Evolution of the southern margin of the Donbas (Ukraine) from Devonian to Early Carboniferous times

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Abstract: A Devonian–Early Carboniferous succession comprising thick clastic and carbonate sediments with interbedded volcanics was examined along the southern margin of the Donbas fold belt, Ukraine. Following initial rifting and subsidence, a continental (fluvial, lacustrine) succession was established. This first phase of synrift activity (Eifelian) was accompanied by the extrusion of basalts. In Late Givetian–early Famennian times, half-graben development was pronounced and a series of E–W-trending half-grabens were formed, along with coeval volcanic activity. Subsequent basin subsidence led to the establishment of a broad marine carbonate platform across the region (Late Famennian–Tournaisian-early Viséan). Renewed uplift led to partial exposure and karstification of the platform. This was accompanied by trachyte intrusion and extrusion. The overlying chert-rich unit was probably deposited under lacustrine conditions, although a shelf environment has also been suggested. Renewed tectonic activity along the main basin-bounding fault resulted in the synsedimentary deformation of this unit.

The NW–SE-trending Pripyat–Dniepr–Donets Basin strikes in a southeasterly direction, and over a distance of *c.*2000 km, from Belarus through Ukraine, where it connects with the deformed southern margin of the East European Platform (Karpinsky Swell) in southern Russia (Fig. 1). The basin is bounded by two Precambrian basement massifs – the Priazov Massif (eastern part of the Ukrainian Shield) to the north, and the Voronezh Massif to the south, together forming the Ukrainian craton (Fig. 1). Together, these massifs define a broad domal structure which is transected by the rift (Wilson & Lyashkevich 1996). The basin system can be subdivided from northwest to southeast into the relatively shallow Pripyat Trough, the much deeper Dniepr–Donets Basin (DDB) and the inverted Donbas fold belt (DF). Sediment thicknesses increase from *c.*2 km in the Pripyat Trough to *c.*22 km in the DF (Chekunov *et al.* 1993; Stovba *et al.* 1996). The Pripyat Trough and the DDB are important hydrocarbon provinces, while the DF contains large coalfields. The Donbas fold belt–Karpinsky Swell region forms the inverted part of the intracratonic Dniepr–Donets rift basin, and is therefore, of key importance both for constraining the anomalous subsidence history (since no subsidence model has yet explained the 22-km thick sediment succession) and formation mechanism of the linked Dniepr–Donets–Donbas–Karpinsky–Peri-Caspian sedimentary basin system and also for solving the puzzle of Late Palaeozoic palaeogeography along the southern part of the East European Craton. The Donbas fold belt (DF) is a 150-km wide region

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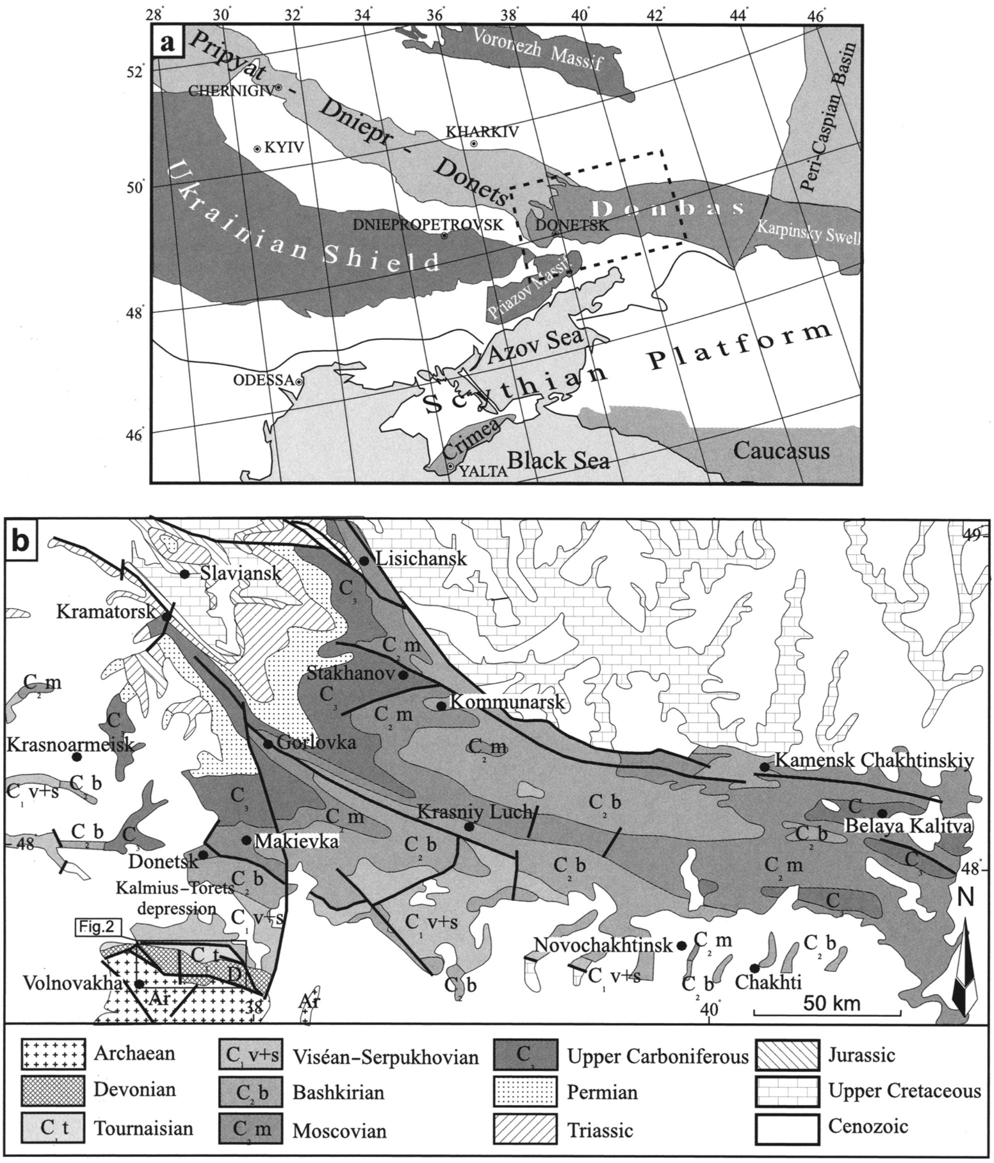


