			284–	284.4 ±0.7		
Sakmar.	7	12.7	8	13			6	10.2		Sakmar.
	290–	281.5–	293–	282–			290–	294.6 ±0.8		
Asselian	6	8,5	5	8			6	4,4		Asselian
	296–	290.0–	298–	290–			296 −4, +3	299.0 ±0.8		
Gzhelian	4.5	5.1	4	6.5			6	4.9		Ghzelian
	300.5–	295.1–	302–	296.5–			302–	303.9 ±0.9		
Kasimov.	4.5	7.9	3	6.5			3	2.6		Kasimov.
	305–	303–	305–	303–			305–	306.5 ±1.0		
Moskov.	9	8.3	6.5	8			7	5.2		Moskov.
	314–	311.3–	311.5–	311–			312–	311.7 ±1.1		
Bashkir.	9	11.5	2.5	12			8	6.4		Bashkir.
	323–	322.8–	314–	323–			320–	318.1 ±1.3		
Serpukh.	7	10.1	11	4			6.5	8.3		Serpukh.
	330–	332.9–	325–	327–			326.5–	326.4 ±1.6		
Visean	20	16.6	18	15			19	18.9		Visean
	350–	349.5–	343–	342–	–	–	345.5–	345.3 ±2.1		
Tournai.	10	13	11	12		(III)	12.5	13.9	(5)	Tournai.
	360–	362.5–	354–	354	362–	359.6–	358 ±4	359.2 ±2.5	360.7 ±0.7	(2)
									361.4 ±0.7	(3)
Famenn.	9	4.5	10.5	10	14.5	5.3	15.5	15.3	15.3	Famenn.
	369–	367–	364.5–	364–	376.5–	364.9–	373.5–	374.5 ±2.6	376.1 ±1.6	
									376.1 ±1.6	(4)
Frasnian	8	10.4	4.5	6	6	8.0	7.5	10.8	7,7	Frasnian
	377–	377.4–	369–	370–	382.5–	372.9–	381	385.3 ±2.6	383.7 ±1.1	
Givetian	6	3.4	9	10	5	10.3	6	6.5	4,4	Givetian
	383–	380.8–	378–	380–	387.5–	383.2–	387–	391.8 ±2.7	388.1 ±0.6	
Eifelian	6	5.2	6	11	6.5	11.6	5	5.7	3.7	Eifelian
	389–	386.0–	384–	391–	394–	394.8–	392–	397.5 ±2.7	391.8 ±0.4	
Emsian	8	4.4	15.5	9	15.5	13.7	15	9.5	18.0	Emsian
	397–	390.4–	399.5–	400–	409.5–	408.5–	407–	407.0 ±2.8	409.8 ±0.3	
Pragian	5	5.9	4	12	4	0.9	4.5	4.2	2.9	Pragian
	402–	396.3–	404.5–	412–	413.5–	409.4–	411.5–	411.2 ±2.8	412.7 ±0.5	
Lochkov.	6	12.2	5.5	5	4.5	2.2	6	4.8	5.4	Lochkov.
	408–	408.5–	410–	417–	418–	411.6–	417.5–	416.0 ±2.8	418.1 ±1.1	
Pridolian	4	2.2	4	2	1	7.2	1.5	2.7		Pridolian
	412–	410.7–	414–	419–	419–	418.8–	419–	418.7 ±2.7		

(I) 252.6 Ma: Kozur (2003); (II) Permian stages: Korte et al. (2005); (III) Devonian stages: Compston (2000).
(1) Roadian to Changhsingian stages, Menning et al. (2005); (2) 360.7±0.7 Ma, Trapp et al. (2004); (3) 361.4 this work, according to Trapp et al. (2004); (4) 376.1±1.6 Ma, Kaufmann et al. (2004); (5) Devonian stages: Kaufmann (2006).

Table 3
Underestimated uncertainties of period and stage boundaries

Jurassic–Cretaceous	130±3 Ma (Odin, 1982)	144±5 Ma (Harland et al., 1982)
Triasssic–Jurassic	205.7±4 Ma (Gradstein and Ogg, 1996)	199.6±0.6 Ma (Gradstein et al., 2004)
Carnium–Norium	227–228 Ma (Muttoni et al., 2004)	216.5±2.0 Ma (Gradstein et al., 2004)
Permian–Triassic	≤252.6±0.3 Ma (Mundil et al., 2004)	251.0±0.4 Ma (Gradstein et al., 2004)
Emsian–Eifelian	391.9±3.4 Ma (Kaufmann, 2006)	397.5±2.7 Ma (Gradstein et al., 2004)

definition of the lower boundary in the Devonshire type area and no definite correlation between its marine facies and the terrestrial Old Red Sandstone which overlies Silurian rocks in Wales. In the Welsh Borderland, the Ludlow Bone Bed was accepted as a convenient lithological marker for the base of the Devonian. Here the last appearance of graptolites in the late Ludlow (Elles and Wood, 1901–1918), below the Ludlow Bone Bed, was taken to indicate the Silurian–Devonian boundary. By 1960, it was realized that this disappearance was a local event, and that the global extinction of graptolites occurred in the Early Devonian. A globally recognizable Silurian–Devonian boundary was accepted by the ICS in 1972. By 1996, all stage boundaries and GSSPs had been recommended by the Subcommission on Devonian Stratigraphy, and ratified by the International Commission on Stratigraphy (cf. Ogg, 2004; Fig. 2). All stage boundaries are defined by conodonts except for the base of the Devonian, which is indexed by a graptolite.

There are age differences between the original and the newly defined stage boundaries, but the same names are used. An excellent example of stage homonyms is the Emsian, which was used historically for the shallow water facies, lacking conodonts, of Germany. However, the currently used Emsian is defined in a conodont rich pelagic facies of Central Asia. Thus, a strict stratigraphic terminology must always differentiate between a global Emsian ("GSSP Emsian") and a regional Emsian ("historical Emsian"). The same is true for most of the other Devonian stages (cf. Fig. 2).

Moreover, the GSSP concept based strictly on the phylogenetic entry of the chosen index species which is usually more uncertain than the position of a "Golden Spike" in the stratotype section, because the index fossil either makes an imperceptible phylogenetic appearance and/or is too rare to detect a FOD. The recent restudy of the Pragian GSSP (see below) provides a good example for such uncertainties. The scientific and, even political, problems involved with the subdivision of the Devonian Period by GSSPs, appear to have been resolved, and thus, the Devonian subdivision has still an official directive status, but with a less flexible relationship to current research.

On the one hand, index conodonts dominate Devonian stratigraphy, and because of their widespread practical use, the standard global conodont zonation has to be regarded as the paramount subdivision of the Devonian. This zonation evolves iteratively by scientific updates and refinements in accordance with current research, and which is, for instance, regularly documented and updated in the Devonian Correlation Table (Weddige, 1996–1998, 2000–2005).

On the other hand, the stage subdivision was originally based on remarkable facies changes, which causes distinct faunal changes, often named "events" (e.g. Walliser, 1985; see below for details). These events often permit a precise lithostratigraphic correlation. Thus, it is also important aim to analyse such events by inorganic geophysical and geochemical criteria for their potential as additional time markers.

3.2. Lochkovian, Pragian, and Emsian stages (Early Devonian Epoch)

3.2.1. GSSPs and numerical ages (K. Weddige, M. Menning)

The basal Devonian GSSP, and the co-incident base of the Lochkovian Stage, is in the Lochkov Formation at the Klonk section southwest of Prague, Czech Republic, where it is recognized by the FOD of the graptolite *Monograptus uniformis* and of the trilobite group of *Warburgella rugulosa*. This section consists of a rhythmic sequence of allochthonous limestones and autochthonous shales, where the "sudden and abundant occurrence" of *M. uniformis* does not necessarily demonstrate reworking. Confirmed by the occurrences of associated fossils, e.g. of trilobites, conodonts, or chitinozoans and spores, the boundary level can be recognized in most parts of the world (Chlupáč and Hladil, 2000).

The base of the *Pragian Stage* is defined by the GSSP in the neritic Praha Formation at Velka Chuchle, near Prague, Czech Republic (Weddige, 1987; Chlupáč and Oliver, 1989; Chlupáč, 2000) by the FOD of the conodont *Eognathodus sulcatus*. However, this index conodont has recently been found one layer below the GSSP (Slavík, 2004). This demonstrates the use of a single index fossil to indicate a time level leads to uncertainty, because its entry depends largely on palaeoenvironmental factors, i.e. on local facies. Such uncertainty could be reduced by considering the use of

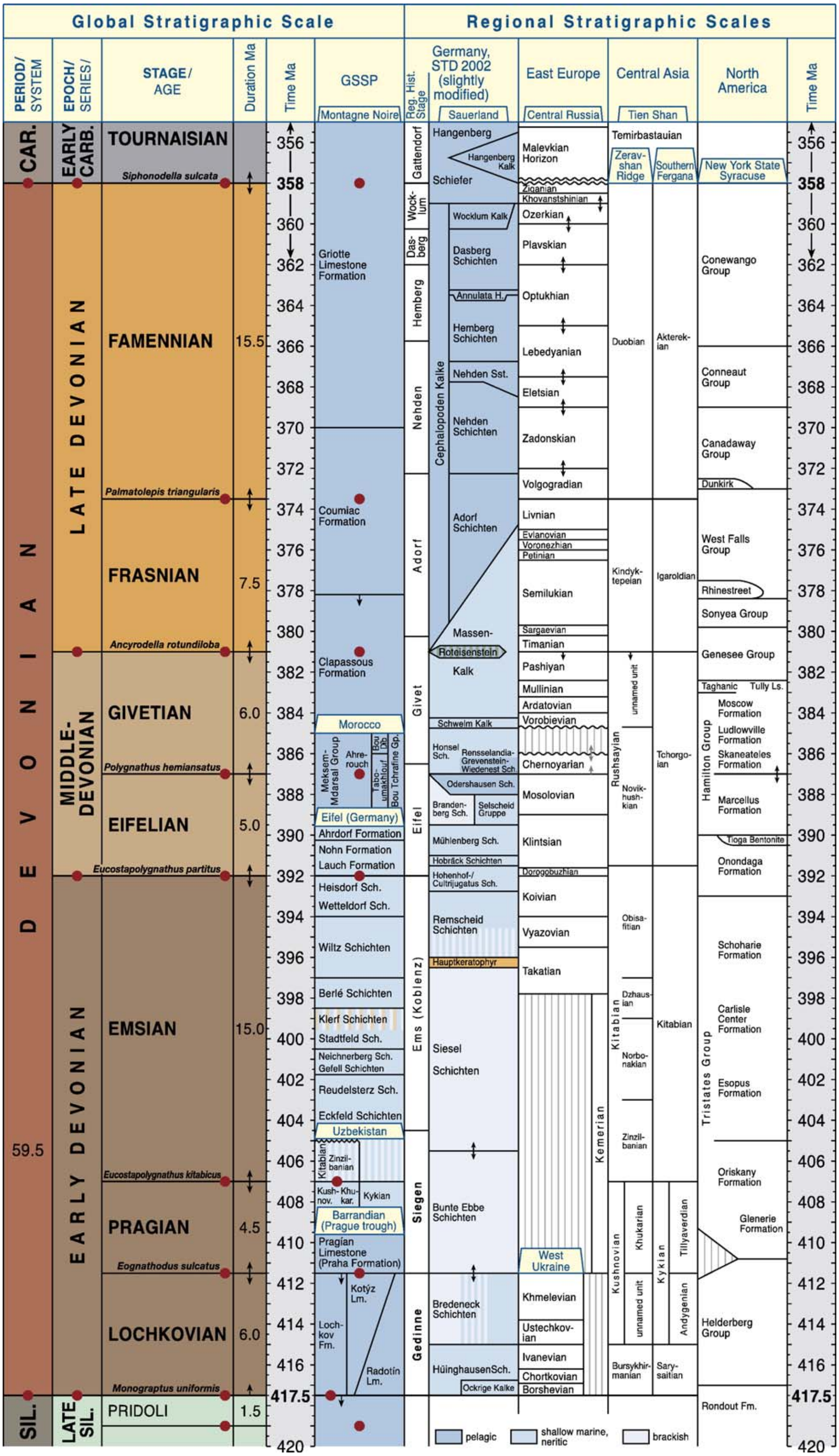

Fig. 2. Devonian Global Stratigraphic Scale as used in the DCP 2003 and Regional Stratigraphic Scales.

associated guide fossils and by interfacial correlations. Thus, the accompanying conodont *Latericriodus steinachensis* appears just below the GSSP boundary. The dacryoconarids *Nowakia sororcula* and *Nowakia acuaria* enter shortly above the boundary, and chitinozoans are useful because they can be linked to the spore zonation. Lithologically, the early Pragian rocks become gradually lighter in colour, although their overall facies remains similar to that of the Lochkovian (Chlupáč, 2000). A relatively minor faunal change in the early Pragian marks the rather insignificant Lochkovian–Pragian Event.

The GSSP for the Emsian Stage is within the Khodzha–Kurgan Formation of the Zinzilban Gorge of the Kitab National Park in Uzbekistan (Yolkin et al., 1997; Yolkin et al., 2000). The boundary is indexed by the conodont *Eocostapolygnathus kitabicus* which is derived, and well differentiated, from its ancestor *Eoctenopolygnathus pireneae*, and thus represents a proven biostratigraphic time marker (Yolkin et al., 1994). The level of this phylogenetic transition is lower than the entry of the somewhat controversial taxon *Eolinguipolygnathus dehiscens*, and the entries of both taxa are lower than the top of the regional historical Pragian Stage. The rarity of conodonts in the late Pragian of the type area, however, does not permit an exact correlation with the Emsian GSSP stratotype in Uzbekistan.

When Yolkin et al. (1994) proposed such a low Emsian base at the expense of the historical Pragian Stage, the authors intended to introduce a further stage for the latest Early Devonian. The early interval of the global Emsian corresponds with a part of the late original Pragian Stage in its Czech type region, and with the so-called Zinzilbanian in Uzbekistan, Central Asia (Fig. 2). For the middle and late intervals of the global Emsian, the Czech terms Zlichovian and Dalejan are currently used informally when open marine to pelagic guide fossils like conodonts, dacryoconarids, trilobites, and goniatites are available. But, when the latter are lacking and only shallow water forms of brachiopods and trilobites are available, the terms of the neritic Rhenish facies, Early and Late Emsian substages, have to be used. The historical base of the original Early Emsian has been correlated, a century later, with the base of the Ulmen Substage i.e., the Eckfeld Schichten (Fig. 2), which may be slightly younger than the base of the Zlichovian (cf. Weddige, 1996). The base of the Late Emsian, in contrast, is more or less coeval with the base of the Dalejan.

With regard to the event stratigraphy, this subdivision of the global Emsian is supported by the Basal Zlichov Event (regressive pulse) and the (basal) Daleje Event (transgressive pulse). Both events are now regarded as more important than the basal Pragian Event, because they are recognized globally.

The *numerical age* for the base of the *Lochkovian* Stage (base of the Devonian Period) was estimated by Tucker et al. (1998) at 418 Ma using three ID-TIMS ages from the Lochkovian and Pragian stages and the Kalkberg age of 417.6 ± 1.9 Ma from the upper half of the *Icriodus woschmidti* conodont Zone (earliest Lochkovian; Kaufmann, 2006). Using the same ages as Tucker et al. (1998), and part of their error bars, an age of 417.5 Ma has been derived for the DCP 2003. Kaufmann (2006) favours an age of 418.1 ± 3.0 Ma which may have been influenced at least from the Early Emsian Bundenbach ID-TIMS age of 407.7 ± 0.7 Ma of Kaufmann et al. (2005) (cf. Fig. 1). House and Gradstein (2004: Fig. 14.5; GTS 2004) estimated an age of 416 Ma using the Ludfordian Upper Whitcliffe ID-TIMS age of 420.2 ± 3.9 Ma (Tucker et al., 1998) as tie point, giving the other ages a lower weight. Compston (2000) suggested a SHRIMP age of 409.9 ± 1.1 Ma for the base of the Devonian, after he reprocessed original data, allowing for the heterogeneity in the zircon SL13 standard. His entire Devonian time scale, except for the upper limit, is in strong contrast to the other scales based mainly on ID-TIMS ages (Fig. 1).

Since the work of Tucker et al. (1998), the Emsian Stage has been thought to have a duration between 15 Ma and 17 Ma, except for the House and Gradstein (2004: Fig. 14.5; GTS, 2004) scale. The latter combines three isotopic ages and the “House scale” which “juggles” the high-resolution ammonoid and conodont (sub) zonations, such that shortest segments are fairly close to equal duration” (House and Gradstein, 2004: 215). This methodology appears to be flawed by the exclusion of some very significant ID-TIMS ages in its calibration procedure. Thus, the age of the upper boundary of the Emsian Stage at 397.5 ± 2.7 Ma may be too old and its duration of 9.5 Ma is probably too short. Most probably, the Emsian Stage is longer than the combined ages of the Lochkovian and Pragian stages (Fig. 1).

3.2.2. Germany (K. Weddige)

Two thirds of the German part of the Rhenish Massif is formed by Early Devonian rocks, which consist of mainly siliciclastic sediments in a neritic shallow-water facies (“Rhenish Facies“). The fauna is dominated by mostly brachiopods and trilobites, whereas (hemi) pelagic global index fossils such as conodonts, goniatites, tentaculites, and graptolites are lacking. Thus, the global stages Lochkovian, Pragian, and Emsian are rarely recognizable, and the regional historical stage subdivision of Gedinne, Siegen, and Ems has to be maintained. The siliciclastic material originated from the NW, from the "Old Red Continent" (part of the northern supercontinent Laurussia), and was accumulated in the Rhenish trough. Accumulation rate and crustal subsidence of the shelf area were both high suggesting a constant palaeobathymetry so that the entire Early Devonian succession became up to 10,000 m thick. The strong subsidence initiated crustal movements, and volcanics sporadically intercalated in the sediments, e.g. the Hauptkeratophyr (Fig. 2). Facies shifts can be observed in sedimentary sequences from lacustrine, brackish, limnic-fluviatile to marine shallow-water (Fig. 2: Bunte Ebbe-Schichten and Siesel-Schichten). The stratigraphy of these Early Devonian successions is still under review. It is suspected, however, that the successions are only local developments, and thus do not reflect extra-regional causes like climatic changes. Even at the end of the Early Devonian, the late Emsian strata are generally transgressive, obviously related to a widespread deepening of the shelf area and/or even with a global rise of the sea level and climatic warming (Fig. 2: Wiltz Schichten and Heisdorf Schichten).

3.2.3. East Europe (A.S. Alekseev, B.I. Chuvashov)

During the Early Devonian most of the East European (or Russian) Platform was an area of subaerial erosion, and only on its southwestern margin (Lvov-Volyn, Ukraine) carbonate marine sedimentation took place (Alekseev et al., 1996). There, the base of the Lochkovian coincides with the lower boundary of the Borshevian Horizon with the FAD of the conodont *I. woschmidti* together with *M. uniformis angustidens* (Drygant, 1984). The Late Lochkovian, Pragian, and lower Emsian (Kemerian) are represented by a lagoonal and terrestrial succession in the Lvov-Volyn and Baltic area (Latvia and Lithuania). Their ages are based on evidence from thelodont and acanthodian fish zonation and spore assemblages (Avkhimovich et al., 1993).

The first Devonian marine transgression is of late Emsian age. Its carbonate and terrigenous sediments are known mainly on the eastern part of the East European Platform (Volga-Ural, western slope of the Ural Mts.). They contain conodont assemblages of the *E. dehiscens, Eocostapolygnathus gronbergi*, *Eucostapolygnathus inversus–Linguipolygnathus serotinus*, and *Eucostapolygnathus patulus* zones. The ages of the Takata and Vanyshkinsk formations are defined by miospores and brachiopods. The Koivinsk and Biysk formations are composed by shallow water argillites, marls, and limestone (Koivinsk Formation) and essentially shallow water limestones (Biysk Formation). Their ages are defined by conodonts, corals, and brachiopods.

3.2.4. Central Asia (N.G. Izokh)

Early Devonian deposits of the Zeravshan Ridge (South Tien Shan) are represented by shallow water reef and marly limestones with diverse benthic corals, brachiopods, trilobites, and other faunal groups, as well as remains of pelagic organisms: graptolites, dacryoconarids, and conodonts (Kim et al., 1984, 1988; Yolkin et al., 2000). The base of the Lochkovian coincides with sharp lithological changes at the lower boundary of the Bursykhirmanian and is defined by the appearance of *M. uniformis*, together with the trilobite *W. rugulosa*, and the brachiopods *Protathyris praecursor* and *Lanceomyonia borealiformis*. The Pragian massive limestones of the Khukarian are characterized by benthic Koneprusy-type faunas and conodont assemblages of the *E. sulcatus, Pseudogondwania kindlei*, and *E. pireneae* zones. The GSSP for the basal Emsian boundary is defined at the base of the *E. kitabicus* Zone (Yolkin et al., 1989, 1994). This boundary is aligned with the base of the Kitabian that embraces the Emsian *E. kitabicus*, *E. excavatus*, *Eolinguipolygnathus nothoperbonus, E. inversus, L. serotinus*, and *E. patulus* conodont zones and the Eifelian *E. partitus* Zone. The Early Devonian of South Fergana is represented by more deep water carbonates and fine clastics that include such index graptolite species as *M. uniformis*, *M. praehercynicus*, and *M. hercynicus* (Kim et al., 1988).

3.2.5. North America (K. Weddige)

The New York State succession is regarded as the classical standard of the North American Devonian, and of the method of sequence stratigraphy (e.g. Brett and Ver Straeten, 1997).

The Rondout Formation, which includes the Silurian–Devonian boundary, in eastern New York represents a subtidal nearshore marine environment. It forms the base of the Helderberg Group in which carbonates of intertidal facies dominate. Then, a transgressive period followed with the widespread and mature shallow marine Oriskany Sandstone of the Early Devonian Tristates Group. Both the Oriskany Sandstone and the approximately coeval Glenerie Formation, represent relatively shallow marine, shore to shelf facies. The overlying Esopus and Carlisle Center formations, restricted to eastern New York, are deeper marine grey to black carbonate mudstones. A subsequent change-over to increasingly carbonate-dominated deposition (Schoharie Formation) culminated in widespread limestones of the partly coeval Onondaga Formation. The change from carbonate to siliciclastic deposition followed by a gradual return to more carbonate-dominated rocks characterizes the first of three to four pulses of the Acadian Orogeny in eastern North America (Tectophase I sensu Ettensohn, 1985, 1987). Volcanic ash beds indicating crustal movements are found in both the Helderberg and Tristates groups and in the Onondaga Formation.

3.3. Eifelian and Givetian stages (Middle Devonian Epoch)

3.3.1. GSSPs and numerical ages (K. Weddige, M. Menning)

The base of the Middle Devonian Epoch, and the coincident base of the Eifelian Stage, is defined by a GSSP within the uppermost Heisdorf Formation at Wetteldorf Richtschnitt in the Prüm Syncline of the Eifel District of western Germany (Ziegler and Werner, 1982). The boundary is defined by the entry of the conodont *E. partitus* (see Weddige, 1982). The stratotype section shows a gradual increase of open marine facies, that may have influenced the entry of the index conodont. The taxon, however, represents a clear phylogenetic step within the *Eucostapolygnathus* lineage which, although less precise than the Golden Spike of the GSSP, can be recognized world-wide (Ziegler, 2000). The transgression culminates slightly higher within the earliest Eifelian that corresponds with the coeval anoxic pulse of the Choteč Event in the Prague area.