Fig. 1. (a) Location map of the Pripyat–Dniepr–Donetsk–Donbas basin system. The various structural units which compose the southern edge of the East European Platform are indicated by a selection of grey shadings. **(b)** Detailed stratigraphic map of the Donbas fold belt (after *Geological map of the USSR and adjoining water-covered areas 1983*). The location is indicated by the dashed outline in Figure 1(a).

which extends 500 km from the eastern Ukraine to southwest Russia (Fig. 1). The DF has undergone a complex tectonic evolution. Seismic evidence reveals that the Dniepr–Donets segment of the rift is characterized by linear bounding faults which penetrate much of the crust, and by significant crustal thinning (ß of approx 1.3; Kusznir *et al.* 1996) and abundant synrift volcanic activity.

The precise area of study comprises *c.* 500 km^2 and is situated about 30 km to the south of the city of Donetsk, and located between the Donbas fold belt and the crystalline Priazov Massif (Figs 1 & 2). The geology of the region is

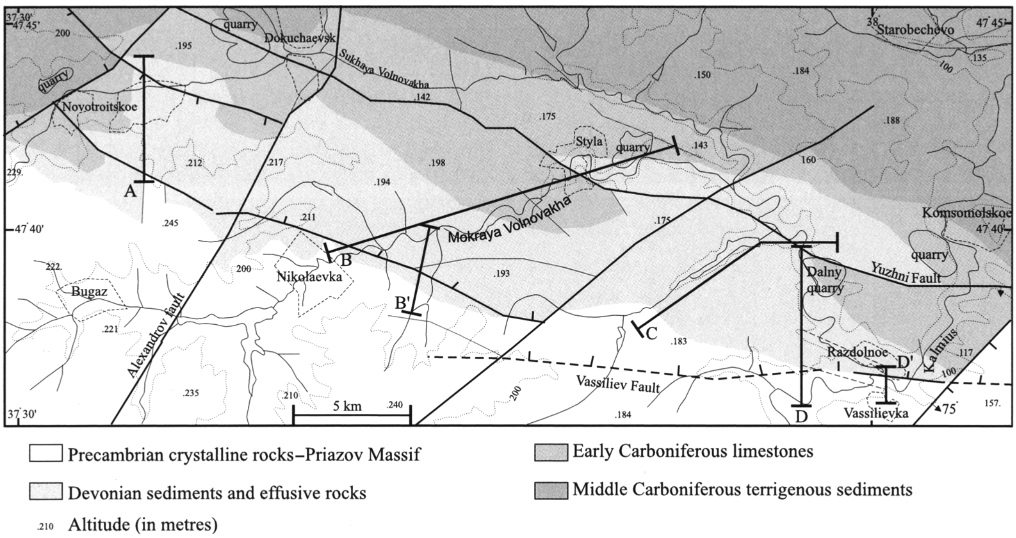


Fig. 2. Detailed map of the southern margin of the Donbas fold belt, showing the locations of the various cross-sections (after the *Geological map of the Ukrainian Donbas fold belt 1:200 000, 1995*).

characterized by a series of half grabens which developed between the Vassiliev Fault, located within the Priazov Massif, and the Yuzhni Fault, marking the southern boundary of the Styla Horst, in response to rifting phases that occurred in Devonian and Early Carboniferous times (Figs 1 & 2). The relationship between sedimentation and volcanism in the southern Donbas region, and at this particular stage in the geological evolution of the entire region, has received little attention in Western literature. Throughout much of the Donbas and Donets region, these strata are overlain by a thick cover of Upper Palaeozoic, Mesozoic and Cenozoic sediments, and thus the precise relationships between the different units are difficult to decipher. In the study area, however, exposure, while not ideal, facilitates a more detailed study of these key exposures for unravelling the early tectonic and sedimentary evolution of the Donbas region. These new data outline the sedimentological, volcanological and tectonic development of this key region from Middle Devonian to Early Carboniferous times.

Geological background

The broad geological evolution of the region is closely linked to the development of the Pripyat–Dniepr–Donets system. Prior to the onset of crustal extension, the Priazov Massif and Voronezh Massif were covered by marine Middle Devonian sediments (Alekseev *et al.* 1996; Nikishin *et al.* 1996). During the Late Devonian, rifting, which was associated with domal basement uplift and magmatism, was widespread from the Pripyat–Dniepr–Donets system in the south to the Eastern Barents Sea in the north (Wilson & Lyashkevich 1996). The evolution of this rift system is thought to be contemporaneous with the development of a major back-arc rift system in Western and Central Europe (Ziegler 1990).

Rift development in the DDB region commenced with a synrift phase in the Devonian and a second rift event in Early Carboniferous times, followed by extensional rejuvenation during the Late Carboniferous and a post-rift phase between the Permian and Quaternary, with thrust and fold development in the Late Cretaceous (Stephenson *et al.* 1993; Stovba & Stephenson 1999). The main extensional phase of the Dniepr–Donets segment occurred between the Late Frasnian (370 Ma) and the end of the Devonian (363 Ma) (Kusznir *et al.* 1996) which is when maximum rates of downwarping occurred. Rifting was accompanied by intense, probably plume-related, volcanic activity (Wilson & Lyashkevich 1996).

Sedimentology of the Southern Donbas region

Middle Devonian (Eifelian–Givetian)

The oldest exposed Devonian sediments in the region (near Styla Quarry and Nikolaevka village) rest unconformably on a Precambrian alkaline porphyritic and/or graphic granite. These sediments are of Eifelian–Givetian age (Aisenverg *et al.* 1975) (Quartz-rich sandstones overlying granite at the base of the Styla section B in Fig. 3). The top of the granite is weathered down to 10–15 m, and is overlain by a feldspar-rich, coarse-grained sandstone. The rocks are clastic, ranging from coarse gravels to micro-conglomerates to fine-grained sandstones and siltstones. In some areas the sandstones are more quartz rich, and may even resemble a quartz-pebble conglomerate. Indeed, in one particular location the rocks were quartz dominated, with mainly quartz clasts set in a quartz cement. Rare carbonates have also been noted (Figs 3 & 4).

Beds are normally graded, and up to 70 cm thick, with sharp upper and lower boundaries,

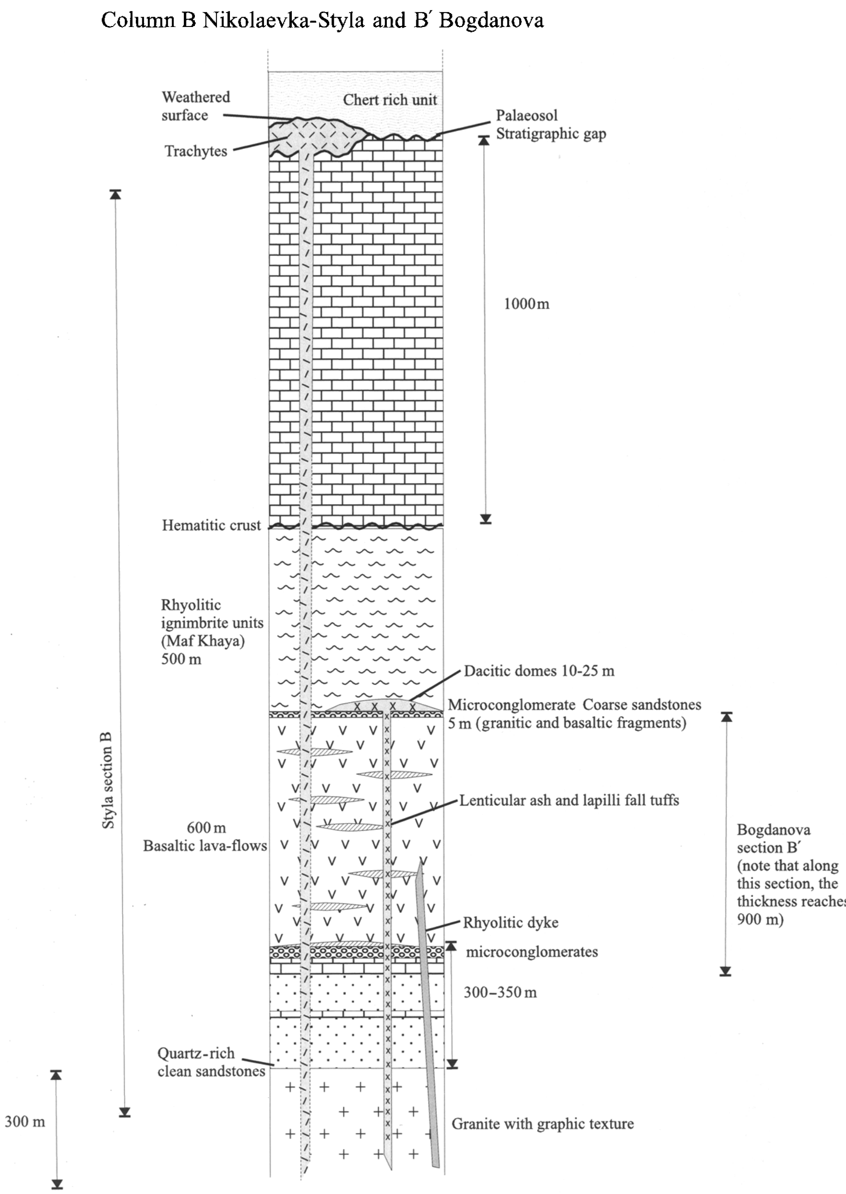


Fig. 3. Schematic stratigraphic column from the Nikolaevka–Styla (column B) and Bogdanova (column B′) areas.