The base of the global Eifelian stage differs only slightly from that of the original Eifel. Well known, stratigraphically significant, trilobite and brachiopod index fossils, and even dacryoconarids, support the boundary definition. Moreover, some bentonite horizons are close to the boundary, one yielding an age of 392.2 ± 1.5 Ma (Kaufmann et al., 2005) (cf. below).

The GSSP of the Givetian Stage is at Jebel Mech Irdane in the Tafilalt area of the Anti Atlas, southern Morocco (Walliser et al., 1995; Walliser, 2000). The entry of the index conodont *Polygnathus hemiansatus* is slightly later than that of the ammonoid *Maenioceras* and slightly earlier than of the spore *Geminospora lemurata* (Loboziak et al., 1990). The lithostratigraphic base of the Assisse de Givet, the name bearer, is vaguely defined and imprecisely dated in the Ardennes, Belgium. However, the historical base appears to be not much higher than the base of the global Givetian, just at the entry of the classical Givetian brachiopod *Stringocephalus*. According to Struve (e.g. 1982) and to correlations by Weddige (e.g. 1977), *Stringocephalus* sensu stricto first appears directly above a transgression event. In the Rhenish Mts. east of the Rhine, however, the base of the transgression event, here called “Odershäuser Transgression”, has traditionally been regarded as the base of an “Upper Middle Devonian”. In the pelagic facies of the Eastern Rhenish Mountains the transgression is marked by anoxic dark shales or limestones. The Devonian rocks near Prague show the same lithologies, but the anoxic horizons are here regarded as the so-called “Kačák Event”. Despite the different names, the same event has a distinct global distribution. In Morocco, the newly defined boundary falls in the upper part of the event interval. The Moroccan facies is purely pelagic, and rather condensed, and the time markers are within a few centimetres only (in sharp contrast to the Eifelian GSSP with high neritic sedimentation rates). Because of its distinct intercontinental distribution, the Kačák Event should be of particular interest for analyses of inorganic time markers.

The numerical age for the base of the Eifelian Stage (base Middle Devonian Epoch) may be in between 394 Ma and ca 390 Ma, whereas the age of 397.5 ± 2.7 Ma of House and Gradstein, 2004 (GTS 2004) seems too old (cf. 3.2.1). There are three U–Pb ID-TIMS ages, which are reasonably close with each other, around the base of the Eifelian Stage (cf. Weddige et al., 2005: Taf. IV): 390.0 ± 0.5 Ma (Roden et al., 1990), 391.4 ± 1.8 Ma (Tucker et al., 1998) and 392.2 ± 1.5 Ma (Kaufmann et al., 2005). Thus, the age of ca 392 Ma used in the DCP 2003 is a reasonable estimate.

The age of the lower boundary of the Givetian Stage varies between 378 Ma (Young and Laurie, 1996) and 391.8 ± 2.7 Ma (House and Gradstein, 2004; GTS 2004) which seems too old (cf. 3.2.1). In the DCP 2003 an age of 387 Ma is preferred, based on House’s (1995) analyses of successions of microrhythms that the Givetian Stage (DCP: 6 my) could be slightly longer than the Eifelian Stage (DCP: 5 my). There is no question, however, that the Middle Devonian Epoch is of shorter duration than either of the Early and Late Devonian epochs.

3.3.2. Germany (K. Weddige)

The base of the global Middle Devonian, and thus the global Eifelian Stage (GSSP), is identical to that of the Eifelian regional historical stage (Fig. 2: Heisdorf and Lauch Schichten). Also, the top of the new global Eifelian is very slightly below the historical top, which in the Eifelian region was drawn previously at the entry of *Stringocephalus* sensu stricto (Struve et al., 1997).

These correlations are based on overall improved biostratigraphic evidence, as the neritic Middle Devonian facies generally shifted to open marine connections with sufficient occurrences of extra-regionally distributed index fossils.

The increase of open-marine and outer-shelf connections may be due to a gradual decrease of the siliciclastic input from the Old Red Continent. In the Eifel area, at that time, limestones and marls already dominated over fine grained siliciclastics. In the Sauerland, however, which was situated closer to the Old Red Continent, siliciclastic sedimentation persisted sporadically until the latest Eifelian (Fig. 2: Mühlenberg and Brandenberg Schichten). This siliciclastic succession is abruptly overlain by widely transgressive deposits starting with anoxic shales, carbonates and pelagic fossils such as tentaculites and goniatites (Fig. 2: Odershausen and Wiedenest Schichten). The anoxic pulse is assigned to the global *otomari* Event or Kačák Event, which marks the latest Eifelian, immediately below the GSSP for the base of the Givetian (cf. 3.3.1). After this interval in the Sauerland, limestone deposition increased from interbeds in local reefs, to more widespread biostrome reefs (Fig. 2: Honsel Schichten, Schwelm Kalk). Finally, bioherms developed when crustal subsidence and reef growth were in a balance during the later Givetian and the early Late Devonian (Fig. 2: Massenkalk).

3.3.3. East Europe (A.S. Alekseev, B.I. Chuvashov)

Middle Devonian shallow-water sediments are widely distributed on the East-European Platform. The Eifelian age of the two lagoonal and evaporitic Dorogobuzhian and Klintsian regional substages is established by spore assemblages (Avkhimovich et al., 1993). The following marine interval of the Mosolovian and Chernoyarian regional substages contains shallow-water conodont assemblages which allow a weak correlation, but it is accepted that they correspond approximately with the *Tortodus kockelianus* and early *Polygnathus ensensis* zones of the GSS (Kononova and Kim, 2005).

In the Ural Mts., this succession is represented by shallow water limestones and dolomites (upper part of the Biysk Horizon), and bituminous marls and argillites with a relatively deep-water fauna (Afoninsk Formation). In most areas a gap occurs between the Eifelian and Givetian successions.

The Givetian is mainly terrestrial, and only in the southeastern part of the platform are thin limestone intercalations containing marine faunas. In previous Soviet stratigraphic schemes the uppermost Givetian Pashiyan Horizon is considered as lower Frasnian, but this was changed in 2002 according to the new GSSP definition of the significantly raised Givetian–Frasnian boundary.

In the Ural Mts., this succession is represented by clastics with small lenses of bauxite (palaeosols) at the base (Chussovsk Formation), overlain by shallow water limestones with a rich fauna of brachiopods and conodonts (Cheslavsk Formation). The upper part of the succession consists of the Pashiyan and Kynovian horizons, the latter containing a rich and diverse fauna; in the recent stratigraphic scale both are coeval to the *Ancyrodella binodosa* conodont Zone.

3.3.4. Central Asia (N.G. Izokh)

The lower boundary of the Eifelian is recorded in the late Kitabian by the FAD of *Eucostapolygnathus partitus*. The Rushsayian (Zeravshan Ridge, South Tien Shan) corresponds to the whole Middle Devonian (Fig. 2). It is represented by shallow-water bedded and massive limestones (bioherms) and siliceous-carbonate rocks, with abundant benthic and pelagic faunas (*Eucostapolygnathus costatus* to early *Mesotaxis falsiovalis* zones) (Kim et al., 1984, 1988; Yolkin et al., 2000). The stratigraphic equivalent of the Rushsayian, the Tchorgoian, is composed mainly of cherts which include limited conodont associations.

3.3.5. North America (K. Weddige)

The Onondaga Formation, which includes the Early–Middle Devonian boundary beds of New York State, consists dominantly of fine to coarse-grained, cherty to non-cherty limestone, and in its upper part, a prominent K-bentonite (Tioga Bentonite). An abrupt transgression with a widespread change from shallow to deeper shelf facies in the upper part of the Onondaga Formation marks the onset of Acadian Tectophase II sensu Ettensohn (1985). Because of the shift in facies, a highly diverse assemblage of brachiopods, corals, crinoids, gastropods, and other forms are present in the Onondaga Formation.

The overlying Hamilton Group is composed of four formations. All are dominantly siliciclastics, in contrast to the earlier carbonate dominance in the Onondaga Formation. This change in deposition was caused by the second pulse of the Acadian Orogeny, which created an extensive highland area east of New York State.

The Marcellus Formation is composed mainly of dark shales and siltstones which represent the initial influx produced by the Acadian Orogeny. Its upper part is marked by anoxic dark shales which indicate the Kačák Event (cf. 3.3.1). Above this level, in central New York State the Skaneateles Formation undergoes a transition from basinal shales to a sandy shallow shelf, succeeded by the Ludlowville and Moscow formations with their succession of fossiliferous mudstone, siltstone and sandstone. The Hamilton Group is succeeded by the Tully Limestone, the age of which has been controversial, but now known to be within the upper portion of the *Polygnathus varcus* Zone (cf. Rickard, 1975; Kirchgasser, 2000). It thus represents the Taghanic Event, which is regarded as the last distinct global event of the Middle Devonian (e.g. Aboussalam, 2003). The Tully Formation with its shallow marine carbonates was abruptly covered by thick black shales of the Genesee Formation. This major change of post-event facies is regarded as a widespread transgression, and thus marks the onset of the Acadian Tectophase III sensu Ettensohn. The overlying Genesee Group extends into the Late Devonian.

3.4. Frasnian and Famennian stages (Late Devonian Epoch)

3.4.1. GSSPs and numerical ages (K. Weddige, M. Menning)

The GSSP of the Late Devonian Epoch, and the coincident base of the Frasnian Stage, is at Puech de la Suque, Montagne Noire, France, within the lower Clapassous Formation (Klapper et al., 1987; House et al., 2000). The boundary is indexed by the conodont *Ancyrodella rotundiloba* early morphotype sensu Klapper. The FAD of this taxon, which is present in both pelagic and neritic facies, almost coincides with the base of the Frasnes Group in Frasnes, Belgium. The level of the boundary is substantially higher than that of previous opinions on the Late Devonian base, which approximated to the base of the Assise de Fromelennes Formation which underlies the Assise de Frasne in Belgium. There were, however, prolonged controversies on the taxonomy of the index conodont and therefore on its position within the conodont zonation (particularly by Ziegler and Sandberg, 1996). The latter authors finally placed the boundary within the Early *M. falsiovalis* Zone, i.e. at the top of the *Polygnathus norrisi* Zone (see Weddige, 1996). For accompanying fossil groups, the boundary is marked by a late stage of radiation of the pharciceratid ammonoids. A new goniatite record, *Neopharciceras*, together with conodont index fossils, permits world-wide correlation. In contrast, a short anoxic pulse, the so-called Frasnes Event, which lies slightly above the base, is of less global significance.

The GSSP for the base of the Famennian Stage was placed within the upper Coumiac Formation slightly above the Upper Coumiac Quarry, near Cessenon, Montagne Noire, France (Klapper et al., 1993; House et al., 2000). The stratigraphic level is very close to that of the base of the Famennian as formerly used in the Famenne area in Belgium (Bultynck and Martin, 1995). The boundary is indexed by the entry of *Palmatolepis triangularis*. The taxon is derived from the *Palmatolepis praetriangularis* stock demonstrating a proven phylogenetic lineage. Thus, the taxonomy of the boundary index is generally accepted, whereas the lithologic position of the GSSP is still under review. The GSSP, with respect to the FOD of *P. triangularis*, is immediately above a global mass extinction event, i.e. the last event of the Kellwasser Events, when, for example, the goniatites *Manticoceras* and *Beloceras* become extinct (e.g. Schindler, 1990; House et al., 2000). A haematite crust then, immediately above the GSSP spike, obviously represents a hardground with condensation, and it is suspected that the GSSP spike itself seems to mark nothing else than a gap in sedimentation, i.e. of documentation (e.g. Ziegler and Sandberg, 1996). Geochemical, magnetostratigraphic and other investigations are still in progress.

The numerical ages for the Late Devonian stages of the DCP 2003 were selected as young as possible using ID-TIMS ages of Tucker et al. (1998) and their error bars (cf. Weddige et al., 2005: Taf. IV). This is necessary to get a balance with the Early Carboniferous SHRIMP ages of Roberts et al. (1995) which are too young because of the heterogenity of the standard SL13 zircon used to determine them. For over 10 years the Famennian has been regarded as considerably longer than the Frasnian Stage, not least because of its division into a greater number of conodont and ammonoid zones. However, this time-relation has yet to be confirmed by a reliable isotopic age determination. Only in the time scale of Compston (2000), using SHRIMP and selected/reinterpreted MSID ages, is the Frasnian longer than the Famennian Stage (Fig. 1). Since the work of Tucker et al. (1998) the Famennian has a duration of about 15 my (Table 2) and, therefore is one of the longest Phanerozoic stages.

3.4.2. Germany (K. Weddige)

In the Rhenish Massif, the regional historical subdivision of the Late Devonian into Adorf-, Nehden-, Hemberg-, Dasberg- and Wocklum-Stufe, was mainly introduced by Wedekind (1913). This regional subdivision has been used extensively by survey geologists for intensive mapping for about one century.

Moreover, pioneering conodont studies started in the Rhenish Massif, where the Late Devonian pelagic limestone succession contain abundant conodont faunas, which have supported and promoted the usage of the regional stratigraphy. The discrepancy with the new global (GSSP) stages, however, is actually not serious, because the basal boundaries of the global Frasnian and the historical Adorf-Stufe are nearly coeval, but the latter is slightly younger (cf. Fig. 2). In contrast, the global Famennian corresponds to four regional historical stages which now could be taken as substages of a four-partite Famennian.

The Late Devonian of the Sauerland starts, locally, immediately above the "Roteisenstein-Horizont", which is derived from volcanic exhalative activities (Fig. 2). Then follow the reefal Adorf-Massenkalk formations and bioherms, which coincide with cephalopod limestones, or inter-reefal siliciclastic Adorf-Bänderschiefer (Fig. 2). These facies variants demonstrate a topographic wave pattern with small rises and troughs. Towards the end of the Frasnian, the reef growth generally slowed and finally stopped because of subsidence, too rapid for reef builders to match. All reefal development was terminated by two anoxic pulses, which are globally recognizable as the Early (Lower) and Late (Upper) Kellwasser events (Schindler, 1990). The latter is characterized by a mass-extinction (cf. Section 3.4.1). Later, slow pelagic sedimentation during the Famennian gradually levelled the Frasnian rise-trough relief. Red or yellow shales, ostracod shales or fine-grained silt- and sandstones or Plattenkalke filled up the troughs and thin cephalopod limestones, with enriched pelagic fossils, covered the rises (Fig. 2: Nehden-, Hemberg- and Dasberg-Schichten). In the Sauerland–as in the Harz Mountains and in Thuringia, too–this Famennian sedimentation pattern was interrupted twice by the semiglobal Annulata Event, Hemberg-Stufe and the global Hangenberg Event, late Wocklum-Stufe, anoxic pulses (Fig. 2; cf. Walliser, 1996).

3.4.3. East Europe (A.S. Alekseev, B.I. Chuvashov)

The marine Frasnian, dominantly carbonates, covers most of the East-European Platform. The lower boundary of the Frasnian, which is fixed by the FAD of the conodont *A. rotundiloba,* is difficult to trace because this level is in a clastic part of the succession. However, in the Timan area (north-eastern platform) the joint occurrence of *A. rotundiloba* and spore assemblage of the Timanian and Sargayan regional substages indicates that the base of the Frasnian lies within or at the base of the Timanian. Besides these two substages, the other 5 Frasnian substages contain abundant shallow-water conodont polygnathid assemblages which have been correlated to the standard zonation by Ziegler et al. (2000). In the Ural Mts. the FAD of the conodont *A. rotundiloba* in limestone of the Sargaevo Formation, marks the lowermost Frasnian Stage. The overlying formations are composed of shallow water limestones, occasional reefs, biostromes, and thin beds of chert, argillites, marls, and limestone. This succession contains the *Polygnathus timanicus–Ancyrognathus triangularis, Palmatolepis gigas, P. rhenana* and *P. linguiformis* zones.

It has long been known that a hiatus exists between the latest Frasnian (Livnian) and the earliest Famennian (Zadonskian). This gap spans mainly the basal Famennian *P. triangularis* Zone. But, rocks of this age (Volgogradian) occur in some restricted deep depressions located in the eastern part of the platform (Galushin and Kononova, 2004). The Famennian is represented mainly by carbonates, often with evaporitic intervals in its upper part. It includes 9 regional substages (Fig. 2). The Famennian conodont assemblages are dominated by shallow-water endemic taxa, but spore zonation gives some arguments for the correlation with the West European succession (Avkhimovich et al., 1993).

In the Ural Mts. the Famennian Stage is mostly represented by a shallow-water limestone succession defined by the *P. triangularis, P. crepida, P. rhomboidea, P. marginifera, P. trachytera, P. postera, P. expansa*, and *Siphonodella praesulcata* conodont zones. Foraminifers, brachiopods, and ammonoids are also used to divide and correlate the Famennian. The short stratigraphic break between the Famennian and Tournaisian stages lies at an undetermined position within the succession.

3.4.4. Central Asia (N.G. Izokh)

Late Devonian sections in the Zeravshan Ridge are represented by bedded and massive shallow-water limestones. They are characterized by diverse conodont associations, with almost all zonal index-species (Kim et al., 1984, 1988). The uppermost Givetian and Frasnian are united in a single unit, the Kindyktepian. It is determined by the presence of *Polygnathus dengleri*, *Mesotaxis asymmetricus*, *Skeletognathus norrisi*, *A. binodosa*, *A. alata*, and *Palmatolepis transitans* (Kim et al., 1984, 1988; Yolkin et al., 2000). The Duobian include Famennian conodont associations (Kim et al., 1984; Yolkin et al., 2000). Late Devonian sections from South Fergana are characterized by diverse Frasnian and Famennian conodont taxa (Kim et al., 1988).

3.4.5. North America (K. Weddige)

Above the Tully Limestone, open-marine black shales transgress from the interior of North America eastward across New York State during four major phases, which indicate sea level highstands: the Geneseo, Middlesex, Rhinestreet and Dunkirk. These define the initiation of the Genesee, Sonyea, West Falls, and Canadaway–Conewango groups (Ettensohn, 1987). Each group actually has to be regarded as a cycle. After the initial highstand each cycle is upward-shallowing and upward-coarsening as a result of basinward, i.e. westward, prograding wedges of the Catskill delta siliciclastics (Brett and Ver Straeten, 1997).

In terms of the Late Devonian stages, the upper part of the Genesee, the Sonyea and West Falls groups are Frasnian, and the Canadaway to Conowango groups are Famennian. The Frasnian–Famennian boundary is distinctly placed for the upper Hanover Shale of the West Falls Group, where the black shales demonstrate an abrupt faunal extinction. This is the well-known Upper Kellwasser Extinction Event, which permits stratigraphically worldwide correlation of the Frasnian–Famennian boundary (cf. Section 3.4.1).

4. Carboniferous GSSPs and Regional Stratigraphic Scales

4.1. General (P.J. Jones, A.S. Alekseev, P.H. Heckel, M. Menning, D. Weyer)

A bipartite division of the Carboniferous has long been used in Central/West Europe and North America as supraregional units, named respectively, the Dinantian and Silesian series or subsystems, and the Mississippian and Pennsylvanian systems. These different supraregional units cover significantly different time spans (Fig. 3). In contrast, the Carboniferous strata of Russia are traditionally subdivided into three series: Lower (Tournaisian, Visean, Serpukhovian stages), Middle (Bashkirian, Moscovian stages) and Upper (Kasimovian, Gzhelian stages). A tripartite division of the Carboniferous also has been used for many decades in China, where a succession previously referred to the Lower Permian, is now placed in the Carboniferous as a result of the introduction of the GSSP concept. In addition, the regional Middle and Upper Carboniferous of the Russian Supraregional Stratigraphic Scale (RSS) were different from those of the traditional upper units of the bipartite subdivisions to which they were considered equivalent (Figs. 3 and 4).

In order to establish a scale for global correlation the SCCS approved, and the International Commission on Stratigraphy decided, to use a bipartite, rather than a tripartite, division of the Carboniferous, with each of the two divisions recognized as a subsystem (Engel, 1989). The SCCS later approved the selection of the lowest occurence of the conodont *Declinognathodus noduliferus* s.l. to define the global Mid-Carboniferous Boundary (MCB), and ratified the GSSP for this boundary at the base of the Pennsylvanian at Arrow Canyon, Nevada (see Lane et al., 1999; Richards et al., 2002). Based on this new definition, Russian stratigraphers lowered the boundary between the Serpukhovian and Bashkirian stages to the new global MCB. The choice of the new global stratotype for the MCB has tended to depreciate the use of the RSS of Central/West Europe as a Global Stratigraphic Reference Scale (GSS) because the global MCB is within the middle of the Namurian (Fig. 3).

The Mississippian (Early Carboniferous) and the Pennsylvanian (Late Carboniferous) are recently subdivided into three series each named according to their position, and seven stages with the names used in Belgium and Russia (Heckel, 2004; Work, 2004; www.stratigraphy.org/gssp.htm). Menning et al. (2001a) preferred to classify the Mississippian and Pennsylvanian as series/epochs (as in STD 2002) in order to avoid the anomalous situation of the Carboniferous being the only system in the GSS having major subdivisions at the subsystem level (Roberts, 2000).

4.2. Tournaisian, Visean, and Serpukhovian stages (Mississippian, Early Carboniferous)

4.2.1. GSSPs and numerical ages (D. Weyer, A.S. Alekseev, F.-X. Devuyst, L. Hance, T.I. Nemyrovska, M. Menning)

The Devonian–Carboniferous (Famennian–Tournaisian) boundary is now defined by the first appearance of the index conodont *S. sulcata* within the evolutionary lineage from *S. praesulcata* to *S. sulcata* in the stratotype section at La Serre, Montagne Noire, southern France (Paproth et al., 1991). This time level is very near to, but slightly older than, the traditional ammonoid boundary–entrance of *Gattendorfia subinvoluta*–accepted by the 2nd International Carboniferous Congress in 1935 (Paeckelmann and Schindewolf, 1937). The golden spike at La Serre is situated within an allochthonous turbiditic succession, and because many of its critical conodont data are based on reworked faunas, it has been the subject of much criticism (e.g. Ziegler et al., 1988). In retrospect, the ratified GSSP appears to be a comparatively bad choice. On the other hand, there has been no proposal since 1991 of an ideal,

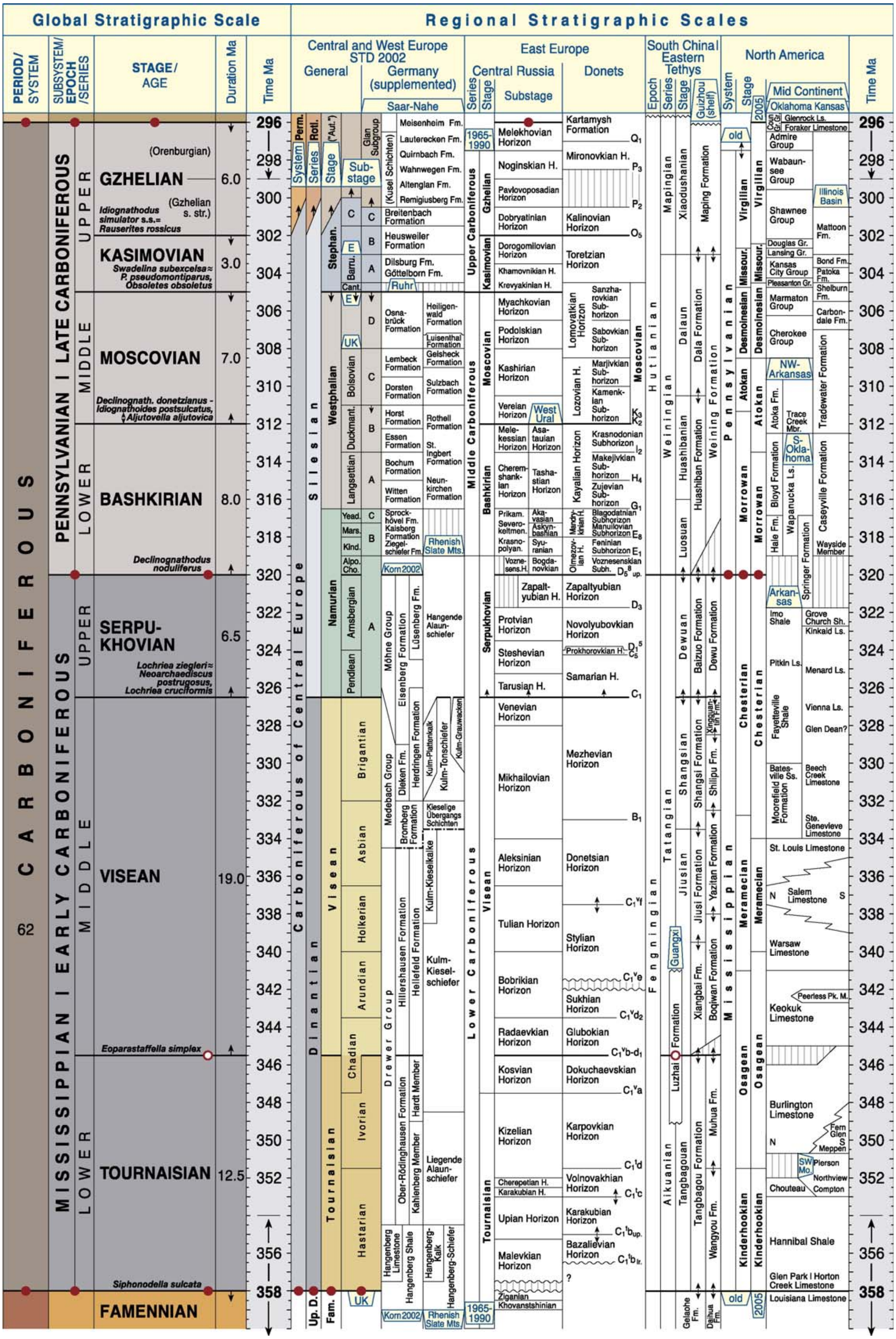

Fig. 3. Carboniferous Global Stratigraphic Scale as used in the DCP 2003 and Regional Stratigraphic Scales.

or more appropriate, section for a possible change in the future.