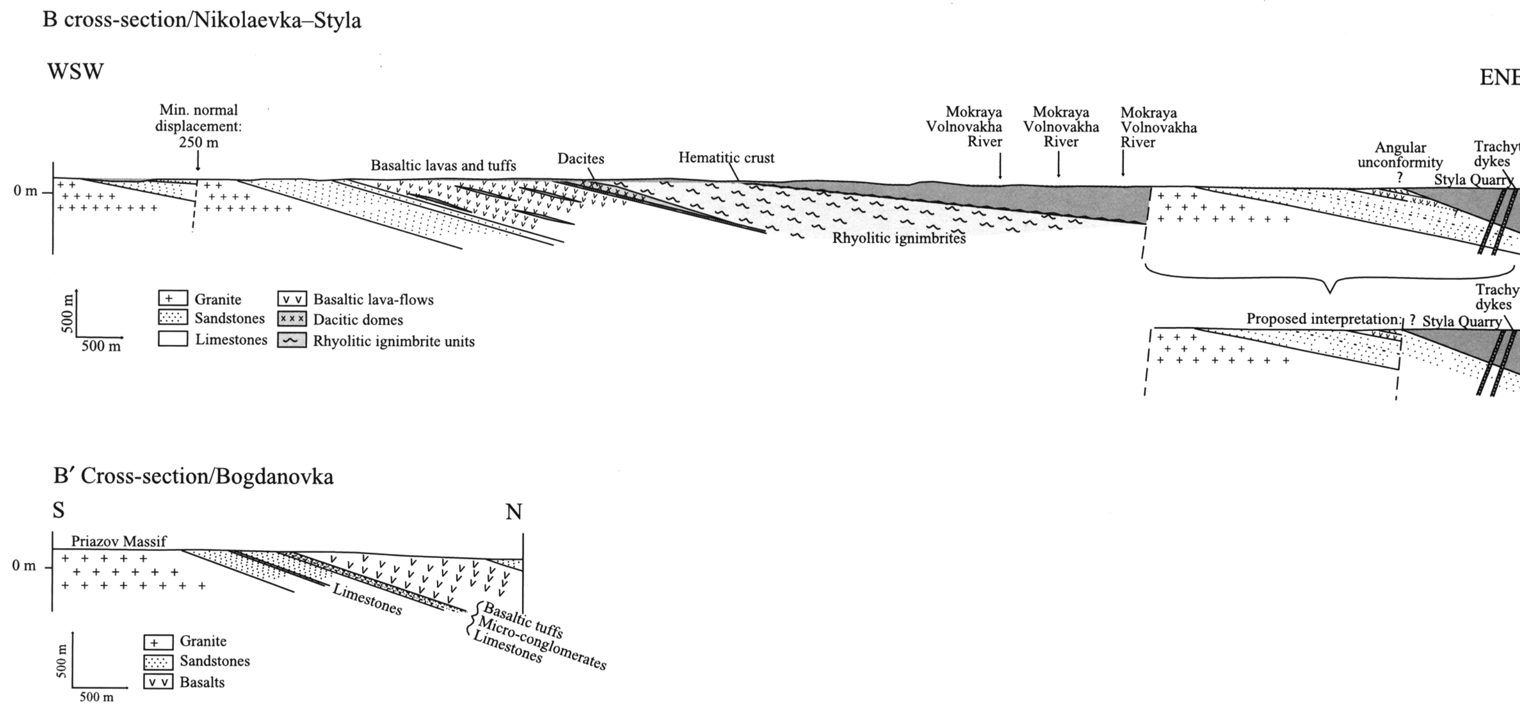


Fig. 4. Geological cross-sections from the Nikolaevka–Styla area (cross-section B) and the Bogdonovka area (cross-section B′). See Figure 2 for a precise location.

and local evidence of scouring and channelling (see below). Internally, the beds are planar and trough cross-bedded (transport direction to the north). There is evidence of local thinning and fining upwards.

Elsewhere within the study area other Middle Devonian outcrops were examined. The sediments here strongly resemble those of the Styla region, comprising mainly coarse-grained sandstones, interbedded with quartz-pebble conglomerates where the individual clasts are up to 3–4 cm in diameter. Bed thickness ranges from 0.5 to 1.0 m and internally they are cross-bedded (080°), becoming increasingly more quartzite-like up section. The beds both thin and fine upwards. Interbedded mudstones are also observed,as are sandstones with some possible tuffaceous interbeds.

As noted above, there is much evidence of channelling within the sandstones of the Middle Devonian, with individual channels cutting down into the underlying sediments. Some of these channels are associated with finer-grained (mud–fine-grained sand) overbank and interchannel deposits. Bed lenticularity may be pronounced. The channel-fill sandstones (typical orientation – 048°) are up to 3.0 m thick, normally graded and coarse grained. Individual clasts within the coarse channel fill may be up to 15.0 cm in diameter. In some locations, the sediments are intercalated with volcanic units – even showing signs of channels cutting into the underlying volcanics.

Petrographically, the samples are coarse- to medium-grained sandstones comprising monocrystalline quartz (Qm) with subsidiary polycrystalline quartz (Qp), plagioclase and a series of accessory minerals. Some samples, in addition to being rich in Qm, also have a quartz cement (sedimentary quartzites). Volcanic fragments are rare within these sandstones.

The predominance of parallel lamination and trough cross-stratification, together with the presence of channel and overbank deposits, suggest that these sediments were deposited in a continental environment, most probably fluvial or delta plain. There was, however, no evidence of any soil development.

?late Givetian/Frasnian–?early Famennian

The uppermost Devonian sediments are predominantly clastic and reddish, or sometimes light grey, in colour and include microconglomerates and coarse- to medium-grained sandstones (see also Aisenverg *et al.* 1975) (Fig. 5). Pebbles of quartz and effusive rocks are contained within the microconglomerates. Individual sandstone beds are up to 100 cm thick, have sharp boundaries, are normally graded (sometimes from a microconglomerate base) and show both trough cross-stratification and parallel lamination with some possible bed amalgamation. Some sections show evidence of coarsening upward cycles. In some areas the succession is siltstone rich, with thinner cross-bedded sandstones up to 50 cm thick. Some of these cross-bedded sandstones show evidence of channelling. The siltstones are laminated and sometimes contain coalified and silicified plant remains such as *Lepidodendron* fragments. The upper part of the succession has a hematitic crust, several metres in thickness (Fig. 5)

Petrographically, the microconglomerates contain a variety of lithic fragments, including: polycrystalline quartz (Qp) in an altered matrix; sediment lithic fragments (mainly siltstone and mudstone); volcanic fragments (mainly basic, but also acidic and ?pyroclastics), some of which are altered. Grains are contained within a matrix, which may also show evidence of alteration. Associated sandstones may be rich in monocrystalline quartz (Qm), subordinate Qp, with plagioclase also present (plagioclase percentages can be up to 15–20% in some samples). These sandstones also contain rare altered volcanic fragments. In some areas the sandstones contain so much Qm that they are almost quartzite-like.

The sediments of the Upper Devonian were probably deposited in a similar setting to those of the Middle Devonian. The evidence of trough cross-bedding, channel and interchannel deposits and the rare plant remains suggest deposition in a fluvial or delta-plain setting.

?Uppermost Famennian–Tournaisian

At the transition from the ?Uppermost Famennian to the Carboniferous there is a basal horizon (conglomerates, sandstones, limestones) 1.5–3.0 m thick. This horizon is overlain by limestones and micritic shales up to 6.0 m thick and is then replaced by massively bedded, planar limestones of the Tournaisian (upper part of Styla Section B, see Fig. 3). The limestones show marked bed continuity, with individual beds being up to 1.90 m thick.

The overlying Tournaisian succession comprises flat bedded, occasionally knobbly and lenticular, laterally continuous micritic carbonates. Bed thickness ranges from *c.*0.6 to 2.0 m, with the strata arranged in thinning-upward sequences, although there does not appear to be any formal pattern to these sequences. Rare, disarticulated bioclastic material (coral, shelly fragments) is present. The limestones appear to

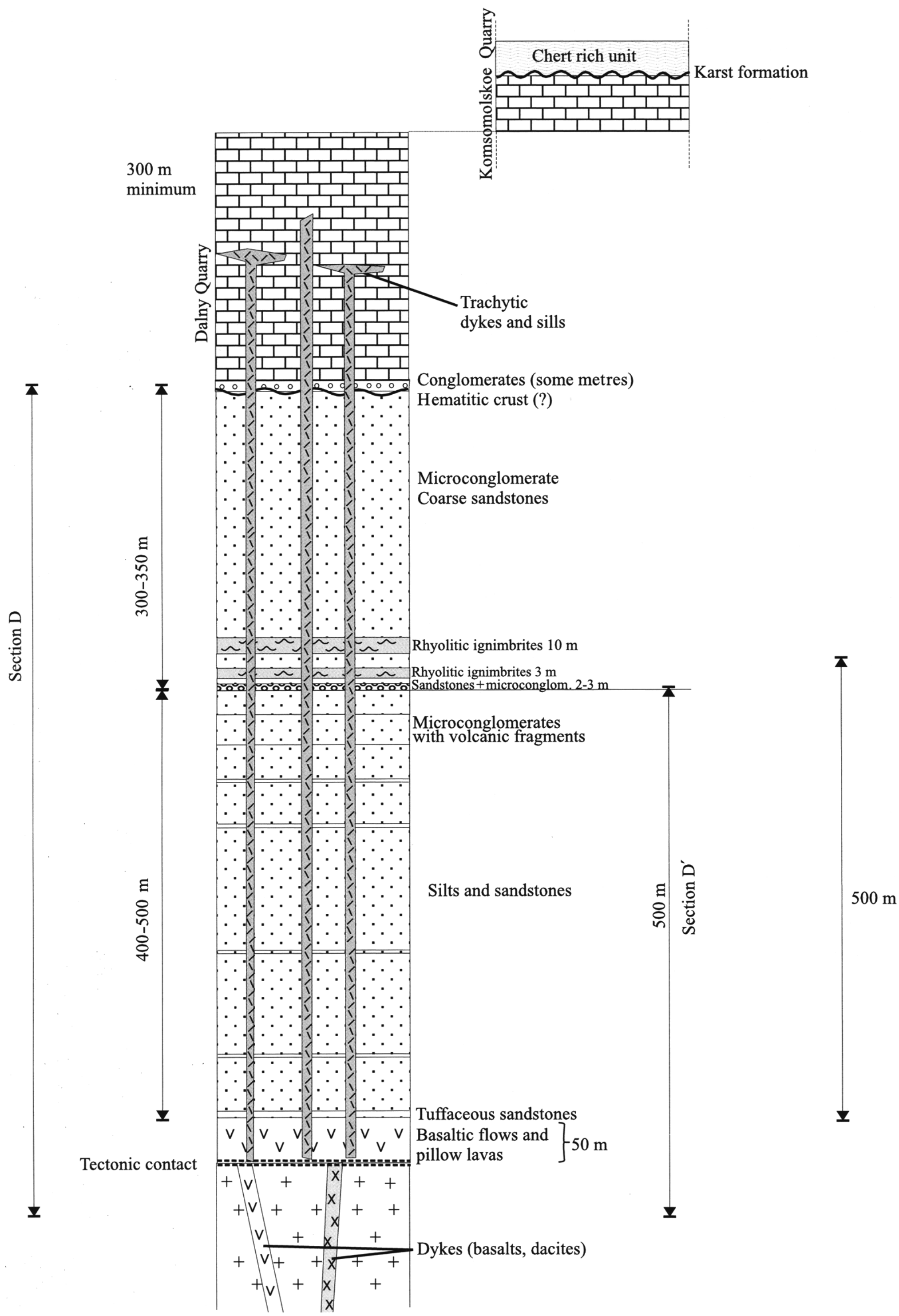


Fig. 5. Schematic stratigraphic column from the Razdolnoe–Dalniy/Vassilievka–Razdolnoe area (column D–D′).

be normally graded, although there is little variation in grain size. Some beds show planar lamination, but there is no evidence of any wave-related structures. Stylolitic layers are often present. Rare, thin mudstones may be intercalated. In some areas, the succession passes upwards into mudstones and fine–medium-grained siltstone.

Petrographically, the limestones of the Tournaisian are mainly pelsparites, with rarer biosparites which include possible ostracods and rare foraminifera. In some areas bioclastic material is more common, and here ooliths and crinoidal material may also be present. Ooliths are generally rare, although they appear to be more common at the base of the succession. The most noticeable feature of these limestones is the lack of any terrigeneous input. The dominance of peloids would suggest a shallow-marine low-energy environment. The presence of shell hash rather than intact bioclastic material suggests periods of increased energy (storms) on the shelf area. In addition, the great lateral continuity of the limestone units (>1 km), together with the lack of any obvious relief would suggest that these carbonate sequences accumulated in a low-energy shelf–platform environment. Carbonate platforms range in size from a few to hundreds of kilometres across, and the larger platforms may have built sediment piles hundreds of metres thick. The platform margin was probably rimmed, since there is some localized evidence of small reefal mounds. This may have restricted circulation with the open ocean, thus sheltering the main platform area.

Viséan

The Viséan succession can be subdivided into two units – early and late Early Viséan in age (see also Poltaev *et al.* 1988) (upper part of Styla section B of Fig. 3). The lowermost Viséan units comprise limestones and are very similar to the underlying Tournaisian strata. Beds are flat-lying and laterally continuous, although bed lenticularity is also noted. Units show evidence of upward-thinning with no sign of any internal structure within the individual beds. There is some rare evidence of parallel lamination. Furthermore, the occurrence of discrete thin, discontinuous layers of bioclastic material within some beds would suggest that bed amalgamation occurred. Beds generally range up to 80.0–150.0 cm in thickness, although in some areas bed thickness is much less (10.0–80.0 cm). Some of the thicker bed bases show evidence of sediment loading. Stylolites are also present.

Bioclastic material (e.g. solitary corals) tends to be much rarer than in the Tournaisian succession. However, in some areas the limestones have a much richer fauna, mainly comprising brachiopods, corals (both colonial and single tetracorals). Some facies successions were also observed. These commence with limestone at the base. These limestones are almost pure micrite, or rarely, sparse biomicrites. The bioclastic material mainly comprises isolated microfossils. These then pass upwards into limestones which are more fossiliferous, contain patchy sparry calcite and have an almost hash-like appearance. Some of these samples contain mixtures of peloids and fossils in a sparry calcite (pel–bio-sparite). In the upper part of the section the limestones are once again micrite rich. The entire unit probably reflects the development of a sequence of varying energy levels.