The base of the Visean was first formally defined in the Dinant Basin (SCCS, 1969). However, subsequent work revealed that the section chosen (Bastion section, near Dinant) was unsatisfactory, not only for long distance correlation, but also at the local scale. Entries of stratigraphically significant foraminifer taxa are facies

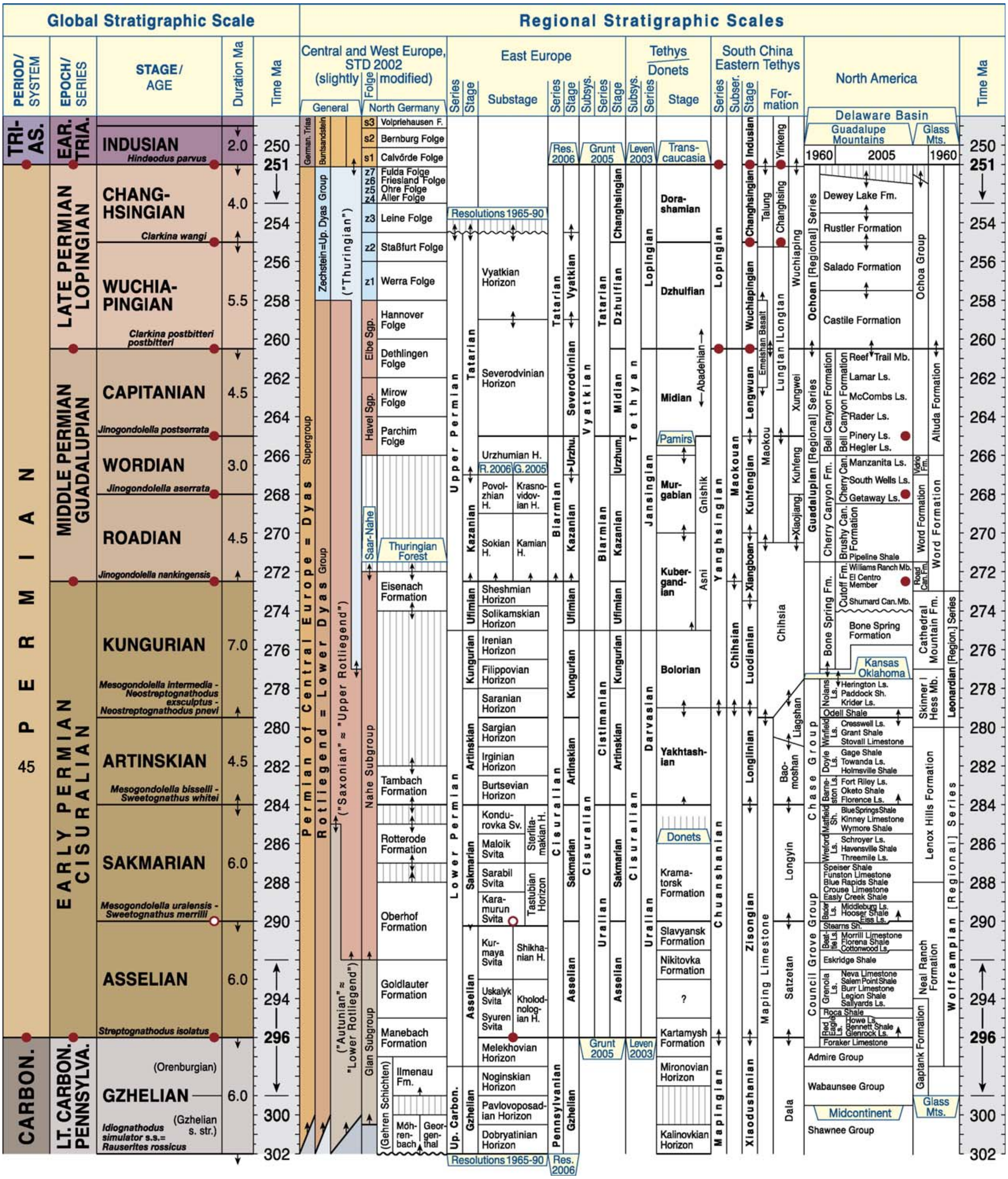

Fig. 4. Permian Global Stratigraphic Scale as used in the DCP 2003 and Regional Stratigraphic Scales.

controlled. In addition, the use of unspecified *Eoparastaffella* (Conil et al., 1969) for defining the base of the Visean was inappropriate. The SCCS Working Group on the Tournaisian/Visean boundary followed the proposal of Hance and Muchez (1995) to search for a criterion within the evolutionary lineage of the genus *Eoparastaffella*. Work by its members, summarized in Devuyst et al. (2003) led to the proposal that the first appearance of *Eoparastaffella simplex* in the lineage *E. ovalis* → *E. simplex* should be used in the choice of the base of the Visean. This proposal was formally approved by the Voting Members of the SCCS in 2002 (Work, 2002) (cf. Fig. 3).

The proposed new GSSP for the base of the Visean at Pengchong (Guangxi) southern China (Devuyst et al., 2003) is in the process of ratification. The section exposes a unit of allochthonous mass-flow deposits (Pengchong Member, Luzhai Formation) which accumulated in a starved, intra-platform basin. The evolution of *Eoparastaffella* is exceptionally well documented. The base of the Visean is defined at the base of bed 85, in which *E. simplex* first appears. The conodont *Gnathodus homopunctatus*, a useful, but cryptogenic, guide for the base of the Visean, enters in bed 86, about 1 m above the boundary (see Section 4.2.5).

This new Tournaisian–Visean boundary is significantly younger than some historical (pre-1969) interpretations. Certain commonly used index macrofossils, formerly regarded as earliest Visean, on microfossil evidence, now indicate latest Tournaisian, e.g. the productid brachiopod *Levitusia humerosa* in the shallow water facies of Western Europe was thought to indicate earliest Visean, in contrast to the long controversial, but now accepted, Russian dating of this biozone as late Late Tournaisian in the Kosvian Horizon of the Ural Mts. Similarly, microfossil evidence indicates the *Ammonellipsites*/*Merocanites* ammonoid community in Europe, North Africa, Middle Asia, and North America should be removed from the basal Visean and restricted to an expanded Late Ivorian.

The Serpukhovian was proposed in 1890 and officially included into the stratigraphic scale of the USSR in 1971 as an equivalent of the Early Namurian. Its type area is at the south limb of the Moscow Basin about 100 km south of Moscow, in the vicinity of Serpukhov. Originally it was defined on the basis of brachiopod assemblages and later subdivided into several foraminiferal zones. The conodont and foraminiferal assemblages of the early Serpukhovian are very similar to those of the late Visean (Gibshman, 2001; Nikolaeva et al., 2002). The late Serpukhovian was redefined mainly using the more complete succession of the Donets Basin (late Arnsbergian Zapaltyubian Horizon).

The boundary between the Visean and Serpukhovian stages has been traditionally recognized on the basis of goniatites, and placed at the base of the *Cravenoceras leion* Zone. Korn (1996) proposed *Edmooroceras pseudocoronula* as a better alternative, because it is geographically less restricted. However, although Mississippian goniatites provide high-resolution regional subdivisions, some endemic taxa are of limited use for effective interregional and intercontinental correlation.

Conodonts have also been proposed for the identification of the Visean–Serpukhovian boundary in Europe (Skompski et al., 1995). A SCCS Task Group led by B. Richards is currently working on finding a suitable criterion and GSSP for the base of the Serpukhovian. The lineage *Lochriea nodosa*–*Lochriea ziegleri*, which is recorded and controlled by ammonoids in the Rhenish Slate Mts., Germany, has been selected for this purpose. The FAD of *L. ziegleri* (Nemirovskaya et al., 1994) has been proposed to define the Visean–Serpukhovian boundary, because of its wide geographic distribution (Nemyrovska, 2005). The species is used to recognize the Visean upper boundary in China (Wang and Qi, 2003) and the Ural Mts. (Nikolaeva et al., 2002). On the other hand Gibshman and Baranova (in press) propose the use of the foraminifer "*Millerella*" *tortula* in the lineage "*Endostaffella" tortula* (part)–"*M." tortula* (part), and of *Janischewskina delicata* in the lineage *J. typica*–*J. delicata*. According to the authors, the recognition of "*M." tortula* at the base of the Serpukhovian in the Moscow Region and the Pericaspian Basin offers the possibility of a correlation with the Middle Chesterian of North America.

The numerical age for the base of the Tournaisian Stage (base of the Carboniferous Period) is between 354 Ma (Young and Laurie, 1996) and 362 Ma (Tucker et al., 1998) (Fig. 1). The 354 Ma age was derived from an Early Carboniferous SHRIMP age (Claoué-Long et al., 1993), which is now known to be too young because of the inherent problem in the SL13 zircon standard used for the determination. Nevertheless, the 354 Ma age was applied as a tie point in the Phanerozoic time scales of Young and Laurie (1996) and Gradstein and Ogg (1996), and in the Carboniferous time scales of Jones (1995) and Menning et al. (2000) (Fig. 1), not least, because the dated sample was collected from the *S. sulcata* conodont Zone in the well known Hasselbach section in the Rhenish Slate Mts.

The older age of 362 Ma was derived mainly from the average U–Pb ID-TIMS age of 363.6 ± 1.6 Ma from the latest Famennian Piskahegan Group of New Brunswick, Canada (Tucker et al., 1998) and, less significantly, from a Rb–Sr isochron age of 361.0±4.1 Ma for the *S. sulcata* Zone in Nanbiancun, South China (Yang et al., 1988). Because of the 8 my discrepancy between the 354 Ma and 362 Ma ages, a compromise age of 358 Ma is used for the STD 2002 and the DCP 2003. Recently, Trapp et al. (2004) derived an age of 360.7 Ma for the DCB from two tuffs from the Hasselbach section, Rhenish Slate Mts. Their age of 360.5±0.8 Ma and the age of 353.7 ±4.2 Ma (Claoué-Long et al., 1993; renormalized by Claoué-Long et al., 1995) are both from the same tuff in bed 79. Using the ages of 360.5±0.8 Ma and 360.2 ±0.7 Ma of Trapp et al. (2004) and the thicknesses of the Hasselbach section instead of the thicknesses of the Lali section (South China, Trapp et al., 2004), the age of the DCB is estimated at 361.4 Ma (Menning, this work, Fig. 1). A Re–Os age of 361.3±2.4 Ma of a black shale from the Exshaw Formation, Canada (Selby and Creaser, 2005) is very close to it.

The duration of the Tournaisian Stage in many charts is shown to vary between 10 my and 14 my (Table 2). However, U–Pb ID-TIMS ages from the Rhenish Slate Mts., which yielded an age of ca. 361 Ma for its base (Trapp et al., 2004) and ca. 342 Ma for its top (Trapp, pers. com.), suggest a longer duration of slightly less than 20 my. Thus, these ages suggest the Tournaisian is approximately as long as the Visean Stage, which to date, is, together with the Norian Stage, the longest of the Phanerozoic Eon (STD 2002; Menning et al., 2005).

The age for the base of the Serpukhovian Stage has been changed from 333 Ma (Harland et al., 1990) to 327–325 Ma, but there are no reliable isotopic ages about this boundary. The DCP age of 326.5 Ma is based on the Pendleian/Arnsbergian coal-tonstein $^{40}Ar/^{39}Ar$ age of 324.6 Ma (Lippolt et al., 1984). If preference is given to the middle Arnsbergian age of 319.5 Ma (Lippolt et al., 1984) in lieu of the Pendleian/Arnsbergian age of 324.6 Ma, an age of 325 Ma or even less, is more probable for the base of the Namurian/Serpukhovian (cf. Menning et al., 2000: Fig. 7, Scale B).

4.2.2. Central and West Europe (D. Weyer, F.-X. Devuyst, L. Hance)

The DCB *S. sulcata* criterion is not applicable in the shallow carbonate platform facies of the Franco-Belgian Basin. Here, the DCB is traditionally placed at the top of the basal (1 m thick) bed (“Tn 1bα”) of the Hastière Formation, a coarse grained bioclastic grainstone with lithoclasts and ooids and a reworked Devonian fauna. This bed is included between “known Devonian” and “known Carboniferous” (Van Steenwinkel, 1992).

The Early Tournaisian (mid-Hastarian) of Western Europe includes a remarkable anoxic black shale, just above the basal *Siphonodella crenulata* Zone. It is seen both in basinal facies (Lower Alum Shale and Rußschiefer of Germany) and in shallow water facies (Schistes du Pont d’Arcole of Belgium, Malevkian on the East European Platform). Cephalopod limestones (*Gattendorfia* Genozone) and ostracod shales (with “finger print” entomozoids) below this semiglobal event are reminiscent of the Frasnian/Famennian dysphotic/aphotic facies development.

The historical stratotype for the base of the Visean is located in Belgium in the Bastion section. The boundary lies within the uppermost part of the Leffe Formation which contains thin grainstone layers derived from shallower areas (Lees, 1997). The first *Eoparastaffella* sp. (a broken specimen unidentifiable at the level of species) is found less than 1 m below the entry of the conodont *G. homopunctatus* (Conil et al., 1989).

The utilization of the original British stages of the Visean (George et al., 1976) has been greatly enhanced by the critical and detailed review of Riley (1993). In the British Chadian stratotype section (Chatburn, Craven Basin, NW England), Riley (1990, 1995) showed that the genus *Eoparastaffella* enters about 300 m higher than originally reported (George et al., 1976) and precedes the incoming of *G. homopunctatus*. Riley (1990) therefore introduced the term ‘late Chadian’ for that part of the Chadian recognized by *Eoparastaffella* and accessory taxa, such as *G. homopunctatus*. Revisions of the widely used Brigantian, Pendleian and Arnsbergian ammonoid zonations of western Europe (Korn, 1996; Korn and Horn, 1997) demonstrate a greater degree of provinciality than previously estimated, when comparisons are made with Mediterranean and Russian ammonoid faunas. The Tournaisian and Visean and the Belgian substages, Hastarian to Warnantian, are revised by Dejonghe (2006).

4.2.3. East Europe (A.S. Alekseev, B.I. Chuvashov, F.-X. Devuyst, L. Hance, D. Weyer)

On the East European Platform the Early Carboniferous deposits are composed of mainly marine carbonates with minor terrestrial intervals. Rocks of the basal Tournaisian Stage are absent in the central part of the platform, and the marker conodont species *S. sulcata* occurs only in sections of the western Ural Mts. Before 1986 (approved officially in Kagarmanov, 1998) the uppermost Famennian was considered as earliest Tournaisian, comparable with the Etroeungtian (“Tournai 1a”) in Central/West Europe, which now also belongs to the Devonian. The Tournaisian includes 6 regional substages — Malevkian (*Siphonodella duplicata*), Upian (*S. kononovae*), Karakubian, Cherepetian (*S. quadruplicata*), Kizelian (*Siphonodella isosticha*, *Dollymae hassi*), and Kosvian. However, before 2002 the Kosvian substage (late *Gnathodus typicus* and *Scaliognathus anchoralis* zones) was regarded as the lowest unit of the Visean (Fig. 3). Whereas the coeval foraminiferal zones correspond approximately to each conodont zone, the brachiopod zones have a coarser stratigraphic resolution. The foraminiferal fauna from the Kosvian correlates well with the latest Tournaisian of Belgium (Brenckle, 1997). The late Tournaisian sediments of the Kizelian and Kosvian regional substages are absent in the central part of the platform, and are developed only in deep troughs in the eastern part, and at the western slope of the Ural Mts. Such gaps and unsuitable facies hinder the recognition of the Tournaisian–Visean boundary in the Moscow Syneclise (Hecker, 2002).

The Visean Stage can be subdivided into two parts. In most places the early part (Radaevkian and Bobrikian regional substages) is a terrestrial coal-bearing succession, with a coeval succession of shallow water limestones, marls, and argillites. The later shallow-water carbonate succession is composed of 4 regional substages — Tulian (*Endothyranopsis compressa–Archaediscus krestovnikovi* foraminiferal Zone), Aleksinian (*Endothyranopsis crassa–Parastaffella luminosa*), Mikhailovian (*Eostaffella ikensis*), and Venevian (*Endothyranopsis sphaerica–Eostaffella ikensis tenebrosa*). The late Tulian, Aleksinian, and Mikhailovian substages are coeval with the conodont Zone *Gnathodus bilineatus bilineatus* (Makhlina et al., 1993).

Marine limestones and dolomites of the Serpukhovian Stage are widely distributed. This stage is subdivided into 5 regional substages: Tarusian, Steshevian, Protvian (*Eostaffella protvae–Eostaffella minifica*), and Zapaltyubian and Voznesenskian (Fig. 3), but rocks of the later substages are often absent due to the Mid-Carboniferous disconformity.

4.2.4. Donets (T.I. Nemyrovska)

The Tournaisian Stage in the Donets Basin is composed of marine carbonates, 90–200 m thick. Its basal parts are absent, and its lower boundary lies within a gap (paraconformity) between the uppermost Devonian and the Bazalievian Horizon. The Tournaisian comprises five horizons: the Bazalievian Horizon (lower C_1t b Zone) composed of limestones and dolomites, 50–60 m thick, with *Bisphaera* and *Siphonodella* aff. *sulcata* in the lowermost beds and primitive Tournaellidae in the upper part, with the conodonts *Patrognathus andersoni* and *Siphonodella* spp.; the Karakubian Horizon (upper C_1^t b Zone), 20–60 m thick, with abundant chernyshinellids and the brachiopod *Eudoxinia media*; the Volnovakhian Horizon (C_1^tc Zone), 25–40 m thick, with siliceous limestones in the middle part, and conodonts of the *Neopolygnathus communis* group; the Karpovkian Horizon, 22–30 m thick (C_1^td Zone) with the conodonts *Cavusgnathus regularis*, *C. unicornis*, and *Hindeodus scitulus* and abundant foraminifers *Carbonella* and *Spinoendothyra* and the first *Dainella*; and the Dokuchaevskian Horizon (C_1^va Zone), 5–6 m thick, consisting of interbedded mudstones, claystones and thin micro-grained limestones with abundant brachiopods and corals, conodonts were not been recorded.

The Visean Stage in the Donets Basin, originally identified by brachiopods, has its lower boundary defined by foraminifers and corals. The evolutionary lineage of the foraminiferid genus *Eoparastaffella,* first established by Vdovenko (1964), has been used to identify its base. The Dokuchaevskian Horizon, previously included in the early Visean, is now placed in the late Tournaisian. The present base of the Visean coincides with the boundary between the local biozones C_1^va and C_1^vb (Vdovenko, 1973; Poletaev et al., 1990; Vdovenko et al., 1990) or between the Dokuchaevskian and Glubokian horizons. This boundary correlates with the base of the Radaevkian of the Moscow Syneclise (Poletaev et al., 1990; Vdovenko et al., 1990). Most of the Visean (200–800 m thick) is represented by limestones (C_1^vb–C_1^vf zones), with a member of mudstones and siliceous rocks intercalated with black radioactive shales in the lower part of the Stylian Horizon (Fig. 3, middle Visean). Mudstones interbedded with cherty and crinoidal limestones of the upper part of the Stylian contain the brachiopod *Gigantoproductus* and the conodonts *Gnathodus girtyi* and *Lochriea commutata*. The late part of the Visean, i.e. Mezhevian Horizon ($C_1^v g_1$) consists of up to 500 m of terrigenous rocks with limestone intercalations.

The beginning of the latest Visean regressive cycle in the Donbas is expressed by a shallowing of the basin, and heralded the advent of the Serpukhovian Stage. Marine carbonates changed to thick, mostly terrigenous claystones, siltstones and sandstones with limestone intercalations and coal seams during the time of the major Mississippian coal formation. Because of unfavourable facies in the Donbas, *L. ziegleri* was not found at the base of the Serpukhovian Stage, but higher, at the level of the Novolyubovkian Horizon. In the Dn–Donets Depression *L. ziegleri* enters at the base of Horizon IX (Skompski et al., 1995), and coincides with the base of the Serpukhovian, as defined by foraminifers (Brazhnikova et al., 1967). The basal part of the Serpukhovian contains land plants, which allow a correlation with the lowermost Namurian. As the Serpukhovian is incomplete in its type area (Moscow Basin), the Zapaltyubinian Horizon from the Donets Basin was added to the Serpukhovian Stage, and therefore, the Donets Basin is regarded as a parastratotype of the Serpukhovian. The end of the Zapaltyubinian is characterized by the mass extinction of the majority of early Carboniferous faunas, among them the conodont genera: *Gnathodus*, *Lochriea* and *Cavusgnathus* (Nemyrovska, 1999).

4.2.5. South China (X.-D. Wang)

The Fengningian Epoch, which was originally defined as System (Ting, 1931), is equivalent to the global Mississippian Subsystem/Epoch. It comprises the Aikuanian and Tatangian series, both of which used to be referred to as stages (Yang et al., 1979). The Aikuanian Series consists solely of the Tangbagouan Stage, whereas the Tatangian Series encompasses the Jiusian, Shangsian, and Dewuan stages (Fig. 3).

The Tangbagouan Stage, first proposed by Jin et al. (2000) as the time equivalent of the Tangbagou Formation (Ting, 1931), comprises 300 m of sandstone, mudstone, and limestone of inner shelf facies. It is widely recognizable by the presence of the rugose coral *Pseudouralinia* in the middle and upper part, coinciding with two ascending brachiopod zones of *Martiniella–Eochoristites* and *Finospirifer shaoyangensis* (Zhang, 1987). The standard conodont zonation (*S. sulcata*, Lower *S. duplicata*, Upper *S. duplicata*, *S. sandbergi*, Lower *S. crenulata*, Upper *S. crenulata–S. isosticha*, *Gnathodus semiglaber–G. typicus*, and *Scaliognathus anchoralis*), is recognized throughout the pelagic Tangbagouan Stage (Wang, 1990a). Therefore, the stage can be correlated in detail with the Tournaisian Stage of West Europe.