Petrographically, the limestones are mainly micrites to sparse biomicrites, or occasionally pel–bio-sparites to biosparites. The bioclastic fraction comprises microfossils, with other bioclastic material being in the form of a shell hash. The presence of rare monocrystalline quartz is indicative of proximity to a terrigenous source. Some micritic envelope structures are also present.

The depositional environment for this lowermost part of the Viséan is similar to that of the underlying Tournaisian. The decrease in bioclastic material and the relatively high percentage of micritic material may be indicative of deposition within a quieter-water environment, probably in a sheltered area of the platform (inner platform), although the area was also subjected to rare terrigenous input. The uppermost parts of this succession record evidence of subaerial karstification, suggesting emergence and weathering (Fig. 3).

In some places these Viséan karstified limestones are overlain by a faulted, folded and slumped siliciclastic succession which is late Early Viséan in age (uppermost part of column B in Fig. 3). The sediments here comprise thin shales and mudstones which are interbedded with silts and fine-grained sandstones. Some samples are micritic with rare microfossils. These latter have been infilled with chalcedony, which also replaces some isolated and fragmentary shell fragments. However, the majority of the sediments are siliceous, where the individual beds comprise a dark fine-grained silica-rich groundmass with some microfossils which have been chertified. In parts the sample resembles a very fine-grained polycrystalline quartz (i.e. chert). Cavities formed within the karstified limestone were infilled with deposits of the overlying chert-rich unit (see below).

This chert layer is microcrystalline in character and comprises mainly sponge spicules. There are also some concentric fragments that may be ooids. Other samples comprise pure chert without any internal structure or fossils. In some locations, these chert-rich units overlie the weathered tops of some trachytic domes, although other trachytic domes – also with weathered tops – are overlain elsewhere within the study area by carbonate deposits (Fig. 6). In these latter areas the chert units are missing, while the limestones are mainly biosparite (microspar) with some biomicrite. The limestones contain up to 15% quartz grains. The precise depositional environment for these chert-rich units is unclear. The lack of any carbonate material would suggest possible deposition below the carbonate compensation depth. Such a rapid change in depth (i.e. from subaerial karstification to deep marine), however, is difficult to explain. A possible water depth of *c.* 500 m has been suggested (A. Alekseev, pers. comm.) which corresponds with the calculated subsidence rate (also ± 500 m) for the region in Early Viséan times (Van Wees *et al.* 1996). Another possible explanation, however, is that the cherts were deposited within a lacustrine environment. Lacustrine silica-rich deposits, where the silica is derived from diatoms and sponges, have been reported from a number of areas (e.g. Owen & Crossley 1992). Lakes fed by silica-rich groundwaters are especially favourable sites for the

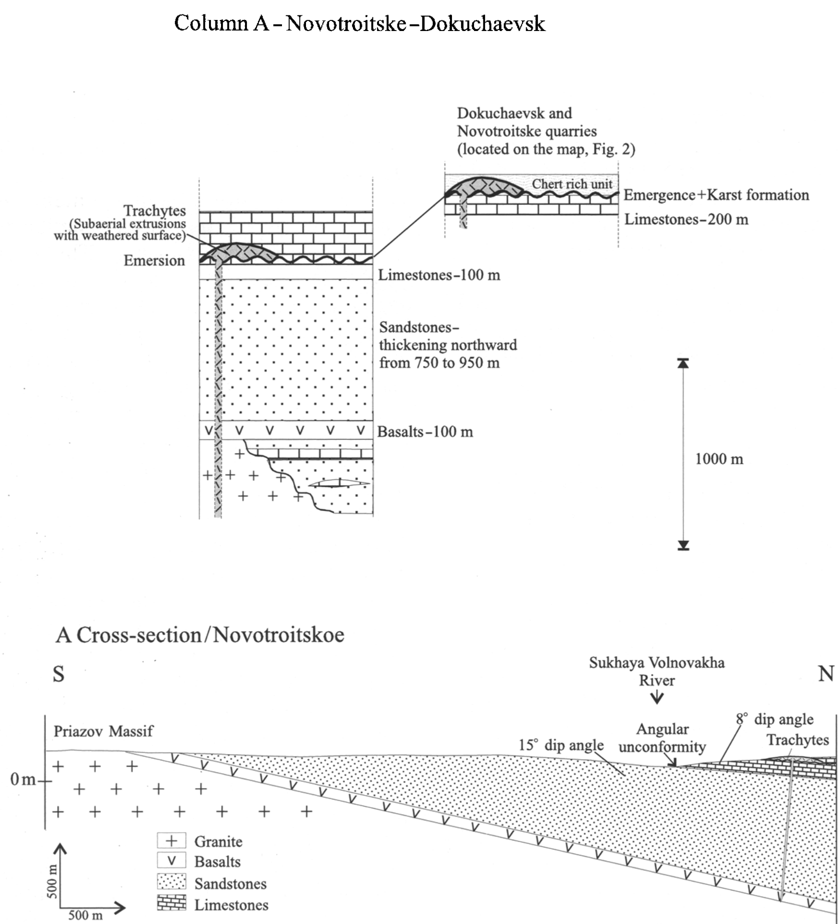


Fig. 6. Stratigraphic column (column A) from the Novotroitske–Dokuchaevsk area, and cross-section from the Novotroitskoe area (cross-section A). See Figure 2 for a precise location.

accumulation of such sediments (Talbot & Allen 1996). Of particular interest in the present situation is the observation that inlakes associated with faults and active hydrothermal systems, alkaline hot springs can be an additional source of siliceous deposits (e.g. Lake Bogoria, Kenya; Renaut & Owen 1988). The lack of regional exposure of these chert-rich units makes it difficult to determine the precise nature of the depositional environment, although a lacustrine setting would be favoured by the predominance of sponge spicules, in association with possible ooids. The extensive volcanic activity, with associated hydrothermal activity, would also have provided suitable conditions for the development of lacustrine silica-rich units.