The Jiusian Stage was named by Jin et al. (2000). In southern Guizhou it consists of limestones with intercalated sandstones in the late part. It is subdivided into three brachiopod zones: *Schuchertella magna*, *Delepinea subcarinata–Megachonetes zimmermanni*, and *Vitiliproductus groeberi*, but only contains the coral Zone *Thysanophylloides* (= *Dorlodotia*). Jiusian conodonts are grouped into the *Gnathodus texanus–G. homopunctatus* Zone and the *Paragnathodus commutatus* Zone in slope facies, southern Guizhou (Xiong and Zhai, 1985; Wang, 1990a). This stage can be correlated with the western European late Chadian to the early Holkerian substages.

The Shangsian Stage was suggested by Zhang (1988) as a chronostratigraphic equivalent of the Shangsi Limestone from southern Guizhou, where corals are used to recognize the *Yuanophyllum* Range Zone. The early part of this zone is characterized by *Palaeosmilia* and *Kueichouphyllum heishikuanense*, the middle part by *Aulina carinata*, and the late part by *Palastraea* (Wang, 1990b). The brachiopods are grouped into the *Balakhonia yunnanensis* Zone and the *Delepinea comoides–Datangia weiningensis* Assemblage. The Shangsian conodont succession consists of the lower *G. bilineatus bilineatus* Zone and the upper *Lochriea* (synonymous with *Paragnathodus*) *nodosa* Zone (Wang, 1990a). The stage may correspond to the late Holkerian to Brigantian substages.

The Dewuan Stage, named from western Guizhou, comprises limestones and dolomites intercalated with marls, shales and chert bands (Hou et al., 1982). It contains the brachiopods *Gondolina weiningensis*, *Latiproductus latissimus*, *Gigantoproductus edelburgensis*, and *Striatifera angulata*, and the corals *Palastraeaeosmilia regia* and *Aulina rotiformis*. It coincides with the ammonoid *Dombarites–Eumorphoceras* Zone in pelagic facies in northwestern Guangxi (Ruan, 1989) and the conodont *Gnathodus bilineatus bollandensis–Adetognathus unicornis* Zone in slope facies in southern Guizhou (Wang et al., 1987). The Dewuan Stage is approximately correlated to the Serpukhovian Stage.

4.2.6. North America (P.H. Heckel, D.M. Work)

The regional Kinderhookian Stage at the base of the Mississippian Subsystem was named from the Mississippi River valley in western Illinois, where it comprises 30–50 m of marine shale and siltstone with limestone at the base and top. The conodont *S. sulcata*, which identifies the Devonian–Carboniferous boundary, occurs in the base, and the stage spans seven ascending conodont zones based on *Siphonodella* species: *sulcata*, Lower *duplicata*, Upper *duplicata, sandbergi*, Lower *crenulata*, Upper *crenulata*, and *isosticha* (Lane and Brenckle, 2001). The deeper-shelf, type Kinderhookian contains only a few earlandiid, tournayellid, and endothyrid foraminifers, but the latter two groups become more abundant and diverse higher on the shelf in north-central Iowa. The lower part of the stage belongs to the *Gattendorfia–Eocanites* ammonoid Genozone, and the upper part belongs to the *Goniocyclus–Protocanites* Genozone. The Kinderhookian is essentially equivalent to the western European Hastarian Substage.

The overlying regional Osagean Stage was named from southwestern Missouri, where it encompasses ~80 m of limestone. It comprises the Burlington and Keokuk limestones, and the lower Warsaw Formation northeastward in the Mississippi Valley. It spans the following conodont zones (in ascending order): *Gnathodus punctatus*, *Neopolygnathus communis carinus, Pseudopolygnathus multistriatus, Doliognathus latus, Bactrognathus lanei, Eotaphrus burlingtonensis, Gnathodus bulbosus*, and *G. texanus* (Lane and Brenckle, 2001), of which the latter extends above the top of the stage. Kammer et al. (1990) raised the top of the Osagean from the top of the Keokuk Limestone to a faunal change among foraminifers and macrofossil groups in the middle of the Warsaw Formation. Foraminifers are sparse throughout most of the Osagean in this region, but a diverse fauna including *Eoendothyranopsis* and *Viseidiscus* in the Peerless Park Member in the upper middle Keokuk allows correlation with the British early Arundian and Belgian mid-Moliniacian substages in the global early Visean Stage (Brenckle et al., 1982). This places the Burlington Limestone as essentially equivalent to the Belgian Ivorian Substage at the top of the Tournaisian Stage. Ascending ammonoid genozones in the Osagean are *Pericyclus*, encompassing most of the Burlington, *Fascipericyclus–Ammonellipsites*, spanning the Burlington–Keokuk boundary, and *Bollandites–Bollandoceras*, which includes the Peerless Park Member and extends into the overlying stage.

The overlying regional Meramecian Stage was named from the St. Louis area, where it consists of ~120 m of limestone comprising the upper Warsaw Formation, Salem Limestone, and St. Louis Limestone. Its base was redefined at the first appearance of the foraminifer *Globoendothyra baileyi* in the middle Warsaw by Kammer et al. (1990). Calcareous foraminifers (including *Eoendothyranopsis* and *Paraarchaediscus*) flourished and diversified during the Meramecian, but have not yet provided formal regional zonation. No change in conodont or ammonoid faunas is yet known at the basal Meramecian boundary. The top of the Meramecian once included the Ste. Genevieve Limestone above the St. Louis, but Maples and Waters (1987) lowered the top to the St. Louis–Ste. Genevieve contact. The upper Meramecian contains the *H. scitulus–Apatognathus scalenus* conodont Zone (Lane and Brenckle, 2001) and the lower part of the *Beyrichoceras* ammonoid Genozone, of which the latter also includes the Ste. Genevieve Limestone. These considerations allow correlation of the Meramecian with the British Holkerian and early Asbian substages and the Belgian Livian and early Warnantian substages in the middle of the global Visean Stage.

The regional Chesterian Stage at the top of the Mississippian was named from southwestern Illinois, where it comprises a succession of alternating limestone, shale, and sandstone up to 400 m thick. Its base is the base of the Lower *G. bilineatus* conodont Zone in the lower Ste. Genevieve Limestone. This zone is followed upward by the Upper *G. bilineatus* Zone, *Cavusgnathus naviculus* Zone, and *A. unicornis* Zone, which is at the top of the Chesterian Stage in its type region (Lane and Brenckle, 2001). A disconformity spanning the Lower and Upper *Rhachistognathus muricatus* zones, which are known elsewhere, separates the type Chesterian from the Early Pennsylvanian in this area. The base of the Chesterian is also characterized by the first appearance of the foraminifers *Asteroarchaediscus* and *Neoarchaediscus* in the lower Ste. Genevieve, but diversity decreased throughout the rest of the early Chesterian, and none of the larger forms characteristic of the late Visean is known here. Foraminifers again become diverse with the appearance of *Eosigmoilina* and *Brenckleina* at the base of the *C. naviculus* Zone in the Menard Limestone, which confirms correlation of the upper Chesterian with the global Serpukhovian Stage (Lane and Brenckle, 2001). Chesterian ammonoid zones are better known in northwestern Arkansas and the Appalachian region, but recognition of the *Lusitanoceras–Lyrogoniatites* Genozone in the mid-Chesterian Beech Creek Limestone allows correlation of the middle Chesterian with the middle of the British Brigantian Substage of the late Visean in western Europe.

4.3. Bashkirian, Moscovian, Kasimovian, and Gzhelian stages (Pennsylvanian, Late Carboniferous)

4.3.1. GSSPs and numerical ages (V.I. Davydov, A.S. Alekseev, T.I. Nemyrovska, M. Menning)

The GSSP for the base of the Pennsylvanian Epoch, and the co-incident base of the Bashkirian Stage, is established in Nevada, USA. The global Bashkirian starts with the FOD of the conodont *D. noduliferus* in the lower Bird Spring Formation at Arrow Canyon in the Great Basin, Nevada, ca. 75 km NE of Las Vegas (Lane et al., 1999; Brenckle et al., 1997). Here, transgressive bioclastic and mixed bioclastic–siliciclastic limestones and fine sandstone successions are separated by regressive mudstones as well as oolitic and brecciated limestone intervals with palaeosols. For more details on the original Bashkirian Stage in Russia see Section 4.3.3.

The index conodont *D. noduliferus* s.l. first appears within the chronocline *Gnathodus girtyi simplex–D. noduliferus* s.l. at 82.9 m above the base of the Arrow canyon section. An alternative evolutionary chronocline for *D. noduliferus* also has been proposed (Nigmadganov and Nemirovskaya, 1992; Nemirovskaya and Nigmadganov, 1994; Nemyrovska, 1999) based on studies in the Donets Basin and Central Asia.

In the ammonoid zonation, the base of the Bashkirian is close to the base of the *Homoceras* Genozone or *Isohomoceras subglobosum* Zone of Great Britain, Nevada, and Central Asia (Ramsbottom, 1977; Ruzhentsev and Bogoslovskaya, 1978; Nikolaeva, 1995). However, it is likely that the first *I. subglobosum* in Nevada and Central Asia occurs slightly earlier in the latest Serpukhovian (Nemirovskaya and Nigmadganov, 1994; Titus et al., 1997). The MCB is well documented and controlled by ammonoids in Stonehead Beck, England, in the early part of the Chokierian (Riley et al., 1987; Varker et al., 1990). Also it is identified in the Donets Basin at the base of the Voznesenskian Horizon, in the Ural Mts. at the base of the Bogdanovkian Horizon, and in the South Tienshan at the base of the Suffinsk Formation. It is also distinguished in the Rhenish Slate Mts., the Pyrenees, and the Cantabrian Mts. The foraminiferid *Globivalvulina bulloides* is a useful marker to approximate the boundary; although in Japan, Arctic Alaska, the Pyrenees, and Donets Basin it appears slightly below the FAD of *D. noduliferus*. Other foraminiferids, *Millerella pressa* and *M. marblensis*, are informal markers for the boundary (Rauzer-Chernousova et al., 1996; Brenckle et al., 1997). In the Arctic, Pyrenees, East European Platform, Ural Mts., Donets Basin, Central Asia, and Japan, the base of the Bashkirian is marked by the occurrence of the foraminiferal species *Plectostaffella varvariensis*, *P. jakhensis, P. posochovae*, and *P. bogdanovkensis*; however, the first primitive representatives of the genus appear slightly below the MCB (Kulagina et al., 1992).

The Moscovian (Nikitin, 1890) is designated as the second stage of the Pennsylvanian Epoch. The GSSP could not be established in the eponymous Moscow Basin, because of a large disconformity at the base of the succession. For additional regional information see Section 4.3.3. Originally the Moscovian Stage was defined by brachiopods, but, more recently, its base was defined by the fusulinid *Aljutovella aljutovica*. Because of fusulinid provinciality, one of the conodonts *Declinognathodus donetzianus* and *Idiognathoides postsulcatus* is more likely to be used for the global boundary (Groves and Task Group, 2004). These index fossils are applied also in the DCP 2003 (Fig. 3). Their FADs may be in a time span ≤ 0.5 my.

In the Soviet Union *A. aljutovica* was a commonly accepted index fossil which defined the base of the Moscovian Stage. It is found together with *D. donetzianus* in the Alyutovo Formation of the Vereian Substage of the Moscow Basin. *A. aljutovica* is used for correlations throughout the northern and eastern margins of Pangaea (Solovieva, 1986), but its potential for this purpose is restricted by fusulinid provinciality. Thus, the species composition in Eurasia and America is quite different (Solovieva, 1977; Ross and Ross, 1988; Davydov, 1996). Instead, the conodonts *D. donetzianus* and/or *I. postsulcatus* are proposed as index fossils because of their wide distribution in the Moscow Basin, Ural Mts., NW Europe, Spain, Canadian Arctic, Alaska, South America, and Japan. Moreover, their evolution is clearly recognized (Nemyrovska, 1999; Goreva and Alekseev, 2001).

This conodont based definition of the Bashkirian–Moscovian boundary does not coincide with the base of the Atokan Stage in North America as suggested by fusulinid studies (Miklukho-Maclay, 1963; Ross, 1979). The base of the Atokan, operationally defined with the FADs of the fusulinids *Eoschubertella* and *Pseudostaffella* (Sutherland and Manger, 1984; Groves, 1986) is significantly older than the base of the Moscovian proposed above. Groves et al. (1999) have suggested that the base of the Atokan lies within the Late Bashkirian, either within the late Tashastian, or at the base of the Asatauian Horizon (South Ural Mts.). The Bashkirian–Moscovian boundary coincides with either the base of the *Winslowoceras–Diaboloceras* ammonoid Genozone of the East European Platform and Ural Mts. (Ruzhentsev and Bogoslovskaya, 1978), or with the *Eowellerites* Genozone in western Europe and North America (Ramsbottom and Saunders, 1984) which may be slightly younger (Popov, 1979).

The Kasimovian Stage is designated as the third stage of the Pennsylvanian Epoch. This interval was separated from the upper part of the Moscovian in 1926 as the “*Teguliferina* Horizon”, a name changed later to Kasimovian Horizon (Dan’shin, 1947) and finally accepted as Kasimovian Stage (Teodorovich, 1949). Its recently nominated type area is in lower reaches of the Moskva River, about 80 km southeast of Moscow, but its name comes from the town of Kasimov in the Ryazan Region. For additional regional information see 4.3.3. The lower boundary was first defined on brachiopod evidence, and in 1962 it was selected at the base of the *Protriticites pseudomontiparus–Obsoletes obsoletus* foraminiferal Zone. In order to define the base of the Kasimovian in the DCP 2003, and to suggest index fossils for global correlation, the FODs of *P. pseudomontiparus, O. obsoletus* and the conodont *Swadelina subexcelsa* (Goreva and Alekseev, 2001) are chosen for the DCP 2003. The FAD of the above-mentioned fusulinids may be in a time span of ≤0.5 my.

However, in recent years both fusulinid and conodont workers have suggested selecting a higher level for the base of the Kasimovian, at the FODs of the fusulinid *Montiparus* and of the conodonts *Idiognathodus sagittalis* (Alekseev and Goreva, 2002; Villa and Task Group, 2004). Because their use would require a redefinition of both the Moscovian and Kasimovian stages, Davydov (2002) proposed the selection of an index species within the evolutionary chronocline of *Protriticites* fusulinids. This fauna is widely distributed throughout the Tethyan and Boreal realms and has also been found in the Great Basin within middle to late Desmoinesian successions (Wahlman et al., 1997; Davydov et al., 1999).

The Gzhelian Stage is designated as the fourth stage of the Pennsylvanian Epoch. It was introduced by Nikitin (1890) defined by a specific brachiopod assemblage, and with a stratotype about 60 km east of Moscow at the village Gzhel. For additional regional information see 4.3.3. The task of defining the base of the Gzhelian was accepted by the Moscovian/Kasimovian Boundary Task Group in 2002. From fusulinid data (Rauzer-Chernousova, 1941) its lower boundary coincides with the base of the *Rauserites rossicus–Rauserites stuckenbergi* Zone. Also the FAD of the fusulinids *Daixina*, *Jigulites*, and *Rugosofusulina* has been proposed as an operational index for the lower boundary of the Gzhelian Stage (Rozovskaya, 1975; Rauzer-Chernousova and Shchegolev, 1979; Davydov, 1990). These genera, however, do not occur in North America (except for the Canadian Arctic).

The FOD of the conodont *Idiognathodus simulator s. s.* is an useful marker because it is easily traced in Europe, Asia and North America (Boardman and Work, 2004; cf. 4.3.6). The *Shumardites–Vidrioceras* ammonoid Genozone has been conventionally placed at the base of the Gzhelian Stage (Bogoslovskaya et al., 1999). In the southern Ural Mts., advanced species of *Shumardites* occur in the late Gzhelian (Popov et al., 1985; Davydov, 2001). In order to define its base for the DCP 2003, and to suggest index fossils for global correlation, the conodont *I. simulator s.s.* and the foraminifer *R. rossicus* are chosen in this paper. The FAD of both species may be in a time span of <0.5 my.

The numerical age for the base of the Bashkirian Stage is between 323 Ma and ca. 318 Ma (Fig. 1) excluding the age of 314 Ma which was derived from the U–Pb SHRIMP age of 314.45±3.3 Ma from the Harewood borehole, middle Arnsbergian, middle Namurian A (Riley et al., 1993; cf. Menning et al., 2000). In the DCP 2003/STD 2002 an age of ca. 320 Ma is used, which has been derived by Menning et al. (2000) using the $^{40}Ar/^{39}Ar$ age of 324.6±8.0 Ma from the Jaklovec Member (Lippolt et al., 1984). This is allocated to approximately the Pendleian–Arnsbergian boundary (Weyer, in Menning et al., 2000: Fig. 6). When the $^{40}Ar/^{39}Ar$ age of 319.5 Ma±7.8 Ma of the Poruba Member (Lippolt et al., 1984) is used instead of the Jaklovec age, an age of 318 Ma or even younger is more plausible. Such an age of 318.1±1.3 Ma has been used by Davydov et al. (2004: Fig. 15.5) in the GTS 2004 (Table 2, Fig. 1), using both the above $^{40}Ar/^{39}Ar$ ages of Lippolt et al. (1984) and a SHRIMP age (using the SL 13 zircon standard; Riley et al., 1993) which was rejected by Menning et al. (2000).

Ages of 314 to 311 Ma are estimated for the base of the Moscovian Stage (Fig. 1). They are based on all the $^{40}Ar/^{39}Ar$ ages from coal tonsteins of Westphalian age of Central Europe (Lippolt et al., 1984). In the DCP 2003/STD 2002 an age of 312 Ma is used. An age of younger than 310.5 Ma can be excluded in the case of the base of the Moscovian being within the Westphalian B (Duckmantian Substage).

Ages of 306.5±1.0 Ma to 303 Ma are estimated for the base of the Kasimovian Stage (Fig. 1). They are based mainly on $^{40}Ar/^{39}Ar$ ages from coal tonsteins of Westphalian and Stephanian age of Central Europe (Lippolt et al., 1984; Burger et al., 1997). In the DCP 2003/STD 2002 an age of 305 Ma is used, which is an optimal estimate using the above-mentioned dates. The age of 306.5±1.0 Ma (Davydov et al. in the GTS, 2004) is an approach to make the largest possible timespan for the Kasimovian and Gzhelian stages, after the maximum age of the top of the Gzhelian was regarded as 299 Ma (Ramezani et al., 2003, ID-TIMS; cf. Section 5.2.1).

Ages of ca. 295 to ca. 304 Ma are estimated for the base of the Gzhelian Stage (Table 2, Fig. 1). The main reason for this relatively wide age variation is the poor biostratigraphic correlation between the southern Ural Mts. (marine, GSSP area) and Central/West Europe (continental, the current source of most of the isotopic age data). In the DCP 2003/STD 2002 an age of ca. 302 Ma is used. It is based on the ages of ca. 305 Ma and ca. 296 Ma for the bases of the Kasimovian and Asselian stages (Menning, 1989; Menning et al., 2000; Menning, 2001b) using Davydov's (1992) data that from geological indications, the Gzhelian should be more than twice duration of the Kasimovian.

4.3.2. Central and West Europe (D. Weyer, M. Menning, J. W. Schneider)

The classical Late Carboniferous successions of Central and Northwest Europe are mainly paralic, with an upward increase of terrigenous material culminating in exclusively continental beds. The stratigraphy is based on very detailed lithostratigraphy in coal basins (with marker horizons as marine bands, coal seams, volcanic ash layers), on macrofloral and miospore zonations, and only to a lesser degree on marine and non-marine faunal zones. Shallow water carbonate faunas (foraminiferids) are totally absent. The most suitable ammonoid faunas are restricted to the early Bashkirian, where they have an extremely fine biozonal resolution from Chokierian to Marsdenian substages (Bisat and Hudson, 1943; Ramsbottom et al., 1979; Ramsbottom and Saunders, 1984). The last Carboniferous conodont assemblage occurs in the early Moscovian (at the base of Bolsovian; =Aegir [Mansfield] marine band). It provides an inter-regional correlation across Europe. Thus, there is an approximate correlation between the base of the Vereian (earliest Moscovian) and the base of the Westphalian C (Bolsovian) (Nemyrovska, herein). However, in Fig. 3 the lower boundary of the Vereian is shown within the upper Westphalian B (Duckmantian), as based on Menning et al. (2001b). Therefore, a downward pointing arrow is shown at the base of the Westphalian C (Fig. 3).

Above this level, the macroflora (Remy and Remy, 1977; Kerp, 1988; Josten, 1991), palynomorphs (Clayton et al., 1977; Peppers, 1996), and terrestrial faunas (Haubold, 1970, 1973; Holub and Kozur, 1981; Schneider, 1982; Martens, 1983a,b, 1984; Schneider, 1985, 1996; Boy, 1987; Werneburg, 1989, 2001; Boy and Martens, 1991; Schneider and Werneburg, 1993; Hampe, 1994; Voigt, 2005) are useful for local and regional correlation, but these fossil groups rarely allow detailed intercontinental correlation. However, the carbonate successions of the Cantabrian Mts. (Spain) are correlated by fusulinids, conodonts, and ammonoids with those of eastern Europe (cf. Villa and Task Group, 2004). The Moscovian–Kasimovian boundary is within the Cantabrian (Fig. 3, arrow down at the base of the Cantabrian).

In the Subvariscan Foredeep, between Upper Silesia (South Poland) and South Wales (UK), the average cumulative thickness is >5000 m for rocks of Namurian and Westphalian age, compared with >3000 m for those of Bashkirian and Moscovian age (Drozdzewski, in Menning et al., 2000: Table 3). In Central Europe, coal seams are concentrated in rocks of Namurian C (Yeadonian Substage) to Westphalian C (Bolsovian Substage) age, e.g. the Ruhr district (Fig. 3), but they start also in the Namurian A (latest Mississippian) in Upper Silesia, and finish also in the early Rotliegend (Gzhelian/Asselian) in the intra-Variscan Saar-Nahe Basin and Saale Basin (STD 2002). In several depocenters, paralic sediments accumulated continuously with minimal gaps, of less than 0.5 my. In the Saar-Nahe Basin, however, with lacustrine, deltaic, and fluvial sediments, some 4 km thick, an "Asturian" gap of unknown, but in maximum 2.5 my, duration (Menning et al., 2005) separates Westphalian D from Stephanian A (Barruelian Substage). This gap could approximately correspond to the regional Cantabrian Substage (Stage, Wagner, 1972).

4.3.3. East Europe (A.S. Alekseev, B.I. Chuvashov, V.I. Davydov)

The Bashkirian Stage was proposed in the South Ural Mts. (Semikhatova, 1934), because it represented an interval of strata that was missing on the central East European Platform. Its base was defined, originally with the brachiopod *Choristites bisulcatiformus*, and later by the foraminifers *Plectostaffella bogdanovkensis*/*Pseudostaffella antiqua* (Semikhatova et al., 1979; Sinitsyna and Sinitsyn, 1987). After the GSSP for the MCB was defined in Nevada (Lane et al., 1999), the conodont *D. noduliferus* was selected also for the base of the Bashkirian in Russia. Thus, there is now total consistency between the base of the global Bashkirian Stage (Pennsylvanian Epoch) in North America and that of the regional Bashkirian Stage in East Europe.

The Bashkirian contains two sets of regional substages: on the platform the Voznesenskian, Krasnopolyanian, Severokeltmenian, Prikamian, Cheremshankian, and Melekessian horizons, and in the foredeep of the southern Ural Mts., the type area, the Bogdanovkian (*D. noduliferus*–*Idiognathodus sinuatus*), Syuranian, Akavasian (upper part: *I. sinuatus*–*Neognathodus askynensis*), Askynbashian (*Idiognathodus sinuosus*), Tashastian (*Declinognathodus marginodosus*), and Asatauian (*Neognathodus atokaensis*) horizons. Recently, the latter two were joined into a single Arkhangelskian substage (Kulagina and Sinitsyna, 2003).

On the eastern East European Platform shallow-marine carbonates dominate. In the western Ural Mts. are different facies: a) near shore sandstones, silts and widespread argillites with lenses of shallow water limestones; b) relatively deep water cherty argillites, marls, limestones, and sandstones incorporating the regional stratotype.

The Moscovian Stage was introduced by Murchison et al. (1845) and formally proposed by Nikitin (1890). In the type area, in the southern outskirts of Moscow, only the uppermost strata are exposed in quarries near the village of Myachkovo. In the southern part of the Moscow Basin outcrops of early Moscovian unconformably overlie limestones of Mississippian age.

The earliest Moscovian (Vereian Horizon) originally included the siliciclastic Shatsk Formation. Recent opinion favouring the base of the overlying fusulinid-bearing shallow-water limestones of the Alyutovo Formation to mark the base of the Moscovian (Makhlina et al., 2001), would relegate the Shatsk Formation to the Bashkirian. As in the Bashkirian, there are two main facies — shallow-water carbonates cover most of the East European Platform, an exception being the Vereian Horizon, which is dominated by clastic sediments, that include two marine intervals, one near its base and another near its top. On the platform, the stage is subdivided into the Vereian, Kashirian (including the Tsna Formation), Podolskian, and Myachkovian substages (Fig. 3). Near the Ural Mts. flysh-like sediments dominate. These are replaced to the west by more deep water argillites, marls, and limestones. The sediments are usually divided and correlated by fusulinids (Solovieva, 1977; Davydov, 1997) and less by conodonts (Goreva and Alekseev, 2001).

On the western part of the East European Platform the Kasimovian Stage is characterized by thick, terrigenous, relatively deep-water clastics, which alternate with shallow-water carbonates, whereas on the eastern part of the platform there are only carbonates. The stage is composed of the Krevyakinian, Khamovnikian, and Dorogomilovian substages (Fig. 3), whereas in the western Ural Mts. it consists of the Orlovian Horizon (foraminiferal zones: *Protriticites–Obsoletus*, *Montiparus;* conodont Zones: *Swadelina subexcelsa*, *Sw. makhlinae, Streptognathodus zethus–St. gracilis*) and the Kerzhakovian Horizon (foraminiferal zones: *Rauserites arcticus*, *R. firmus*; conodont Zone: *St. firmus*) (Chernykh, 2002).

On the platform the "*Teguliferina* Horizon" (see 4.3.1: Kasimovian) was originally recognized mainly by the presence of endemic bivalves and brachiopods. Later, Rauzer-Chernousova and Reitlinger (1954) defined the Kasimovian Stage with the FOD of the fusulinids *Obsoletes obsoletus* and *P. pseudomontiparus*. Recently, the base of Kasimovian has been placed in the Moscow Basin at the base of a local unit "sviniya" within the Peski formation which traditionally was included in the latest Moscovian Stage (Makhlina et al., 2001). This boundary coincides with an eustatic lowstand (Kabanov and Baranova, in press), and consequently with disconformities in several sections (Ross and Ross, 1988; Davydov, 1997; Ehrenberg et al., 1998; Davydov, 1999; Davydov and Krainer, 1999; Leven and Davydov, 2001). In the North American Midcontinent it occurs two major cycles below the previously proposed basal regional Missourian Stage boundary at the base of the Exline cyclothem in the Pleasanton Group (Heckel et al., 1998). The earliest Kasimovian (Krevyakinian Horizon) is well correlated with the latest Desmoinesian of North America (Heckel et al., 2002) (Fig. 3).

On the East European Platform the regional Gzhelian Stage is composed of the Dobryatinian, Pavlovoposadian, Noginskian, and Melekhovian horizons (substages). The Melekhovian Horizon is moved from the earliest Permian to latest Carboniferous (Makhlina and Isakova, 1997) because it predates the FOD of the index conodont *Streptognathodus isolatus* within the Aidaralash section selected as GSSP for the Asselian Stage (Davydov et al., 1998).

The traditional lower boundary of the Gzhelian Stage is at the FAD of the fusulinid species *R. rossicus* and *R. stuckenbergi* (Rauzer-Chernousova, 1941; Rozovskaya, 1950). However, in the type location near the village Gzhel, only *R. rossicus* has been recovered, although it is likely that *R. stuckenbergi* appears slightly higher in the section (Rauzer-Chernousova, 1958; Davydov, 1986). In the 1960s and 1970s the upper part of the Gzhelian was accepted as latest Carboniferous Orenburgian Stage (Fig. 3: informal substage) as proposed by Ruzhentsev (1945) using ammonoids.

In the western Ural Mts. the Gzhelian Stage consists of the Azantshian Horizon (foraminiferal zones *R. stuckenbergi* and *Jigulites jigulensis*; conodont zones *Streptognathodus simulator*; *S. vitali*; early *S.virgilicus*), the Mortukian Horizon (foraminiferal *Daixina sokensis* Zone; conodont zones late *S. virgilicus*, *S. simplex*, *S. bellus*), and the Nikolsky Horizon (foraminiferal Zone *Daixina* (*Ultradaixina*) *bosbytauensis–Schwagerina* (=*Globifusulina*) *robusta*, conodont Zone *S. longilatus–S. wabaunsensis*). The latest zones are considered also as of Permian age (Regional Stratigraphic Charts, 1994). The flysh-like terrigenous sediments in the east are replaced in westerly direction by relatively deep-water argillites, marls and cherts, and further to the west by carbonate deposits with reefs and bioherms.

4.3.4. Donets (T.I. Nemyrovska, V.I. Davydov)

The base of the Bashkirian Stage is defined in the Donets Basin by the FAD of the conodont *D. noduliferus* at the base of the Voznesenskian Subhorizon of the Olmezovian Horizon (limestone $D_{5\mathrm{upper}}^{8}$ of the Kalmius Section) (Fig. 3). Here, a thick succession

(1000–1200 m) of terrigenous clastic rocks with intercalations of thin limestones and coal seams, contains rich assemblages of marine and terrestrial fossils that allow correlation with the global Bashkirian Stage (Ayzenverg, 1969). Six conodont zones are distinguished in the Bashkirian: *Declinognathodus noduliferus*, *Idiognathoides sinuatus–Id. sulcatus, Id. sinuosis–Id. sulcatus parvus*, "*Streptognathodus" expansus*, and *Id. tuberculatus–Id. fossatus*, and *D. marginodosus* (Nemyrovska, 1999).

The controversial base of the Moscovian Stage coincides, according to the Resolutions (1990a) using foraminifers, with the base of the limestone K_3 of the Lozovian Horizon. The SCCS Task Group on the Bashkirian–Moscovian boundary proposed to use the FADs of the conodonts *D. donetzianus* or *I. postsulcatus* (Groves and Task Group, 2004). This would lower the boundary down to the base of limestone K_2. According to the macrofloral evidence the base of the Moscovian is above the sandstone with Duckmantian floras (e.g. Wagner and Higgins, 1979). Conodont evidence suggests an approximate correlation of the base of the Moscovian Stage, the Bolsovian/Westphalian C Substage in Central/West Europe, and the early/middle Atokan boundary of North America (Nemyrovska, 1999; Goreva and Alekseev, 2001). However, in the DCP 2003 the base of the Moscovian is within the Westphalian B/Duckmantian (STD 2002). The N_3–N_5^1 limestones of the Donbas, classified as latest Moscovian, the Krevyakinian Horizon of the Moscow Basin, classified as earliest Kasimovian, and the latest Desmoinesian of North America, can all be correlated using the conodont Zone *Sw. subexcelsa* (Fig. 3). Advanced *Protriticites* and *Obsoletes* are found in the N_5^1 and O_1 limestones (Davydov, 1990, 1992); their evolutionary development suggests a position at the base of the Kasimovian somewhat below the limestone N_5.

The FAD of *R. rossicus* in limestone O_5 indicates a correlation with the base of the Gzhelian. However, in the Donets Basin *R. rossicus* is represented by a phylogenetic group of primitive to advanced forms which appear up to limestone O_7. Thus, the possibility of a diachronous occurrence of the *R. rossicus* species-group in the Donets and Moscow basins cannot be excluded. Nevertheless, the fusulinid assemblage from limestone O_7 suggests a correlation with the upper Rusavkino Formation (earliest Gzhelian) in the Moscow Basin. Therefore, the base of the global Gzhelian could be between the limestones O_5–O_7 of the Donets Basin. A significant unconformity within the Gzhelian as used in the DCP 2003 spans most of its middle part, whereas a complete Gzhelian succession is in the Pre-Donets Trough (Davydov, 1992).

4.3.5. South China (X.-D. Wang)

In some areas of South China with shelf facies a major regressive event can be recognized at the end of the Fengningian (Mississippian) by the absence of late Dewuan (Serpukhovian) strata. Subsequent transgression occurred in South China during the early Luosuan Stage (Bashkirian) and resulted in the formation of a wide, uniform, shallow water platform, on which massive and pure limestones with diverse benthic fossils were deposited. Exposure surfaces (e.g. Lin et al., 2002) give evidence for several interruptions of the uniform deposition of shelf limestones during the late Hutianian (late Pennsylvanian). The Luodian Section of southern Guizhou, which is composed of wackestone interbedded with packstone, grainstone and cherty beds, representing a slope facies without any hiatus, contains continuous both conodont and fusulinid successions spanning the late Early Mississippian to the Guadalupian, and therefore it may serve as an effective linkage for the correlation between the successions of pelagic facies and shelf facies (Wang and Jin, 2003).

The Hutianian, introduced as a South China regional equivalent to the Pennsylvanian Epoch (Jin et al., 2000), includes the Weiningian and the Mapingian series. The Weiningian consists of the Luosuan, Huashibanian, and Dalaun stages, and the Mapingian consists of the Xiaodushanian Stage.

The Luosuan Stage (Rui et al., 1987) is from southern Guizhou, where its slope deposits comprises wackestones with interbedded packstones, grainstones and cherty beds. Its basal boundary is marked by the FOD of the conodont *D. noduliferus,* and the foraminifer *Millerella marblensis*, which coincides with the MCB. Wang and Qi (2003) subdivided this stage into nine conodont zones: *D. noduliferus*, *Idiognathoides sulcatus sulcatus*, *Id. sinuatus*, *Id. corrugatus–Id. pacificus*, *Neognathodus symmetricus*, *Idiognathodus primulus–N. symmetricus*, *Id. primulus–Neognathodus bassleri*, *Id. sulcatus parva*, and *Streptognathodus expansus*. The Luosuan fusulinids were assigned to the *Millerella marblensis–Eostaffella postmosquensis* Zone. The ammonoids were grouped as the *Homoceras* Zone in northwestern Guangxi (Ruan, 1981). The brachiopod assemblage with characteristic elements such as *Weiningia* dominates in the Luosuan shelf facies (Li, 1987).

The Huashibanian Stage was named from western Guizhou (Yang et al., 1980), where it consists of packstone, grainstone, and dolomite. Its base is marked by the appearance of the fusulinid *Pseudostaffella*.

Pelagic ammonoids from northwestern Guangxi belong to the *Reticuloceras*, *Cancelloceras*, and *Gastrioceras–Branneroceras* zones (Ruan, 1981). In the slope facies of southern Guizhou Wang and Qi (2003) recognized the conodont zones *Idiognathoides ouachitensis* and *Diplognathodus orphanus–D. ellesmerensis*.

The Dalaun Stage was named from SW Guizhou, where it comprises massive bioclastic limestones, intercalated with dolomites. It is subdivided into the *Profusulinella* and *Fusulinella–Fusulina* foraminiferal zones. Abundant ammonoids were found from the pelagic facies in western Guizhou and grouped as the *Winslowoceras decoratum* Zone and the *Owenoceras bellilineatum shuichengense* Zone (Yang, 1978). In the slope facies of southern Guizhou the highly diverse conodonts are grouped into eight zones: *Mesogondolella clarki–Idiognathodus robustus*, *Id. podolskensis*, *Streptognathodus nodocarinata*, *St. clavatulus*, *Streptognathodus cancellosus*, *St. gracilis–St. excelsus*, *St. guizhouensis*, and *St. simulator* (Wang and Qi, 2003).

The Xiaodushanian Stage was named from eastern Yunnan, where it consists of massive limestone (Zhou et al., 1987). Its basal boundary is distinguished by the appearance of *Protriticites*, while the top is marked by the FOD of the members of the Pseudoschwagerininae, a level roughly coincident with the global Carboniferous–Permian boundary. In most areas of South China it is subdivided into two fusulinid genozones, the *Montiparus* Zone and the *Triticites* Zone. The Xiaodushanian conodonts found in southern Guizhou contain the *Idiognathodus nashuiensis*, *Streptognathodus firmus*, *St. tenuialveus*, and *St. wabaunsensis* zones (Wang and Qi, 2003).

4.3.6. North America (P.H. Heckel)

The Mid-Carboniferous (Mississippian–Pennsylvanian) subsystem boundary was selected in Arrow Canyon, Nevada, at the first appearance of the conodont *Declinognathodus noduliferus*. The regional Morrowan Stage at the base of the Pennsylvanian Subsystem was named from northwestern Arkansas, where it comprises ~100 m of fossiliferous shale, sandstone, and limestone. It is largely confined to more basinal areas of the Midcontinent and Illinois basins, but includes sandstone and shale-filled palaeovalleys on higher parts of the shelf. The Morrowan coincides with the Zone of *Millerella* in which this primitive fusulinid occurs by itself. It is zoned using reticuloceratid and schistoceratid ammonoids, which allow correlation with the late Namurian and the Langsettian Substage of the early Westphalian in western Europe (Saunders et al., 1977). It is also zoned by conodonts, based on the work of H. R. Lane and coworkers, in ascending order: Early *D. noduliferus*, Late *D. noduliferus, Id. sinuatus*, *N. symmetricus, N. bassleri, Idiognathodus sinuosus*, *I. klapperi,* and *Idiognathoides convexus*. It is correlated with the early and middle global Bashkirian Stage using both conodonts and foraminifers by Groves et al. (1999).

The overlying regional Atokan Stage was named from southeastern Oklahoma, where it comprises up to ~5000 m of sparsely fossiliferous shale and sandstone that dominate the foreland Arkoma Basin in the southern Midcontinent. Both its lower and upper boundaries are poorly characterized biostratigraphically (Sutherland and Manger, 1984). Although its lower boundary was traditionally placed at the FAD of the fusulinid *Profusulinella* in other areas, transferral of the Trace Creek Member from the top of the type Morrowan to the base of the Atokan (cf. Zachry and Sutherland, 1984) led Groves (1986) to define the base of the Atokan at the first appearance of the primitive fusulinids *Eoschubertella* and *Pseudostaffella*. This placed the *Diaboloceras* ammonoid Zone in the basal Atokan, which made the lower Atokan equivalent to the Duckmantian Substage (Westphalian B) in western Europe. The Atokan now encompasses three fusulinid genozones, in ascending order: *Eoschubertella–Pseudostaffella, Profusulinella*, and *Fusulinella*. Conodont faunas are not yet well enough known to define the base of the Atokan, but zones based on *N. atokaensis* and *N. colombiensis* (the latter zone including *N. bothrops*) characterize the middle and late parts, respectively (Barrick et al., 2004).

The overlying regional Desmoinesian Stage was named from central Iowa. It comprises ~160–250 m of cyclic shale, coal, limestone, and sandstone across the northern Midcontinent shelf, thickening southward into the basin. It coincides with the Zone of *Fusulina* (now *Beedeina*), with *Wedekindellina* also characterizing the lower half. Boardman et al. (1994) recognized it to comprise the *Wellerites* ammonoid Genozone overlain by the *Eothalassoceras* Genozone. Barrick et al. (2004) recognized three ascending conodont zones based on *Neognathodus* (*caudatus, asymmetricus, roundyi*), and five zones based on idiognathodids (*Idiognathodus obliquus, I. iowaensis, I. delicatus, Swadelina neoshoensis, Sw. nodocarinata*). The base of the Desmoinesian has not yet been formally defined in the Midcontinent, but in the most complete succession in NE Oklahoma, *I. praeobliquus* followed by *I. obliquus* appear not far above the traditional boundary (Boardman et al., 2004). The two *Swadelina* zones at the top indicate correlation with the Krevyakinian Substage, which is currently the base of the global Kasimovian

Stage. The early and middle Desmoinesian are equivalent to the late global Moscovian Stage. The Desmoinesian is also characterized by arborescent lycopods and their palynomorphs (Peppers, 1996), and is roughly equivalent to the latest substage (D, now named Asturian) of the Westphalian in W Europe.

The overlying regional Missourian Stage was named from NW Missouri, and it comprises ~150–200 m of cyclic limestone and shale (with some sandstone) across the northern Midcontinent shelf, thickening southward into the basin. It coincides with the lower part of the Zone of *Triticites*, although that genus first appears only in the fourth transgressive–regressive cycle above the base, above the first appearance of *Eowaeringella* in the third cycle. The two lower cycles lack large fusulinids (Heckel, 1999). Boardman et al. (1994) recognized in ascending order the *Pennoceras, Preshumardites*, and *Pseudaktubites* ammonoid genozones, with the latter zone extending above the top. Heckel et al. (2002) designated the base of the Missourian at the FAD of the conodont *Idiognathodus eccentricus* in the Exline cyclothem, and Barrick et al. (2004) recognized five ascending Missourian conodont zones (*I. eccentricus, St. cancellosus*, *S. confragus, S. gracilis*, and *I.* aff. *simulator*). Missourian palynomorphs suggest correlation with the Stephanian of western Europe, but the exact correlation of its basal Cantabrian Substage is not yet established.

The regional Virgilian Stage at the top of the Pennsylvanian was named from SE Kansas, and it comprises ~420 m of cyclic limestone, shale and sandstone across the northern Midcontinent shelf. It coincides with the upper part of the *Triticites* Zone, and it contains *Waeringella* near the base, *Dunbarinella* through the middle, and *Leptotriticites* throughout the top (G.P. Wahlman, manuscript in review). It now includes strata that had been called "Bursumian" and the basal part of the Wolfcampian (Foraker Limestone) below the FAD of the conodont *S. isolatus*, which now defines the global base of the Permian System/Period. Boardman et al. (1994) referred the earliest Virgilian to the late part of the *Pseudaktubites* ammonoid Genozone and the remainder to the *Shumardites* Genozone. Heckel (1999) tentatively defined the base of the Virgilian at the first appearance of the conodont *St. zethus* in the base of the Cass cyclothem, and Barrick et al. (2004) recognized five ascending Virgilian conodont zones (*St. zethus, I. simulator, St. virgilicus, St. brownvillensis*, and *St. wabaunsensis*). The best position for defining the base of the global Gzhelian Stage appears to be at the first appearance of *I. simulator*, which coincides with the base of the *Shumardites–Vidrioceras* global ammonoid Genozone (Boardman and Work, 2004), and is above the base of the regional Virgilian Stage.

5. Permian GSSPs and Regional Stratigraphic Scales

5.1. General (M. Menning, T.A. Grunt, G.V. Kotlyar, H. W. Kozur)

The Permian System/Period (Murchison, 1841) is traditionally subdivided into the Lower Series and Upper Series (Early Epoch and Late Epoch). In East Europe the Early Permian consists of the Asselian, Sakmarian, Artinskian, and Kungurian stages and the Late Permian, starting at the base of the Solikamskian Horizon, includes the Ufimian, Kazanian, and Tatarian stages (Resolutions, 1965). These seven (supra)regional stages were also used as global stages (Fig. 1), although the base of the Ufimian was not well defined in large areas of the East European Platform (Grunt, 2005).

The Permian of Central/West Europe, the Dyas (Marcou, 1859; Geinitz, 1861–1862), comprises the groups Rotliegend and Zechstein. The Rotliegend starts in the latest Carboniferous Gzhelian Stage and continues into the Late Permian Wuchiapingian Stage (Fig. 4). Therefore, it cannot be considered as "Lower Permian" as widely mentioned in the literature. The Zechstein can be correlated with the late Wuchiapingian and most (>95%) of the Changhsingian Stage. The overlying Buntsandstein Group of the Germanic Trias starts in the latest Changhsingian Stage (Kozur, 1999) (Fig. 4).

The Permian of China currently consists of the Chuanshanian, Yangsingian, and Lopingian regional series/epochs (Sheng and Jin, 1994). The Chuanshanian Series plus the Chihsian Subseries correspond to the Cisuralian Series (Fig. 4). The Permian of North America consists of the Wolfcampian, Leonardian, Guadalupian, and Ochoan regional series (Dunbar et al., 1960; Fig. 4).

Jin et al. (1994) proposed officially the tripartite Permian scale which is now the internationally accepted, global standard (Fig. 4). The Middle Permian (Guadalupian Epoch) of this scale starts within the Late Permian (Ufimian, Kazanian, Tatarian) of the bipartite scale. According to Grunt (this work), the bipartite scale corresponds better to the biotic evolution than the tripartite scale. Therefore, Leven (2003) proposed to reactivate the global bipartite scale, by classifying its parts as subsystems (Cisuralian and Tethyan) each divided into two series (Fig. 4). The same structure utilizing proper names for subsystems and series has been proposed for the East-European Permian Scale (Grunt, 2005; Fig. 4): the Lower Permian Subsystem (Cisuralian) includes the Uralian Series (Asselian and Sakmarian stages) and the Cistimanian Series (Artinskian and Kungurian regional stages). The Upper Permian (Vyatkian) Subsystem consists of the Biarmian Series with the regional Ufimian, Kazanian, and Urzhumian stages and the Tatarian Series with the Midian, Dzhulfian, and Changhsingian stages.

However, the introduction of a new stratigraphic category (subsystem), the re-classification of e.g. the Cisuralian (from a series to a subsystem) as well as of the Tatarian (from a stage into a series), and the significantly reduced stratigraphic range of the Tatarian (excluding the Urzhumian Horizon) would in consequence lead to even more confusion than to more clarity (cf. Fig. 4, 5.3.3) (Menning and Kozur (this work). Moreover, the strong extinction event at the Guadalupian–Lopingian boundary evidenced in many fossil groups shows that the differences between the Guadalupian and Lopingian faunas are much stronger than those between the Cisuralian and Guadalupian faunas (e.g. conodonts, fusulinids). Therefore, the general pattern of biotic evolution corresponds better to the tripartite scale than to the bipartite scale (Kozur, this work).

5.2. Asselian, Sakmarian, Artinskian, and Kungurian stages (Cisuralian, Early Permian Epoch)

5.2.1. GSSPs and numerical ages (V.I. Davydov, B.I. Chuvashov, M. Menning)

The Cisuralian Epoch/Early Permian Epoch/Lower Permian Series has been formally established by Jin et al. (1994). Its type area is the southern Ural Mountain region (Urals), comprising the Asselian, Sakmarian, Artinskian, and Kungurian stages. These stages were initially defined and widely recognized on ammonoid phylogenies (Karpinsky, 1891; Ruzhentsev, 1937, 1950, 1951, 1955, 1956). However, the boundaries and subdivisions of those stages were established using fusulinaceans, the most abundant, and one of the best studied Late Palaeozoic fossil groups of the southern Urals. It is expected that the boundary stratotypes (GSSP) for all Cisuralian stages will be formally established in this region.