Volcanism during the initial stages of evolution of the Donbas Basin

Along the southern part of the Donbas fold belt, volcanic rocks are interstratified with Middle Devonian–Lower Viséan clastic and carbonate sedimentary rocks (Fig. 4). The volcanics directly overlie the Proterozoic basement of the Priazov Massif in the eastern part of the studied region. Two eruptive cycles of varying importance are identified: the older one, which is ?late Middle Devonian–Upper Devonian in age, is rather diverse in terms of composition and is well developed at a regional scale; the younger, which is intra-Lower Viséan in age, is less extensive, more homogeneous in composition, and probably much shorter in duration. The volcano-sedimentary series strikes 70° to 100° and dips at 15–20° (up to 30°) degrees to the north. The thickness of the older volcanic sequence is around 1000–1100 m (only 800 m; after Pucharovsky, 1947), while it is up to 30 m for the younger sequence. Four main volcanic lithologies have been mapped. These are described in more detail below.

Basaltic (s.l.) lava-flows, pillow-lavas and pyroclastic fallout

Massive or sometimes vesiculated lava-flows (metres to tens of metres in thickness), interstratified with thin and lenticular flow-breccias, comprise the basaltic pile which has a total thickness of up to 600 m (Figs 3, 4, 7 & 8). Lenticular, thin (1–5 m) and rather poorly-sorted lapilli- and ash-fall tuffs precede, and are interstratified with, this series of lava flows. The scarcity of pyroclastic deposits suggests fissural eruptions occurring over a relatively short period of time. In addition, all of these volcanic products are interstratified and overlain by thin and lenticular tuffaceous microconglomerates and arkoses (some metres in thickness), which were deposited in a fluvial environment (see previous section). Pyroclastic deposits related to fallout in a subaqueous environment were not observed. All of these features suggest continental subaerial volcanic activity. An interesting lateral variation exists in the eastern part of this basin margin (SE of Razdolnoe; Fig. 5) where pillow lavas form a chaotic edifice with a flat shape, *c.* 10 m thick at the base of the basaltic sequence, and directly overlying the granitic basement. The absence of any sedimentary deposits between the basement and the basaltic pillows suggests subaqueous emplacement on a rather rapidly subsiding basin floor at the time of eruption. All of these basaltic units are probably ?Lateate Givetian–early Frasnian in age (Fig. 5). Volcanic activity was thus coeval, being subaerial in the west and subaqueous in the east.

Petrographically, these basalts (picrobasalts to basalts, basanites and trachy-basalts (cf. LeBas *et al.* 1986; Chalot-Prat, unpublished data) are microlitic, mostly porphyritic, and frequently microvesiculated. The microlitic texture is often fluidal, suggesting low magma viscosity during emplacement. Vesiculation occurred during and/or after flow. In the pillow basalts, quenched textures on plagioclase microlites are frequent. Phenocrysts (1 to 8 mm in diameter; 1 to 20% in volume) are predominantly Ti-augite (frequent sector-zoning and concentric zoning), and/or pseudomorphs of olivine, and/or plagioclase, and magnetite. The main microlites present include plagioclase, augite and magnetite.

Dacitic (s.l.) dykes and extrusions

These are found either as dykes (metres in thickness) within the Priazov Massif and the basaltic units, or as extrusive domes intercalated between the basaltic flows and the overlying rhyolitic pyroclastic fallout and flow deposits (Figs 5, 7 & 8). Near the centre of the Styla Block, the rhyolitic deposits are absent and there the dacites, and their caps of flow breccias, are directly overlain by intra-Lower Viséan trachytic lava-flows. The dacitic domes are rather flat (500–750 m in diameter and 10–25 m in height) with well-developed columnar jointing perpendicular to the convex upper shape. At the central dome axis, the planar jointing is concordant with the stratification of sedimentary deposits below and above the domes. These edifices result from the localized emplacement of small volumes of highly viscous magma.

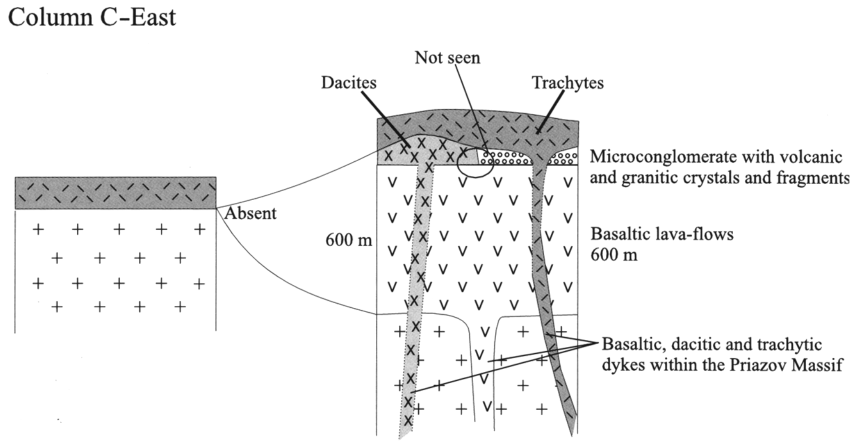


Fig. 7. Stratigraphic column from the eastern part of the study area (column C–east).

In terms of their petrography, these light-grey dacites (cf. LeBas *et al.* 1986; Chalot-Prat unpublished data) are always porphyritic, with a texture which normally varies from microlitic to microgranular towards the core of the extrusion. Phenocrysts (1 to 5 mm in diameter; 10 to 30% in volume) are mainly green–brown amphibole (frequent concentric zoning often around a resorbed blue–green core) associated with Fe-augite (or hedenbergite?), plagioclase, titanite, apatite, magnetite and rare biotites. Microlites are mainly plagioclase (and alkali feldspars?) associated with amphibole and augite. Amphibole and clinopyroxene frequently display resorption features (sinuous contours and resorbed cores), attesting to magma mixing, probably within the same reservoir. Some rounded or shredded metamorphic quartz xenocrysts are also observed.

Based on their geometric relationships with the surrounding rocks, together with their internal structure and texture, these domes are interpreted as extrusive. They are probably Upper Devonian in age, although they had previously been mapped as Permian- or Triassic-age intrusive bodies.