The GSSP for the base of the Asselian Stage, the co-incident base of the Permian Period, is established at Aidaralash Creek, Aktöbe (formerly Aktyubinsk) region, northern Kazakhstan. The position of the GSSP is at the FOD of the conodont *S. isolatus* (Davydov et al., 1998). The FOD of *S. invaginatus* and *S. nodulinearis* nearly coincide with the FOD of *S. isolatus* and therefore, can be used as accessory indicators for the boundary. The GSSP is 6.3 m below the traditional fusulinid boundary, i.e. the base of the *Sphaeroschwagerina vulgaris aktjubensis–S. fusiformis* Zone (Davydov et al., 1998). The latter can be widely correlated via Spitsbergen, the East European Platform, the Ural Mts., Central Asia, China and Japan. It is of practical value to identify the top of the informal Orenburgian Substage of the late Gzhelian Stage. The traditional ammonoid boundary, 26.8 m above the GSSP, includes the termination of the *Prouddenites–Uddenites* lineage and the introduction of the Permian taxa *Svetlanoceras primore* and *Prostacheoceras principale*.

At the Usolka Section numerous volcanic ash beds offer great potential for radiometric dating (Davydov et al., 2002). The Asselian conodont succession of Aidaralash and Usolka is also displayed in the Red Eagle cyclothem of the Midcontinent (Boardman et al., 1998), the Gaptank Formation in the Wolfcamp Hills, West Texas (Wardlaw and Davydov, 2000; Fig. 4), and in China (Wang, 2000). A zone of normal polarity in most of the *Ultradaixina bosbytauensis–Schwagerina robusta* fusulinid Zone (Khramov and Davydov, 1993), just below the CPB in Aidalarash could correlate with the questionable normally magnetized zone in the Manebach Formation of the Thuringian Forest in Germany (Menning, 1987).

The stratotype of the Sakmarian Stage (Ruzhentsev, 1950) and its boundary GSSP (Chuvashov et al., 2002b) were proposed at the Kondurovka section (Karamuruntau Ridge, southern Orenburg Province). It was divided on the basis of its ammonoid and fusulinid assemblages (Rauzer-Chernousova, 1965). However, no ammonoids are known in the early Sakmarian, and thus, the base has been traditionally defined by the FAD of *Schwagerina moelleri* (=*Pseudofusulina moelleri* in Russian nomenclature). This definition is effective only in the Boreal and in the western Tethyan realms because of provinciality. A conodont chronomorphocline exhibits the evolutionary change from *Sweetognathus expansus* to *Sweetognathus merrilli* at 115 m above its base (Chuvashov et al., 2002b). The first joint occurrence of both *Mesogondolella uralensis* and *Sw. merrilli* is an accessory indicator for the boundary. *Sw. merrilli* is a widespread taxon, and its FAD is well constrained throughout Kansas in the upper part of the Eiss Limestone of the Bader cyclothem (Fig. 4).

The Artinskian Stage (Karpinsky, 1874) included all the Late Palaeozoic clastic deposits of the Pre-Uralian Foredeep (Artinskian Sandstone of Murchison et al.,

1845) overlying Carboniferous carbonates. Later, the Artinskian Stage was restricted to its upper part (Ruzhentsev, 1954). Recently, the uppermost part, the Saranian Horizon, has been referred into the Kungurian (Chuvashov et al., 1999; Chuvashov and Chernykh, 2000; Resolutions, 2006). Thus, the Artinskian of Karpinsky (1874) has been reduced two times: significantly at its base and moderately at its top (Fig. 4). Thus, the upper part of the Sakmarian Stage is composed of dark-coloured marl, argillite, or less commonly, detrital limestone with fusulinids, radiolarians, rare ammonoids, and bivalves.

The Artinskian stratotype section is in a succession of sandstone quarries on Mount Kashkabash near the Arti village at the River Ufa (Chuvashov et al., 2002c). The Artinskian is characterized by clastic-rich deposits. Ammonoids, best represented in the Pre-Uralian Foredeep of the southern Ural Mts. (Ruzhentsev, 1954), and fusulinids, best known from the eastern margin of the East European Platform (Rauzer-Chernousova, 1949), are provincial and, therefore, conodonts are the favoured GSSP fossils.

The Sakmarian–Artinskian boundary deposits are well represented in the Dalny Tyulkas (=Tulkus) section which is suggested as stratotype (GSSP candidate) for the base of the Artinskian. A potential GSSP is located within a few meters of the FAD of *Sweetognathus whitei*, which is sometimes associated with the conodont *Mesogondolella bisselli*. Several levels yielded radiolarians of the *Enactinosphaera crassicalthrata–Quinqueremis arundinea* Zone (Chuvashov et al., 2002c). According to Kozur (this work), the *Trilonche* (not *Enactinosphaera*) *crassicalthrata–Q. arundinea* Zone is a junior synonym of the *Patrickella cathaphracta* Zone (Kozur and Mostler, 1989). The definitive chronomorphocline of *Sweetognathus binodosus* to *S. whitei* also can be recognized in the lower Great Bear Cape Formation on southwestern Ellesmere Island, in the Sverdrup Basin, Canadian Arctic (Henderson, 1988; Beauchamp and Henderson, 1994; Mei et al., 2002), and in the Schroyer to Florence limestones of the Chase Group in Kansas (Wardlaw et al., in press) (Fig. 4).

Chuvashov and Chernykh (2000) proposed to define the base of the Kungurian Stage at the FAD of *Neostreptognathodus pnevi* within a chronomorphocline from advanced *Neostreptognathodus pequopensis*. This was later accepted (Resolutions, 2006). The proposed stratotype section (Chuvashov et al., 1993, 2002c) is exposed along at the Yuryuzan River downstream of the Mechetlino settlement. Beds 13–17 of the Sargian Horizon Gabdrashitovo Formation (1–18) include typical Artinskian ammonoids, fusulinids, and conodonts like *Neostreptognathodus kamajensis*, *N. pequopensis*, and *N.* aff. *ruzhencevi*. The basal part of bed 19 (earliest Saranian Horizon) yield *Neostreptognathodus clinei* (Kozur, this work: *pseudoclinei*, is partly assigned to the Roadian *N. clinei*), *N. pnevi*, *N. kamajensis*, *N. pequopensis*, and *Stepanovites* sp.

The defining chronomorphocline can also be recognized in the upper Great Bear Cape Formation and upper Trappers Cove Formation on southwestern Ellesmere Island, Sverdrup Basin, Canadian Arctic (Henderson, 1988; Beauchamp and Henderson, 1994; Mei et al., 2002).

The proposed global Kungurian is significantly extended, in relation to the supraregional Kungurian (restricted on the Filippovian and Irenian horizons), by the inclusion of the former Artinskian Saranian Horizon and the Ufimian, Solikamskian, and Sheshmian horizons. Thus, in contrast to the GTS 2004, the global Kungurian may be the longest global Permian stage (cf. Fig. 4).

Numerical ages of 290 to 299±1.0 Ma are given for the base of the Asselian Stage (Fig. 1). An age of 296 Ma (Menning, 1989; ff.) is used in the DCP 2003/STD 2002, which is derived from ages from Central Europe (Lippolt and Hess, 1983; Lippolt et al., 1984; Lippolt and Hess, 1989; Lippolt et al., 1989). This age range encompasses those previously suggested viz., ca. 292 Ma (Chuvashov et al., 1996, SHRIMP; cf. Menning et al., 2000: Fig. 7), 298 Ma (Jones, 1995, Central European isotopic ages), and the 299 Ma (Ramezani et al., 2003, ID-TIMS) recently used in the GTS 2004 (Wardlaw et al.) as a maximum age. On one hand, the 299 Ma reduces the timespan for the Kasimovian and Gzhelian stages to a maximum of 7.5 my (306.5–299 Ma; Section 4.3.1). On the other hand, it significantly extends the timespan for the Cisuralian Period to 36.6–40 my. Consequently, the Early Permian stages are also extended. There is more than sufficient time to insert the Early Permian successions into the Early Permian time scale.

Numerical ages for the base of the Sakmarian, Artinskian, and Kungurian stages are estimated using mainly geological time indications, because there are no U–Pb ID-TIMS ages from the type area. Isotopic ages from Central and West Europe as well as from Australia (Roberts et al., 1996), are difficult to transpose into the East European Reference Scale, because of weak biostratigraphic correlations. The SHRIMP ages of Chuvashov et al. (1996) may be too young in relation to the "ID-TIMS Time Scale" which currently appears to be in greater use. ID-TIMS ages are needed to decide which time spans are more reliable, for the Asselian Stage [the ca. 6 my of the DCP 2003 or the 4.4±0.8 my of the GTS 2004 (Wardlaw et al.)]; for the Sakmarian Stage [the ca. 6 my of the DCP 2003 or the 10.2±0.8 my of the GTS 2004 (Wardlaw et al.)], for the Artinskian Stage [the ca. 4.5 my of the DCP 2003 or the 8.8±0.7 my of the GTS 2004 (Wardlaw et al.)], and for the Kungurian Stage [the ca. 7 my of the DCP 2003 or the 5.0±0.7 my of the GTS 2004 (Wardlaw et al.) (Table 2, Fig. 1).

The DCP 2003/STD 2002 time scale is based on Menning (2001). The age of ca. 284 Ma for the Sakmarian–Artinskian boundary results from the SHRIMP ages of 280.3±2.6 from the latest Sakmarian and of 280.3±2.4 Ma from the earliest Artinskian (Chuvashov et al., 1996) and from a correction of these ages, according to Compston (2000), of ca. 1.3%.

5.2.2. Central and West Europe (M. Menning, H.W. Kozur, J.W. Schneider)

According to the STD 2002 the diachronous lower boundary of the Rotliegend Group starts at ca. 302 Ma, an age based on a 302±3.0 Ma U–Pb SHRIMP-II zircon age from the Holzmühlenthal ignimbrite (Breitkreuz and Kennedy, 1999), which is allocated to the lowermost Rotliegend of the Central European Basin (Fig. 4). Also SHRIMP-II ages between 300±3 Ma and 297±3 Ma were found in Rotliegend volcanics from six deep bore holes in that basin. From the Saar-Nahe Basin an ^{40}Ar/^{39}Ar age of 300.0±2.4 Ma is cited (but not published in detail) for the "Stephanian/Rotliegend boundary" (Burger et al., 1997). Earliest Rotliegend volcanics from the Thuringian Forest (Möhrenbach Fm., Georgenthal Fm.=lower Gehren Schichten) yielded variable data: 299.1±5.1 Ma to 292.0±2.4 Ma ^{40}Ar/^{39}Ar ages, an 299.8±4.4 Ma Rb–Sr model age (Goll et al., 1996) and ^{40}Ar/^{39}Ar ages from 295±3 Ma to 291.0±2 Ma (Goll and Lippolt, 2001). Thus, the age of ca. 302 Ma for the base of the Möhrenbach Formation (Fig. 4, STD 2002), which was derived by also taking into account the above ages of 302±3.0 Ma, 300±3 Ma to 297±3 Ma, and 300.0±2.4 Ma, is slightly too old (Fig. 4).

The "Lower" and "Upper" Rotliegend are mapping units in Central Europe. The stratigraphic position of the boundary between these units changes in different locations. Thus, it corresponds traditionally to the boundary between the Rotterode and Tambach formations (sometimes, recently, at the Oberhof–Rotterode boundary) in the Thuringian Forest (Fig. 4, ca. 285–284 Ma, STD 2002) and the Glan and Nahe subgroups (former Tholey/Grenzlager boundary) in the Saar-Nahe Basin (Fig. 4, ca. 292 Ma, STD 2002). Thus, the use of the terms "Lower" and "Upper" Rotliegend is under discussion, whereas there is no question about the use of the term Rotliegend and its well defined formations and subgroups (Fig. 4, STD 2002). "Lower Rotliegend" successions have their maximum duration in the Thuringian Forest (≤302 Ma to ~285 Ma). Nevertheless, they span a significantly shorter time interval than the "Upper Rotliegend", which starts earliest in the Saar-Nahe Basin at ~292 Ma with the Nahe Subgroup and finishes in the Central European Basin with the top of the Elbe Subgroup at ~258 Ma (Fig. 4). The overlying Zechstein Group starts with the Kupferschiefer (Copper Shale, Marl Slate), a most useful lithostratigraphic marker horizon between Lithuania in the east and the United Kingdom in the west.

Furthermore the use of the terms "Autunian", "Saxonian", and "Thuringian" in a time sense is not advocated, because of their widely variable application and poor definition; otherwise, their citation should be accompanied by the name of the author and year of publication (Schneider, 2001). Even the definition of the base of the "Autunian" with the FOD of *Autunia* (former *Callipteris*) *conferta* is flawed, because Doubinger (1956) and Broutin et al. (1990) have shown that in successions several hundred meters thick, alternations between "Autunian" rocks with *A. conferta* (meso- to xerophylous) and "Stephanian" rocks (hygro- to hydrophylous) occur. The base of the "Saxonian" was first defined by Haubold and Katzung (1972), but the new "index" fossils are found only in one outcrop of the Tambach Formation of the Thuringian Forest (Fig. 4). Often the term "Saxonian" is synonymously used for "Upper Rotliegend", meaning that the "Saxonian" extends up to the base of the Kupferschiefer. For more details see Kozur (1980). A "Thuringian" age is usually cited by palynostratigraphers, using palynomorphs, which are seldom found in the almost exclusively red coloured beds of monotonously thick successions of late Rotliegend. The macroflora, also, lacks the time sensitivity to serve as diagnostic age markers in the middle and late Permian of Central and West Europe. Therefore, the positions of the age proposed for the base of the Thuringian are highly variable, ranging from about the early Wellington Formation (early Leonardian; global early Kungurian; Remy, 1975) up to the base of the Zechstein Group (Wuchiapingian) (Fig. 4).

The Rotliegend usually begins with thick deposits of volcanic rocks or coarse grained sediments. It consists of very variable successions of coarse to fine grained clastics of fluvial, lacustrine (including black shales and carbonates), aeolian, and sabkha origin. The oldest beds are often grey coloured, whereas the late Rotliegend is exclusively red coloured except for some thin limnic horizons in the Central European Basin (Southern Permian Basin).

On one hand, the Rotliegend starts in the Gzhelian (Kozur, 1977, 1978b; Menning, 1986; Boy, 1987; STD 2002) using a) biostratigraphic indications, b) a numerical age of 296 Ma for the Carboniferous–Permian boundary (Menning, 1989, 1995a,b, 2001), and c) the isotopic ages from the Saar-Nahe Basin (Figs. 3 and 4). The latter (Lippolt and Hess, 1989; Burger et al., 1997; Königer et al., 2002) are systematically older than those of the Thuringian Forest. On the other hand, the Rotliegend starts in the earliest Asselian (Schneider et al., 2003; Lützner et al., in press) using a) an age of 299 Ma age for the CPB (Ramezani et al., 2003), b) an age of ca. 297 Ma for the base of the Möhrenbach Formation of the Thuringian Forest (Lützner et al., in press: Fig. 5), and c) the error bars of isotopic data of the Saar-Nahe Basin to get minimum ages. This example shows that the correlation to East Europe is based on isotopic age data, rather than on biostratigraphic evidence like the FOD of the pteridosperm *A. conferta*, with its FOD in the early Early Rotliegend of the Thuringian Forest (Gehren Schichten), and in the Middle Gzhelian of the Donets Basin.

The Cisuralian–Guadalupian boundary cannot be recognized in the Rotliegend. In most sections large gaps of unknown duration and changing positions are present within the late Rotliegend (Holub and Kozur, 1981; Menning et al., 1988; Schneider, 2001; STD 2002; Menning et al., 2005).

5.2.3. East Europe (B.I. Chuvashov, V.I. Davydov, T.A. Grunt, G.V. Kotlyar)

The (supra)regional Asselian Stage was established by Ruzhentsev (1954). It has sometimes included the Melekhovian Horizon, but since the introduction of the GSSP concept (base Asselian = FAD of the conodont *S. isolatus*), this horizon has been placed in the global Gzhelian Stage (cf. Section 4.3.3). Ruzhentsev distinguished the Asselian by an ammonoid assemblage which, in the Ural Mts., corresponds to the entire range of the fusulinid *Sphaeroschwagerina*. Recently, the regional Asselian Stage was redefined to a reduced Kholodnologian Horizon and the Shikhanian Horizon (East European Platform). Fig. 5 presents its fusulinid and conodont zones. Two ammonoid zones (*S. primore* and early *Sv. serpentinum*) correspond to the Kholodnologian Horizon, however, their geographic distribution elsewhere has yet to be confirmed.

The (supra)regional Sakmarian Stage (Chuvashov et al., 2002b), and the proposed global one, has its stratotype section at the Sakmara River near the Kondurovka Station (Chuvashov et al., 1993b), and a parastratotype at the Usolka Section (Chuvashov et al., 1993a), where a complete succession of deep-water conodonts and numerous volcanic ash beds are present. Thus, the *Mesogondolella uralensis* conodont Zone is useful in the Tastubian (Chernykh and Chuvashov, 2004). In shallow water facies the Tastubian Horizon contains two foraminiferal zones, and one conodont Zone (Chuvashov et al., 2002a; Fig. 5) The Sterlitamakian Horizon is defined by one foraminiferal Zone, and one conodont Zone (Chernykh and Chuvashov, 2004; Fig. 5). The late Tastubian corresponds to the *Synartinskia principalis* and the Sterlitamakian to the *Andrianovia sakmarae* ammonoid zones (Chuvashov et al., 2002b; Fig. 5).

The fusulinid and conodont zones of the original (supra)regional Artinskian Stage are shown in Fig. 5. Corresponding ammonoid zones are *Neoshumardites triceps* (Burtsevian), *Aktubinskia notabilis* (Irginian), and *Neocrimites fredericksi* (Sargian). But, Chuvashov et al. (1999) suggested the Saranian Horizon should be referred to the revised regional Kungurian Stage by the selection of the same conodont, *Neostreptognathodus pnevi*, as the index-fossil for the base of the global Kungurian (Kozur, 1995). This change has now been accepted (Resolutions, 2006). However, the ammonoid evolution in the Irginian and Sargian continues into the Saranian (Bogoslovskaya and Shkolin, 1998). Furthermore, the brachiopods of the Saranian and Sargian horizons are similar, whereas they are remarkably different from those of Kungurian assemblages. Thus, these data do not support the inclusion of the Saranian Horizon into the global Kungurian (Bogoslovskaya and Grunt, 2004). The ammonoid and conodont zonations are established in the Pre-Uralian Foredeep in siliciclastic or mixed carbonate-siliciclastic facies.

The original (supra)regional Kungurian Stage (Stuckenberg, 1890; Resolutions, 1990b) consists of the Filippovian Horizon, and the overlying ammonoid-bearing Irenian Horizon (e.g. *Uraloceras fedorovi*, *U. sofronitskyi*, *U. alekense*, *U. tchuvaschovi*, *Paragastrioceras* ex. gr. *suessi*, *Thalassoceras gemellaroi*, *Baraioceras kungurense*, *Paragastrioceras jossae*, whereas the proposed global Kungurian includes five horizons: Saranian, Filippovian, Irenian, Solikamskian, and Sheshmian (Figs. 4 and 5). The regional Kungurian Stage corresponds to the beginning of a huge intraregional regression and, consequently, it is represented partly by terrestrial and/or sabkha facies. Essentially marine carbonates, marls, argillites and sandstones occur at the eastern rim of the Kungurian Basin. Its

Regional Stage 1990	eastern East European Platform	Pre-Uralian Foredeep	Foraminifers	Conodonts	Global Stage 2005
Ufimian	Sheshmian		---	---	Kungurian
	Solikamskian		---	---	
Kungurian	Irenian	Koshelevka Fm. Absal Fm.	---	---	
	Filippovian	Mysova Fm.	*Nodosinelloides pugioidea – Pseudofusulina sp.*	---	
Artinskian	Saranian	Gabdrashitovo Fm.	*Hemigordius saranensis*	*Neostreptognathodus pnevi*	
	Sargian		*Parafusulina solidissima*	*Neostreptognathodus pequopensis*	Artinskian
	Irginian	Yangantau Fm.	*Pseudofusulina juresanensis – Parafusulina lutugini*	*Neostreptognathodus clarki – Neostreptognathodus ruzhenzevi*	
	Burtsevian	Balzyak Fm.	*Ps. pedissequa – Ps. concavutas*	*Sweetognathus whitei*	
Sakmarian	Sterlitamakian	Kondurovka Fm.	*Pseudofusulina urdalensis*	*Sweetognathus primus*	Sakmarian
		Maloik Fm.	*Pseudofusulina callosa*	*Sweetognathus anceps*	
	Tastubian 1990 2005	Sarabil Fm. Karamurun Fm.	*Pseudofusulina verneuili* *Schwagerina ("Ps.") moelleri*	*Sweetognathus merrilli* *Diplognathodus* aff. *stevensi – Sweetognathus* aff. *merrilli – Mesogondolella uralensis*	
Asselian	Shikhanian	Kurmaya Fm.	*Sphaeroschwagerina sphaerica – Schwagerina ("Gl.") firma*	*St. barskovi – St. postfusus* *Streptognathodus constrictus*	Asselian
	Kholod-nologian	Uskalyk Fm.	*Schwagerina ("Gl.") nux* *Sphaeroschwagerina fusiformis*	*St. cristellaris – St. sigmoidalis* *Streptognathodus glenisteri*	
		Syuren Fm.	*Sph. vulgaris aktjubensis*	*Streptognathodus isolatus*	
	Melekhovian	Orenburgian	*Ultradaixina bosbytauensis – Schwagerina robusta*	*Streptognathodus wabaunsensis* *Streptognathodus fissus*	Gzhelian
Gzhelian	Noginskian		*Daixina sokensis* *Daixina enormis*	*Streptognathodus bellus* *Streptognathodus simplex*	

Gl.-Globifusulina, Ps.-Pseudofusulina, Sph.-Sphaeroschwagerina, St.-Streptognathodus,

Fig. 5. Early Permian (Cisuralian) zonations of foraminifers and conodonts on the East European Platform and in the Pre-Uralian Foredeep.

correlation potential is limited to the Ural Mts. However, when the Saranian Horizon is included (Resolutions, 2006), the base of the Kungurian is close to the base of the (supra)regional North American Leonardian Series (Fig. 4).

The (supra)regional Ufimian Stage includes the Solikamskian and Sheshmian horizons. In its stratotype area the Solikamskian Horizon (120–130 m) is composed of grey limestones, dolomites, marls, mudstones, siltstones, and sandstones containing rare beds and lenses of mineral salt, gypsum, and anhydrite. The Liur-yaga section (northeastern Pai-Khoi Ridge) is important for clarifying the lower boundary, where mostly primitive species of the ammonoid genus *Epijuresanites* (*E. primarius*) replace Kungurian *Tumaroceras dignum* and *Medlicottia postorbignyana* (Kotlyar, 2002). *E. primarius* is associated with the index brachiopod *Sowerbina granulifera* (Bogoslovskaya and Grunt, 2004). This section could be a suitable stratotype for the base of a supraregional Upper Permian Subsystem (and the Ufimian Stage, Grunt, 2005). *S. granulifera* (Solikamskian) is reported from the Kanin Peninsula, Russia (Grunt et al., 2002), and the lower Kapp Starostin Formation (Voringen Member) of Central Spitsbergen (Nakamura et al., 1992), where the conodont evidence suggests a correlation with the Mead Peak Member of the Phosphoria Formation (?early Roadian) or the late Leonardian Victorio Peak Member of the Bone Spring Limestone of SW North America (Szaniawski and Malkowski, 1979; cf. DCP 2003). Therefore, the Solikamskian Horizon, the beds with *S. granulifera* of the Kanin Peninsula and Spitzbergen, and the Victorio Peak Member (latest Leonardian) may be of comparable age.

Thus, the (supra)regional Ufimian Stage is mostly, if not entirely, older than the Roadian Stage, and therefore, Grunt (2005) favours it as a stage of the GSS. However, the referral of the Solikamskian Horizon to the revised Kungurian (Chuvashov et al., 2002a; Kotlyar, 2002) is not supported by the evidence of temperate-water foraminifers, brachiopods, bivalves, fish remains, charophytes, and macrofloral remains. In contrast, it yields numerous taxa unknown from the Early Permian, but widespread in the Late Permian, although admittedly, the Solikamskian Horizon does not contain any warm-water fusulinids or ammonoids (Grunt, this work).

The Sheshmian Horizon (84 to 92 m) is composed of reddish-brown, brownish-red, yellowish-brown clastics, marls, and few limestones. It contains mainly non-marine fossils.

5.2.4. Donets (T.I. Nemyrovska, V.I. Davydov)

In the Donets Basin the CPB lies within a terrestrial succession of monotonous sandy–clayey red beds. However, in the Tormosin Depression (Pre-Donets

Trough), the CPB is situated in the shallow marine Kartamysh Formation (Fig. 4) at limestone Q_7 (Davydov, 1992). This position corresponds well with palynological (Inosova et al., 1976) and palaeomagnetic (Khramov and Davydov, 1984) data. The overlying Nikitovka and Slavyansk formations probably correspond to the Kurmaya Formation of the Ural Mts. In the Donets Basin Middle Asselian rocks are either missing or represented by continental facies, whereas they are complete in the Tormosin Depression and represented by clayey, sandy–clayey and carbonate deposits. Late Sakmarian, Artinskian, and Kungurian deposits are missing.

5.2.5. Tethys (mainly western part) (H. Forke, V.I. Davydov, H.W. Kozur)

The Early Permian (Cisuralian Period) is characterized by an increasing faunal provinciality. Therefore, the late Cisuralian Tethyan regional stages, Yakhtashian, Bolorian, and early Kubergandian (which possibly began in global Kungurian time), differ considerably from the East European regional stages. The stratotypes of those Tethyan stages are located in SW Darvaz (Central Asia, Tadjikistan). Their boundaries are traditionally defined by fusulinids, whereas there are only sparse conodont and ammonoid data (Kozur, 1978a; Reimers, 1991; Kozur, 1995; Forke, 1995; Buser and Forke, 1996; Forke, 2002), which do not allow detailed zonations.

In the early Early Permian the regional stage names still coincide with those of the East European stages. The Asselian is recognized by the appearance of the fusulinoid genus *Sphaeroschwagerina,* and characterized by *Pseudoschwagerina*, *Paraschwagerina*, and advanced *Dutkevitchia*. The Sakmarian is characterized by the fusulinoid genera *Zellia* and *Robustoschwagerina*. However, as these genera are absent in the Ural Mts., the correlation is based on the appearance of the Tastubian species *Schwagerina* (=*Pseudofusulina*) *moelleri* supposed to be present in both provinces. Sakmarian ammonoid genera are also common in Darvaz and the Ural Mts. (Leven et al., 1992).

The stratotype of the regional Yakhtashian Stage (Leven, 1980), a name derived from Mount Yakhtash, is in the Vozgina Section. Biostratigraphically, the base of the stage is poorly defined; thus the base of the Zygar Formation is used. It is marked roughly by a faunal change (disappearance of Asselian–Sakmarian fusulinid associations and the introduction of the newly evolved genera *Pamirina* (*Levenella*) and *Chalaroschwagerina*). The approximate correlation of the Yakhtashian with the Artinskian is based solely on sparse conodont and ammonoid data. The late Yakhtashian of the stratotype yields *Sweetognathus inornatus* and *M. bisselli*, which correlate with the early Artinskian (Kozur et al., 1994). However, other sections in the western and central Tethyan Realm indicate a late Artinskian age for the Yakhtashian (Forke, 2002; Fig. 4: arrow up at its base).

The stratotype of the regional Bolorian Stage is at the watershed of the rivers Charymdara, Gundara, and Zidadara. Its name is derived from Bolor, a mythic mountain ridge that Alexander von Humboldt thought was present in the area of Darvas and the Pamir Mts. The base of the Bolorian Stage is defined by the appearance of the genus *Misellina* (*Brevaxina*) (Leven, 1979).

5.2.6. South China (Y.-G. Jin)

The Chuanshanian Series/Epoch (Huang, 1932) encompasses the strata with *Pseudoschwagerina*. Its older Zisongian Stage originally included the *Ps. uddeni–Ps. texana* Zone and *Sphaeroschwagerina* Genozone with its stratotype in the Yangchang Section, Ziyun County, Guizhou Province. It contains three fusulinid zones: *Sphaeroschwagerina fusiformis–Pseudoschwagerina*, *Sph. moelleri*, and *Robustoschwagerina schellwieni–R. ziyunensis*, and three conodont zones: *Streptognathodus wabaunsensis*, *St. barskovi*, and *M. bisselli*. The exclusively carbonate facies is conventionally named as Maping Formation, or Chuanshan Formation. The frequent sea level changes resulted in alternating white and black limestones, as well as alternating dolomites and limestones.

The Longlinian Stage was proposed between the FAD of *Pseudoschwagerina* and the FOD of *Misellina*. The global regression occurring during the Longlinian and Luodianian stages induced strong facies differentiation. The dominant fusulinid assemblages varies as a result of changing carbonate environments. In the shallow inner shelf they are characterized by *Nankinella, Staffella*, and *Pamirina*, in the outer shelf and slope by *Pamirina,* and in shelf depressions or local basins by *Chalaroschwagerina* and *Pseudofusulina*. Coastal deposits consist of mainly interbedded shale, siltstone, bauxitic clay beds and thin-bedded limestone. Deltaic facies, composed of sandstone containing productive coal seams are developed locally.

The Yangsingian Series/Epoch (Huang, 1932) comprises the fusulinid genozones *Misellina* and *Neoschwagerina*. It corresponds to the succession between the maximum regression occurring at the very beginning of the Chihsian Subseries and the commencement of the Lopingian transgression. The Chihsian covers a broad stratigraphic range corresponding to the *Misellina* and *Cancellina* genozones. It can be approximately correlated to the global

Kungurian Stage and the regional Cathedralian Stage in the United States. However, the lower boundary of the latter two stages, defined by the basal *N. pnevi* Zone or *N. exsculptus* Zone, are within the Longlinian Stage.

Sheng and Jin (1994) proposed to introduce the Luodianian Stage for the outer shelf corresponding to the stratigraphic range of the genus *Misellina* which can usually be subdivided into the *Brevaxina dyhrenfurthi*, *Misellina claudiae*, and *Shengella* zones. The conodonts of this stage are grouped in the *Mesogondolella gujioensis–M. intermedia* Zone. In South China the ammonoids from the late Luodianian were assigned to the *Pseudohalorites* fauna, and those from the beds of the *M. claudiae* Zone include species of *Propopanoceras, Metaperrinites, Eothinites, Neocrimites, Parapronorites, Marathonites*, and *Agathiceras*.

The Xiangboan Stage was proposed for strata containing the *Cancellina* Genozone (Fan et al., 1990), which are best developed in the Houchang Section, Ziyun County, Guizhou Province. The base is distinguished by the appearance of the primitive forms of *Cancellina*. Three fusulinid range subzones are recognized at this section: *Cancellina elliptica*, *C. liuzhiensis,* and *Neoschwagerina simplex*.

Early Chihsian deposits consist of argillaceous carbonates on the inner shelf, whereas those in the outer shelf and basin resemble the Longlinian facies. The late Chihsian shows strong facies differentiation, with alternating shales and thin-bedded limestones dominant in depressions, and nodular limestones with a characteristic brachiopod fauna of *Cryptospirifer* and fusulinids (e.g. *N. simplex* and *Verbeekina grabaui*) on the highs. Dolomite and reef facies dominate towards the outer shelf, and cherty limestones are present in local basins.

5.2.7. North America (V.I. Davydov, H.W. Kozur)

The traditional reference sections of the North American Permian are in West Texas (Delaware Basin) and the Midcontinent (mainly Kansas and Oklahoma). The original Early Permian is composed of the Wolfcampian and Leonardian Series (cf. Dunbar et al., 1960). In the stratotype sections in the Glass Mountains, West Texas, they can be subdivided into the regional Nealian, Lenoxian, Hessian and Cathedralian stages. These are homonymous, but not time-equivalent, with formations from which their names were derived. The original Wolfcampian starts in the late Gzhelian, the original Leonardian (Glass Mts.) ends within the late Kungurian (Fig. 4).

The base of the Wolfcampian has been modified several times, moreover, it is placed at different stratigraphic levels in the Glass Mountains and the Midcontinent. The fusulinid assemblages of both areas are quite provincial, although some rare taxa allow a correlation with the Tethys.

In the Glass Mountains conodonts from the earliest Nealian of the type section (section 22 of King, 1931) indicate a latest Asselian age. The index species of the Sakmarian, *Sw. merrilli* (cf. 5.1.1) was found 52 m above the base of the Nealian type section. The base of the global Permian is within the Gray Limestone Member of the Gaptank Formation (Wardlaw and Davydov, 2000).

In the Midcontinent the traditional base of the Wolfcampian is in the Admire Group (Thompson, 1954) which since the introduction of the GSSP concept, would be of late Gzhelian age. Later, the boundary was proposed at the Neva Limestone (Baars et al., 1994) of Middle Asselian age. Recently, the CPB is precisely placed at the base of the Glenrock Limestone of the Red Eagle Cyclothem (Fig. 4) within the conodont succession *S. wabaunsensis–S. isolatus* at the FAD of the latter species (Chernykh et al., 1997; Boardman et al., 1998).

Ross and Ross (1987) proposed the Bursumian as the new terminal stage of the Pennsylvanian. However, for several reasons (Davydov, 2001) this usage was never adopted. The Bursumian covers a late Gzhelian to middle Asselian time span. In its type section of Lenox Hill in the Glass Mts. the Lenoxian Stage contains typical Artinskian conodonts as *Sw. whitei* and *Neostreptognathodus transitus* and the Yakhtashian fusulinid *Chalaroschwagerina* (Wardlaw and Davydov, 2000; Mei et al., 2002).

In its type section the Hessian Stage contains at its base *N. pequopensis*. At 17 m above the base, a plethora of species of *Mesogondolella* and *Neostreptognathodus* occur, including *N. exsculptus* which approximately marks the base of the global Kungurian. In the Pequope Mts. the later taxon occurs together with *N. pnevi* (Wardlaw et al., 1998), the proposed index-species for the base of the global Kungurian (cf. Section 5.2.1). Thus, the regional Hessian Stage starts in the latest Artinskian. The Hessian and Cathedralian stages as well as the Hess Member, the Cathedral Mountain Formation and the lower Road Canyon Formation correlate approximately with the global Kungurian (Fig. 4). In the Midcontinent the FAD of the conodont *Sw. merrilli* places the base of the Sakmarian quite precisely within the Eiss Limestone of the Council Grove Group and the FAD of the conodont *Sw. whitei* places the base of the Artinskian within the Florence Limestone of the Chase Group (Boardman et al., 1998; Mei et al., 2002; Fig. 4).

5.3. Roadian, Wordian, and Capitanian stages (Guadalupian, Middle Permian Epoch)

5.3.1. GSSPs and numerical ages (M. Menning, H.W. Kozur)

The three GSSPs are located in the Guadalupe Mts., West Texas (for details: Section 5.3.6); the stages are defined by conodonts (Glenister et al., 1999). The term Roadian is derived from the Road Canyon Formation of the Glass Mts., the former lower Word Formation (for details cf. DCP 2003). The base of the Roadian Stage is defined in the El Centro Member, middle Cutoff Formation, by the FAD of *Jinogondolella nankingensis*. In the STD 2002/DCP 2003 its numerical age is taken at ca. 272.5 Ma (Menning and Deutsche Stratigraphische Kommission, 2002); it may be a little younger, but certainly older than 270 Ma in order to avoid an overstretching of the Early Permian. In contrast, other age estimates for this boundary are 269 Ma (Kozur, 2003) and 270.6 ± 0.7 Ma (Gradstein et al., 2004). The base of the Wordian Stage is defined by the FAD of *Jinogondolella aserrata* within the Getaway Limestone, lower Cherry Canyon Formation. The base of the Capitanian Stage is defined with the FAD of *Jinogondolella postserrata* in the Pinery Limestone, lower Bell Canyon Formation at the Nipple Hill. Its age is ca. 265.0 Ma, derived from the U–Pb ID-TIMS age of 265.3 ± 0.2 Ma (Bowring et al., 1998) of a tuff between the Hegler Limestone (below) and the Pinery Limestone. It confirms the age of ≤ 265 Ma which was derived using geological time indications (Menning, 1995a,b). To date, there are neither isotopic, nor cyclostratigraphic data on which to estimate the numerical age of the Roadian–Wordian boundary.

5.3.2. Central and West Europe (M. Menning)

The stratigraphic positions of the bases of the Roadian and Wordian stages within the terrestrial late Rotliegend are unknown. In the Central European Basin both boundaries are in a large gap between the Müritz Subgroup and the Havel Subgroup (Menning, 1995a,b; STD 2002: ca. 282 Ma to 266 Ma; Fig. 4). The position of the Illawarra Reversal has been recognized in the Parchim Formation (Menning et al., 1988) and in the latest Wordian of the Guadalupe Mts. (Menning et al., 2005), and thus, the base of the Capitanian Stage lies within that formation. The Rotliegend of the intra-Variscan basins is almost exclusively reversed magnetized. Therefore, it has a pre-Illawarra, pre-Capitanian age of ≥ 265 Ma (Menning et al., 1988, 2005) excluding the latest Rotliegend Eisleben, Cornberg, and Battenberg formations (STD 2002).

The four formations of the Havel and Elbe subgroups are nearly isochronous time slices which have been termed "Folgen". The Dethlingen Folge and the Hannover Folge consists on 7 cycles each. A cycle has a duration of ca. 0.4 my. Thus, the Elbe Subgroup has a duration of ca. 5.6 my (Gast and Menning, in Menning et al., 2005; Fig. 4: 4 my). Only the latest cycle, directly below the Zechstein, shows marine influence. The fossils of the late Rotliegend are indicators of facies, rather than of time (Schneider et al., 1995).

5.3.3. East Europe (T.A. Grunt, G.V. Kotlyar, M. Menning)

The Late Permian of East Europe consisted for 40 years of the three regional stages Ufimian, Kazanian, and Tatarian (Resolutions, 1965), but these names were also used for stages of the GSS (Fig. 1). Their stratotype and reference sections are described in detail in a monograph by Esaulova et al. (1998). The introduction of the GSSP concept resulted in the controversial subdivision of the Late Permian into the Middle and Late Permian, the substitution of the East European stages by North American and Chinese ones including new defined stage boundaries. In order to improve the equivalence with the new Middle and Late Permian GSS (cf. Sections 5.1 and 5.3.1), the All-Russian Conference (2002, 2004) and Grunt (2005) introduced the term "Biarmian" in a redefinition, and reclassification of East European units (Fig. 4).

Kazanian deposits are wide-spread in Povolzhe and Prikame. Marine clays, marls, and limestones interfinger with terrestrial red-beds (muddy sandstones). On one hand, the Kazanian Stage is subdivided into the Sokian Substage (type locality: Sok River Basin, Baitugan Village; Baitugan, Kamyshla, and Barbashi formations; Forsh, 1955) and the Povolzhian Substage (type locality: Volga river, near the Pechishchi village; Esaulova et al., 1998). On the other hand, the Kazanian Stage is subdivided into the Kamian Substage (Baitugan, Kamyshla, and Krasnoyarsk formations, total thickness 75–85 m, and the Krasnovidovian Substage (Prikazan, Pechishchi, Verkhniya Uslonka, and Morkvashino formations, total thickness 48–50 m; Grunt, 2005) (Fig. 4).

The base of the Kazanian Stage was defined at the base of "Spiriferida" (*Licharewia*) Beds. Recently, the base of *Licharewia schrencki* Beds in the Cheshskaya Bay Section of the Kanin Peninsula (Grunt et al., 2002) has been proposed as the lower boundary stratotype. The conodont *Kamagnathus (=Stepanovites) khalim-*

badzhae from the Baitugan Formation was discovered together with the Roadian index conodont *J. nankingensis* (*Mesogondolella serrata*) in the Phosphoria Formation of Utah (Chernykh et al., 2001; Chernykh and Silantiev, 2004). Moreover, the presence of the ammonoid *Sverdrupites* in the late Early Kazanian (Leonova et al., 2002) allows the approximate correlation of the early Kazanian Stage (RSS) with the early Roadian Stage (GSS).

According to Chernykh, the entire Kazanian corresponds to the *Kamagnathus* Genozone and could be correlated with the Roadian Stage and therefore, the entire Ufimian Stage is probably older than the Roadian Stage. However, the suggestion of Kotlyar (2002) that the latest Ufimian Stage correlates with the earliest Roadian Stage, cannot be excluded (Fig. 4: arrow up at the upper boundary of the Ufimian Stage).

The original Tatarian consists of the Early Tatarian Urzhumian Horizon and the Late Tatarian Severodvinian and Vyatkian horizons (Fig. 4). Recently, it was decided, to separate the Urzhumian from the Tatarian in order to restrict the latter to the former Late Tatarian, and to re-classify this reduced Tatarian as a regional Tatarian Series/Epoch corresponding to the global Capitanian, Wuchiapingian, and Changhsingian stages (Resolutions, 2006). The older, newly introduced regional Biarmian Series/Epoch, consists of the regional, Kazanian and Urzhumian stages, which correspond approximately to the global, Roadian and Wordian stages (Fig. 4).

The boundary between the newly introduced supraregional Biarmian and Tatarian epochs/series approximates to the Illawarra Reversal, which separates the Carboniferous–Permian (Kiaman) Reversed polarized Superchrone/Megazone from the Permian–Triassic (Illawarra) Mixed polarized Superchrone/Megazone. It is the most pronounced magnetostratigraphic marker in the Palaeozoic Era.

The Urzhumian Horizon (Fredericks, 1918) is now reclassified as regional stage (Resolutions, 2006) which corresponds approximately to the global Wordian Stage. The upper boundaries of the Wordian and Urzhumian stages correlate exactly when using the Illawarra Reversal, which is positioned in the latest Urzhumian (Khramov, 1963) and the latest Wordian (Glenister et al., 1999; Burov et al., 2002; Menning et al., 2005). The Urzhumian is composed of interbedded sandstone-shale and rhythmic mudstone and limestone, and coincides with the *Platysomus biarmicus–Kargalichthys efremovi* ichthyological Zone. Its lower boundary is marked by the base of the *Palaeodarwinula fragiliformis–Prasuchonella nasalis* ostracod Zone.

According to the Resolutions (2006), the new supraregional Tatarian Period/Series consists of the Severodvinian and Vyatkian stages, which were introduced by Ignatiev (1962) as horizons. The Severodvinian Horizon (Fig. 4) is composed of mainly alluvial–limnic sandstone–siltstone and variegated marly limestone. The basal boundary of its Early Substage is defined at the base of the *Suchonellina inornata–P. nasalis* ostracod Zone. Its Late Substage is fixed at the base of the *S. inornata–Prasuchonella stelmachovi* ostracod Zone, coinciding with the *Deltavjatia vjatkensis* tetrapod Zone.

The Vyatkian Horizon (Fig. 4) is composed of fluvial, alluvial plains, and limnic facies. Its base is fixed at the bases of the *Wjatkellina fragilis–Dvinella cyrta* ostracod Zone, the *Toyemia blumentalis–Strelnia certa* ichthyological Zone, and the magnetostratigraphic Zone N_2P.

The magnetostratigraphic evidence permits a correlation of the marine parts of the Kazanian and Urzhumian successions of East Europe with parts of the entirely terrestrial Rotliegend of Central and West Europe (Fig. 4; Menning, 2001), and a correlation of the marine early Zechstein of Central and West Europe with the entirely terrestrial late Tatarian (Vyatkian) deposits of East Europe. Consequently, marine fossils cannot be used to demonstrate that the successions in Central and West Europe are approximately isochronous with those of East Europe. Thus, marine fossils like the brachiopod *Pterospirifer alatus* and the conodonts *Merrillina divergens* and *Mesogondolella britannica* were found in the early Zechstein and its equivalents in Greenland and Spitsbergen, but not in the much older Kazanian rocks in East Europe (cf. Section 5.4.2). Also the mostly facies-restricted palynomorphs provide little basis for correlation between East Europe and Central/West Europe. Thus the traditional interpretation of the Kazanian as the Russian time-equivalent of the German Zechstein is completely erroneous. The rocks of the regional Kazanian Stage in East Europe are much older than those of the Zechstein Group, because the exclusively reversed polarized Kazanian rocks belong to the Carboniferous–Permian Reversed Superchrone/Megazone (Kiaman) whereas the mixed polarized Zechstein rocks belong to the Permian–Triassic Mixed Superchrone/Megazone.

5.3.4. Tethys (mainly western part) (V.I. Davydov, G.V. Kotlyar, H.W. Kozur)

Originally, the Guadalupian and Lopingian of the Tethys were subdivided only in two fairly large stages, the Murgabian and Pamirian, defined by fusulinid genozones of Miklukho-Maclay (1958). Leven (1975) proposed a very comprehensive Middle–Late Permian Tethyan time scale with five stages. The three earlier stages are based on fusulinid phylogenies and the two Lopingian stages mainly on ammonoids and other pelagic fossils. The stratotype of the Kubergandian Stage is at the Kubergandy River, SW Pamir Mts. Originally, the stage corresponded to the Kubergandy Formation, except of the lowermost part of that formation containing the famous Kubergandian ammonoid assemblage, and its base was defined by the FAD of the genus *Cancellina* (Leven, 1963, 1975). However, Leven (1979) lowered the base at the FAD of *Armenina* and advanced species of *Misellina* below the occurrence of the Kubergandian ammonoid assemblage. The early substage of the Kubergandian corresponds to the *M. claudiae–Armenina* fusulinid Zone and the late substage corresponds to the *Cancellina curtalensis* fusulinid Zone. Both substages are correlated Tethys-wide. In the type-section the ammonoids of the early Kubergandian are considered as of Roadian age (Chediya et al., 1986). A similar assemblage is found in the late Kubergandian of Afghanistan (Termier and Termier, 1970). Kubergandian conodonts from the Pamir Mts. and Transcaucasia are poor and lack evidence for a Roadian age. At present, conodont data from the Luodian section of South China cannot be used for reliable correlations (Kozur et al., 2001b; Mei and Henderson, 2002).

The regional Murgabian Stage (Dutkevich and Khabakov, 1934) corresponds to the *Neoschwagerina*, *Yabeina*, and *Lepidolina* genozones. Its stratotype is at the Mount Dhzamaltau at the Murgab River in the SW Pamir Mts. The lower boundary is defined by the FAD of *Neoschwagerina* and *Praesumatrina*. Its three fusulinid zones were originally established in the Akasaka Limestone of Japan (Ozawa, 1925; Honjo, 1959) (upwards): *N. simplex*; *Neoschwagerina deprati* (=former *N. "craticulifera"* — cf. Leven, 1993 for details) and *Neoschwagerina margaritae*. Leven (1993) excluded the latter zone from the Murgabian Stage and proposed a more detailed fusulinid zonation on the basis of sumatrinid fusulinids, but it has yet to be adopted. Murgabian rocks are poor in conodonts and/or ammonoids (*Mesogondolella siciliensis* and *Waagenoceras* faunas respectively). Thus, the Murgabian Stage is only provisionally correlated with the GSS. Usually the Murgabian is correlated with the entire Wordian; however, late Wordian conodonts and fusulinids were found in the regional early Midian Stage (Kozur and Davydov, 1996; Kozur et al., 2001a) and therefore, the Murgabian–Midian boundary is in the late Wordian (Fig. 4: ~266 Ma).

The Arpa and Khachik formations of Transcaucasia were introduced as equivalent of the latest Guadalupian stage of the Tethyan time scale as Midian Stage to fill the time span between the Murgabian (Pamir Mts,) and Dzhulfian (Transcaucasia) stages (Leven, 1980, cf. Kotlyar et al., 1983b). Initially, the Midian was thought to correspond to the *Yabeina–Lepidolina* fusulinid Genozone. The early Midian overlaps with the late Murgabian *N. margaritae* fusulinid Zone. As mentioned above, the latter zone has recently been excluded from the Murgabian Stage and has been transferred to the Midian Stage. The Midian is divided into two or three fusulinid zones: *Yabeina archaica* (or *Yabeina ozawai–Lepidolina igoi*); *Yabeina globosa–Lepidolina multiseptata*; *Lepidolina kumaensis* (Ozawa, 1970; Davydov, 1995; Leven, 1996), however, the definition of these zones, boundaries, and their relationship is not well constrained. It cannot be excluded that the two later zones partly overlap each other and that the uppermost one is of Dzhulfian age (Ozawa, 1970; Davydov et al., 1991; Leven, 1996). Ammonoids and conodonts provide a correlation of the Midian Stage with the GSS. The classic ammonoid locality Rupe del Passo di Burgio in Sicily contains late Wordian ammonoids known in the Guadalupian Mts., Midian fusulinids and a deep-water *M. siciliensis* conodont fauna (Kozur and Davydov, 1996). Similar fusulinid, ammonoid, and conodont assemblages are found in Oman, where they co-occur with the index of the Wordian, *J. aserrata*. Additionally, in the Wadi Wasit section the ammonoid genus *Timorites* is present, which begins in the Guadalupian type area in the late Wordian. Therefore, the assignment of these beds and the similar fauna of Rupe del Passo di Burgio to the Kungurian by Mei and Henderson (2002) is surely incorrect. However, the definition of the base of the Roadian Stage within the Tethyan Scale is complex and under discussion.

5.3.5. South China (Y.-G. Jin)

The Maokouan Subseries conforms to a fusulinid succession combining the *Praesumatrina*, *Neoschwagerina,* and *Yabeina* genozones. It may cover approximately the same time span as the Guadalupian Series because *J. nankingensis,* the index species of the Roadian base (Fig. 4), occurs in the Luodian sections in association with the fusulinids of the *Neoschwagerina craticulifera* Zone, but may extend downward into the *N. simplex–Praesumatrina neoschwagerinoides* Zone. In Tethyan areas the base of the Guadalupian has not been finally determined (Jin et al., 1998b).

The Early Maokouan Kuhfengian Stage was recommended for the stratigraphic succession from the base of the *J. nankingensis* Zone to that of the *J. postserrata* Zone (Sheng and Jin, 1994), with the Houchang section as the stratotype. The Kuhfengian fusulinids are grouped into the *N. craticulifera* and the *N. magaritacea* zones. At the Sazhi section in western Guizhou, where the Maokou Formation was named, the fusulinids of these zones are dominated by *Afghanella schenki*; the ammonoids were referred to the *Waagenoceras* Zone, the *Kufengoceras* Zone or the *Altudoceras–Paraceltites* Zone.

In the Penglaitan section, the (supra)regional stratotype, at Laibin, Guangxi, the Late Maokouan Lengwuan Stage contains the conodont zones *Clarkina postserrata*, *Jinogondolella altudaensis*, *J. prexuanhanensis*, *J. xuanhanensis*, *J. granti*, and *Clarkina postbitteri hongshuiensis*. The base of the Lengwuan Stage is placed at the base of the *J. postserrata* Zone, which is approximately the same level as the first appearance of the genus *Yabeina* (Sheng and Jin, 1994). In eastern parts of South China, Lengwuan deposits are characterized by the fusulinids *Eopolydiexodina*, *Metadoliolina*, *Codonofusiella*, *Minojapanella*, *Reichelina* as well as abundant ammonoids including species of *Shouchangoceras*, *Santzites*, *Altudoceras*, *Cibolites*, and *Paraceltites*. In the USA *C. postbitteri hongshuiensis* is in the topmost regional Capitanian Stage. Therefore, the latter is well correlated with the topmost Lengwuan Stage.

In eastern South China the facies pattern distribution changed remarkably at the beginning of the Maokouan, when the pre-deltaic and basinal deposits of the Kuhfeng Formation and related formations gradually shift upwards into deltaic facies or patch-reef facies.

5.3.6. North America (M. Menning, H.W. Kozur)

The original Guadalupian Series comprises the Cherry Canyon and Bell Canyon formations (Dunbar et al., 1960). It corresponds roughly to the recently defined global Guadalupian Series/Epoch.

Nevertheless, it is important to observe the need for caution to avoid confusion caused by subsequent interpretation of the "original" terminology used in the correlation chart of Dunbar et al. (1960) (cf. Fig. 4) for three reasons: a) the introduction of new lithostratigraphic units has led, in three cases, to a reduction of original ones, b) the revised correlation between the Guadalupe Mts. and Glass Mts. (Harris, 2000), and c) the homonymous regional and global terms which have arisen from the introduction of the GSSP concept.

With respect to a): The original Bone Spring Formation is now subdivided into the new Bone Spring Formation and the Cutoff Formation. The original Word Formation is now subdivided into the Road Canyon Formation (First Limestone of King, 1931), the new Word Formation, and the Vidrio Formation (Harris, 2000). The original Cherry Canyon Formation is now subdivided into the Brushy Canyon Formation and the new Cherry Canyon Formation.

With respect to b): The revised correlation (Harris, 2000), opens a gap of ca. 1.5 my duration between the top of the original Leonardian Series (top of the Cathedral Mountain Formation, ca. 273 Ma, Glass Mts.), and the base of the original Guadalupian Series (base Brushy Canyon Formation, original Cherry Canyon Formation, ca. 271.5 Ma, Guadalupe Mts.) because the original Word Formation (Glass Mts.) does not start with the Cherry Canyon Formation (Guadalupe Mts.) as shown by Dunbar et al. (1960), but about 1.5 my earlier (this work, Fig. 4; cf. Dunbar et al., 1960: chart-columns 69, 68).

With respect to c): The three GSSPs for the global Middle Permian were ratified in 2001. Their stratotypes are in the Guadalupe Mts. (Delaware Basin) of West Texas (Ogg, 2004; cf. 5.3.1, Fig. 4). The base of the Roadian Stage is defined in the upper part of the original Bone Spring Formation (now the Cutoff Formation). In the Glass Mts. this level is within the lower part of the original Word Formation, now Road Canyon Formation, the former "First Limestone" of King (1931), which was included in the original Guadalupian Series by Dunbar et al. (1960).

The base of the Wordian Stage is in the Getaway Limestone Member of the new Cherry Canyon Formation, which correlates with parts of the original middle and upper Word Formation of the Glass Mts. The base of the Capitanian Stage is within the Pinery Limestone Member of the Bell Canyon Formation which correlates to the main part of the Capitan Limestone (Dunbar et al., 1960, chart-column: American Standard Section).

5.4. Wuchiapingian and Changhsingian stages (Lopingian, Late Permian Epoch)

The name of the Lopingian Series is derived from the Loping Coal-bearing "series" in South China, named by von Richthofen in 1884 (Jin et al., 2003). Huang (1932) later established it as a formal series to include all Permian deposits that overlie the Maokou Limestone. The Lopingian is subdivided into the Wuchiapingian and Changhsingian stages. Their names are derived from the Wuchiaping and Changhsing (Changxing) formations (Furnish and Glenister, 1970; Kanmera and Nakazawa, 1973). The Lopingian finishes at the base of the Triassic (Yin et al., 2001), which is defined by the FAD of the conodont *Hindeodus parvus*.

5.4.1. GSSPs and numerical ages (Y.-G. Jin, M. Menning)

The GSSP for the basal boundary of the Wuchiapingian Stage, and also for the Lopingian Series/Epoch is placed at the FAD of the conodont *Clarkina postbitteri postbitteri* within an evolutionary lineage from *C. postbitteri hongshuiensis* to *C. dukouensis* in Bed 6k at the Penglaitan Section, Laibin area, Guangxi Province, South China (Jin et al., 2001). The FAD of *C. postbitteri* sensu lato could also be used to approximate this boundary, as it is only 20 cm below the defining point at the base of Bed $6i_{upper}$ at the Penglaitan Section. Historically, this boundary was intended to coincide with a global regression and has been documented as a level coincident with an important mass extinction event. This event is marked by a rapid change from *Jinogondolella* (*Mesogondolella*) to *Clarkina* for offshore conodonts, and *Sweetognathus* to *Iranognathus* for nearshore shallow water conodonts, and the sudden extinction of verbeekinid fusulinids (Jin et al., 1998a). Wuchiapingian fusulinids are dominated by the fusulinids *Codonofusiella* and *Reichelina*.

The stage is composed of the Laibinian and Laoshanian substages bounded by the base of the *Clarkina leveni* Zone. This level is approximately correlated with the base of the *Anderssonoceras–Prototoceras* ammonoid Zone and coincides with the surface of the Lopingian transgression. The Laibinian Substage contains the *C. postbitteri postbitteri*, *C. dukouensis*, and *C. asymmetrica* conodont zones, and the Laoshanian Substage contains the *C. leveni*, *C. guangyuanensis*, and *Clarkina orientalis* conodont zones as well as the *Anderssonoceras–Prototoceras*, *Araxoceras–Konglingites*, and *Sanyangites* ammonoid zones.

The Changhsingian Stage was formally proposed as the last stage of the Palaeozoic Era with its stratotype in the Section D at Meishan, Changxing County, Zhejiang Province of China (Zhao et al., 1981) with a basal boundary between the *C. orientalis* and *Clarkina subcarinata* zones. Recently, the boundary GSSP was ratified by the IUGS at the FAD of the conodont *Clarkina wangi* within the lineage from *Clarkina longicuspidata* to *C. wangi* at a point 88 cm above the base of the Changhsing Formation in the lower part of Bed 4 (base of 4a-2). This point is just above the flooding surface of the second parasequence in the Changhsing Limestone (Jin et al., 2004). The basal part of this stage is also marked by the occurrence of advanced forms of *Palaeofusulina*, and the tapashanitid and pseudotirolitid ammonoids. The Baoqingian and the Meishanian substages have been suggested, bounded by the FOD of *Clarkina changxingensis* within Bed 10. The Changhsingian–Indusian boundary GSSP is close to the top of the Changhsing Formation within the lowermost Yinkeng Formation (Fig. 4).

The numerical age of 260.5 Ma for the Capitanian–Wuchiapingian boundary in the STD 2002/DCP 2003 is an estimation which is based on a) the 265.3 ±0.2 Ma ID-TIMS age (Bowring et al., 1998) for a bentonite bed just below the base of the Capitanian Stage in its boundary stratotype at the Nipple Hill near the Guadalupe Mts., West Texas, b) the numerous ID-TIMS ages from the sections Meishan and Shangsi around the PTB (Bowring et al., 1998; cf. Menning, 2001: Fig. 1) as well as the SHRIMP age of 251.2 ±3.4 Ma for bed 25 in the GSSP of Meishan (Claoué-Long et al., 1991) which is only slightly older than the PTB in Bed 27c (Yin et al., 2001), and c) geological time indications (cf. Section 2.1). The latter suggest that the age of 253.4±0.2 Ma from just above the base of the Changhsingian Stage (Bowring et al., 1998) may be little too young. In the DCP 2003 are allocated durations of ~4 my to the Changhsingian, of ~5.5 Ma to the Wuchiapingian, and of ~4.5 my to the Capitanian stages (Figs. 1 and 4).

When using the age of 252.6±0.2 Ma for bed 25 in Meishan (Mundil et al., 2004), an age of ~252.5 Ma has been derived for the PTB (Fig. 1; Menning et al., 2005). The allocation of ~0.4 my to each of the 25 Lopingian parasequences (17.3 in the Wuchiapingian and 7.7 in the Changhsingian) of Chen et al. (1998: Fig. 9) results in durations of ~10 my for the Lopingian Epoch, of ~6.9 my for the Wuchiapingian Stage and of ~3.1 my for the Changhsingian Stage (Menning et al., 2005; Fig. 1). Thus, the age for the Capitanian–Wuchiapingian boundary is ~261 Ma according to an age of ~251 Ma for the PTB (STD 2002; GTS 2004) or ~262.5 Ma according to an age of ~252.5 Ma for the PTB (Menning et al., 2005). The corresponding ages for the base of the Changhsingian Stage are 254.1 Ma and 255.6 Ma respectively, whereas in the STD 2002/DCP 2003 an age of ~255 Ma is suggested. According to the orbital interpretation of the 25 Lopingian parasequences of Chen et al. (1998: Fig. 9) the duration of the Wuchiapingian is slightly underestimated and that of the Changhsingian is slightly overestimated.

The ages of ~261 Ma and ~262.5 Ma for the base of the Wuchiapingian Stage (Menning et al., 2005) are consistent with a zircon age of 259.3±3 Ma from the Emeishan volcanics occurring around the Guadalupian–Lopingian boundary (Zhou et al., 2002), and approximates with the age of 260.8±0.8 Ma from the

Shangsi Section (Mundil et al., 2004). The latter age is not Early Wuchiapingian as suggested, but Late Wuchiapingian because Bed 7 contains conodonts of the *C. orientalis* Zone (Jin, this work). The age of 257.3±0.3 Ma (Mundil et al., 2004) for the Early Changhsingian is significantly older than the orbital age of 255.6 Ma derived by Menning et al. (2005), which is based on the 252.6±0.2 Ma age for bed 25 in Meishan (Mundil et al., 2004).

5.4.2. Central and West Europe (M. Menning, H.W. Kozur, J.W. Schneider)

In Central and West Europe the stratigraphic range of the succession equivalent in age to the Lopingian Epoch (STD 2002/DCP 2003: ~260.5 Ma to 251 Ma) is under discussion. Whereas sediments of that age are rarely developed in the intra-Variscan basins, a complete succession is represented in the foreland basin between Central England and East Poland. There is no evidence of a break in time of ≥0.1 my in the migrating basin centre in North Germany during the 266 Ma to 229 Ma interval (STD 2002; Menning et al., 2005), but Wardlaw et al. (2004: Fig. 16.2) show a gap between the top of the Zechstein at 259.8 Ma and the base of the Buntsandstein at 251.0 Ma.

In the Central European Basin late Rotliegend sediments consists of fluvial, aeolian, playa, and sabkha sediments, which are up to 2000 m thick (Deutsche Stratigraphische Kommission, 1995; Schröder et al., 1995). These sediments are covered by Zechstein and younger sediments (STD 2002), which are some kilometres thick and penetrated by more than 1000 wells. The mainly brownish red clastics of the Havel Subgroup include aeolian sandstones with significant resources of natural gas. The clastic–evaporitic Elbe Subgroup covers a much larger area and is better subdivided and locally correlated than the Havel Subgroup. In NW Germany it accumulated mainly in a large perennial salt lake (Gast, 1991; Legler et al., 2005) and its orbitally estimated duration is ca. 5.6 my (cf. Section 5.3.2; STD 2002/DCP 2003: ~4 Ma). This duration is consistent with the age for the Rotliegend–Zechstein boundary at ~258 Ma in the STD 2002/DCP 2003, as well as the Re–Os age of 257.3±1.7 Ma of the Kupferschiefer (Copper Shale) at the base of the Zechstein (Brauns et al., 2003).

The marine–lagoonal–terrestrial Zechstein, which is subdivided and correlated in detail using marker horizons (Richter-Bernburg, 1955), consists of fine grained clastics, carbonates, sulfates, and chlorides in the central basin and terrestrial conglomerates, sandstones, and claystones in marginal areas. As in the late Rotliegend, the cyclic sedimentation is related to climatic variations.

According to magnetostratigraphic data (Menning et al., 1988), correlations (Menning, 1986, 2001) and geological time indications, the uppermost part of the Rotliegend Group and the lower part of the Zechstein Group belong to the Wuchiapingian Stage, whereas the main part of the Zechstein Group belongs to the Changhsingian Stage (Fig. 4; STD 2002). The base of the Wuchiapingian Stage is within the upper Rotliegend Elbe Subgroup (Fig. 4; STD 2002) or Havel Subgroup (Menning et al., 2005: Plate IX).

In contrast, the latest occurrence of the conodont *Merrillina divergens* in the Early Dzhulfian (Early Wuchiapingian; Kozur, 1994) and its occurrence in the Zechsteinkalk (Bender and Stoppel, 1965), early Zechstein 1, Werra Folge, is evidence for a slightly older base for the Zechstein. The occurrence of the conodont *Mesogondolella britannica* in the earliest Zechstein of England (Kozur, 1998), the Kupferschiefer of the German North Sea (Legler et al., 2005), and the Dzhulfian of the Salt Range (Kozur, 1998) is consistent with Fig. 4.

According to bio- and chemostratigraphic evidence (Kozur, 1999) and magnetostratigraphic data of Li and Wang (1989) and Szurlies et al. (2003), the Zechstein–Buntsandstein boundary is within the latest Permian, ~0.1 my older than the Changhsingian–Indusian boundary (Kozur, 2003; Menning et al., 2005).

5.4.3. East Europe (see Section 5.3.3.)

5.4.4. Tethys (mainly western part) (V.I. Davydov, G.V. Kotlyar. H.W. Kozur)

The two Lopingian Tethyan regional stages, the Dzhulfian and Dorashamian, are widely used in the literature, especially in the central and western Tethys. They are well defined by ammonoids, conodonts, and brachiopods in very fossiliferous pelagic deposits of Transcaucasia close to the Araxes River West of Dzhulfa (Julfa, Jolfa) and they are also well exposed in adjacent NW Iran and Central Iran. The early Dzhulfian consists of black to light grey limestones, and the late Dzhulfian and Dorashamian is composed of predominantly pinkish to red pelagic limestones, marls, and shales.

The Dzhulfian Stage (Schenck et al., 1941, based on the *Prototoceras* Horizon) has its stratotype in the Dorasham II-1 section West of Dzhulfa (Kotlyar et al., 1983a). Its range, proposed by Leven (1980) from the base of the *Araxilevis* (brachiopod) Beds up to the base of the *Phisonites triangulus* ammonoid Zone, corresponds to the range of the global Wuchiapingian Stage.

However, the position of the basal boundary of the Dzhulfian Stage has been slightly changed through time. Thus, it was placed a) at the base of the *Araxilevis* Beds (e.g. Kozur, 1977, 1978a; Leven, 1980), b) at the base of the *Araxoceras* (ammonoid) Beds (e.g. Kotlyar et al., 1983a) and c) at the base of the underlying shallow water *Codonofusiella* (foraminiferal) Beds (Kozur, 2004, 2005). The widely distributed ammonoids are not endemic as those of the Wuchiapingian of South China. Partly, they are even present in the Western Hemisphere (*Eoaraxoceras*, Spinosa et al., 1970). The conodont succession correlates well with that of the Wuchiapingian.

The Dorashamian Stage was introduced by Rostovtsev and Azaryan (1971) for latest Permian deposits which were long considered as of Triassic or Permian–Triassic age. Only the basal Dorashamian *P. triangulus* Zone was always considered to be of Permian age, and for a long time assigned to the latest Dzhulfian. The stratotype is the Dorasham II-3 section (Kotlyar et al., 1983a). It comprises the upper Akhura Formation and the basal Karabaglyar Formation (in NW Iran the Alibashi Formation and the basal Elikah Formation). In contrast to the endemic index forms of the Changhsingian, the Dorashamian ammonoids have a wide regional distribution in the Tethys. The conodonts are present in all low latitude open sea deposits of the Tethys and Panthalassa. Although some important index forms are absent in the intraplatform basins of South China, most conodont zones can be correlated with the Changhsingian stratotype. The base of the Dorashamian was defined by the FAD of the ammonoid *P. triangulus*, which is not present in the Changhsingian stratotype. In order to bring it in line with the base of the global Changhsingian, Kozur (2005) proposes a slightly later base at the FAD of the conodont *Clarkina hambastensis*, which appears also at the base of the Changhsingian at Meishan (Kozur, 2004, 2005). As in the Chinese intraplatform basins, the uppermost Permian is characterized by the Boundary Clay. The Triassic starts with the FAD of the conodont *H. parvus*.

5.4.5. South China (Y.-G. Jin)

The eruption of the Emeishan Basalt was a prominent feature in South China during Late Maokouan and Early Lopingian time. This volcanic group covers a wide area (330,000 km^2) over western parts of South China. The volcanic rocks, of mostly continental origin, rest directly on Maokouan limestone. Around the Capitanian–Wuchiapingian boundary a significant regression led to the extinction of numerous species. The transgressive coal-bearing Hsuanwei Formation of early Wuchiapingian age overlies the Emeishan Basalt. Laterally it changes via the Wuchiaping Formation carbonates into the Talung Formation of the basin facies.

A significant regression occurred about the Wuchiapingian/Changhsingian boundary. Changhsingian black laminated wackestone accumulated on slopes, and thick-bedded, massive, light-coloured packstone and grainstone formed in carbonate shoals and reefs; and dolomite and dolomitized limestone dominated the back reef facies. All these carbonates are included in the Changhsing Formation, whereas the basinal deposits of the Talung Formation consist of shale and cherty beds.

5.4.6. North America (H.W. Kozur, M. Menning)

The Ochoan Series (Dunbar et al., 1960) is classified here as the Ochoa Group (Fig. 4), because it “consists largely of saline deposits, chiefly anhydrite in the Castile Formation, halite with interbeds of potassium salts in the Salado Formation, and thin dolostones in the Rustler Formation” (Dunbar et al., 1960: 1768). These rocks do not provide significant biostratigraphic and magnetostratigraphic interregional time information and, therefore, numerous arrows are placed at formation boundaries (Fig. 4). The significance of gaps above those formations which are allocated to the latest Permian, the Quartermaster (Oklahoma), Elk City (Kansas), and Dewey Lake (West Texas) formations, are open questions.

The latest horizon which can be well correlated with open sea Tethys and South Chinese intraplatform basins, the Altuda Formation, lies immediately below the hypersaline Castile Formation of the basal Ochoa Group, and is characterized by the FAD of *Clarkina postbitteri* (Wardlaw et al., 2001). This indicates that the top of the regional Capitanian Stage and the base of the global Wuchiapingian Stage are nearly contemporaneous. The stratigraphic position of the Wuchiapingian–Changhsingian boundary is unknown. Conodonts are present in dolomitic limestones of the Rustler Formation above an evaporitic complex, but they are long-ranging, compound forms.

Using magnetostratigraphic data of Peterson and Nairn (1971) and Menning et al. (1988), the Illawarra Reversal lies between the Seven Rivers and Yates formations of the Delaware Basin, as well as in the Upper Rotliegend Havel Subgroup of Germany. This suggests, that the evaporites of the Ochoa Group of North America and Zechstein Group of Central and West Europe are approximately coeval (Menning, 1986: Fig. 2; Fig. 4).

Acknowledgement

We are indebted to all authors who have already contributed to other columns in the DCP 2003, which unfortunately cannot be presented here. We thank all colleagues who have helped significantly by discussions, particularly Andreas Hendrich for the drawing of the figures, which he continuously improved at every stage, from data we provided.

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