Rhyolitic pyroclastic flow, surge and fallout deposits

These pyroclastic deposits cover a limited area (Fig. 3) in the western part of the study area. Elsewhere they are absent (Fig. 6) or rare (Fig. 5). In Figure 3, it can be seen that these units form an uninterrupted volcanic sequence of about 500 m in thickness above the basaltic flows or the dacitic domes and below the grey Upper Devonian–early Lower Viséan limestones. These green, pink, white, or even black, rocks form lenticular stratified deposits resembling fine-grained shales. They are often massive, but may also have planar jointing. Some regularly stratified layers are well sorted and show direct or inverse graded-bedding; some display internal oblique stratification. No sedimentary deposits are intercalated within these pyroclastic deposits, attesting to the rapidity of deposition of these units. Previous studies (e.g. Makoukina 1961; Rotai 1934; Puscharovsky 1947) were confined to the easternmost part of the study area, and did not attempt to integrate their findings into a broader regional interpretation.

Petrographically, these rhyolites (ultrapotassic rhyolites – cf. LeBas *et al.* 1986: Chalot-Prat unpublished data) are medium to fine grained with an unwelded pyroclastic (lapilli and/or ash size) texture. Varying percentages of phenoclasts are present, including alkali feldspars and/or quartz, well-preserved pumice fragments and shards, and small microlitic lava xenoliths, all of which are in a fine ash matrix. Depending on both the internal structure of the individual layers and the degree of sorting of the pyroclasts, these lithologies may be interpreted as subaerial ignimbritic flow units (*sensu* Freundt & Rosi 2001) including deposits of both pyroclastic flows (massive without any clast sorting at thin section scale), or surges (oblique stratification), and also fallout (well stratified with frequent sorting of clasts according to their composition and size: crystal tuffs, pumice tuffs, etc.). Furthermore, these ash and lapilli fallout layers frequently include well-preserved Upper Famennian plant remnants (Rotai 1944; Puscharovsky 1947) without any sedimentary clasts. The plants were incorporated with the pyroclasts, following explosive eruption. The rare lithic clasts comprise microlitic lava and/or rhyolitic ignimbrites from previous deposits.

It should also be noted that the rhyolitic ignimbrites directly overlie the previously described dacitic domes. The well-preserved texture of pumice fragments and shards of these pyroclastics precludes the possibility of any contact metamorphism by a subsurface intrusion, which is an additional argument in favour of extrusive, rather than intrusive, emplacement of the dacitic domes.

Trachytic sills, dykes and extrusions

These occur as dykes (metres to tens of metres in thickness), or sills, cross-cutting the basaltic flows and the Priazov basement to the south, and as vertical dykes and sills, 50-200 cm in thickness, within the ?upper Famennian–lowermost Viséan limestones. They also form domes or restricted lava flows occurring above the previously described basaltic flows and/or the dacitic domes. These extrusions are directly overlain by the upper Lower Viséan fine-grained silicic or calcareous sediments (Fig. 6). Domes and lava-flows are rather thin (up to 30 m) and typically display columnar jointing. The uppermost *c.* 1 m of the domes is weathered. The overlying sediments display a texture and composition which shows no evidence of any metamorphic alteration (neither baking, nor the growth of contact metamorphic minerals). This would suggest subaerial, rather than intrusive emplacement of these trachytic rocks, as had been previously suggested (Makoukina 1961), followed by their *in situ* alteration (this alteration occurred in a tropical or equatorial climate setting and was synchronous with the karstification of the early Lower Viséan limestones), and subsequent rapid subsidence and marine flooding of the basement. The numerous dykes cross-cutting the basaltic pile, and sometimes the Styla Block (Fig. 8) and the absence of any related pyroclastic deposits, suggest fissural eruptions and a rather short volcanic episode.

In terms of their petrography, these trachytes (cf. LeBas *et al.* 1986; Chalot-Prat unpublished data) are vesiculated, finely microlitic and porphyritic. Phenocrysts (up to 6 mm in diameter; up to 20% in volume) and microlites (0.5 mm in diameter; up to 40% in volume) are mainly plagioclase (with frequent concentric zoning), alkali feldspar, green amphibole, titanite, apatite and magnetite. Phenocrysts are often partially resorbed, which suggests some disequilibrium related to magma mixing within the reservoir.

Structural setting

The region of study on the margins of the southern Donbas is bounded by the WNW–ESE-trending Vassilievka Fault to the south and the WNW–ESE-trending Yuzhni Fault to the north (Fig. 2). The latter fault dips approximately 70° to the south. Structurally, the area adjoins the southeastern flank of the Kalmius–Torets Depression and consists of the oldest exposed sedimentary sequence within the Donbas

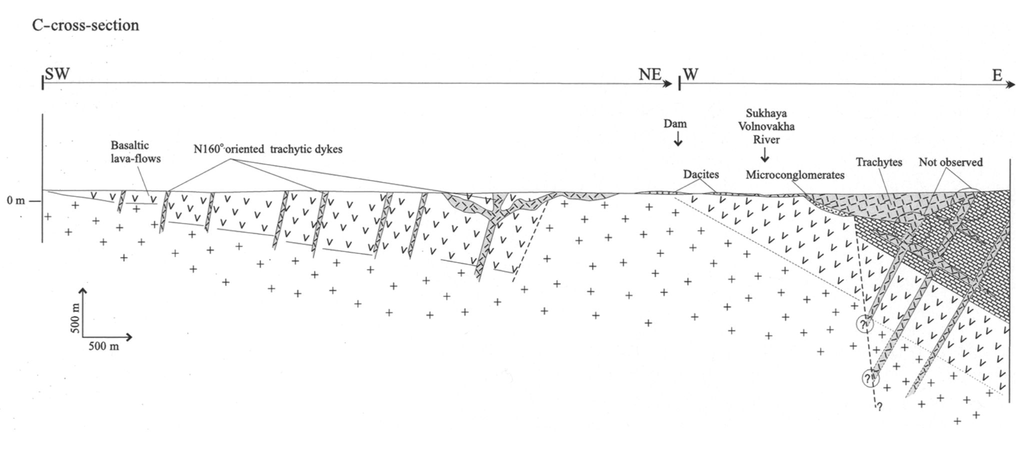
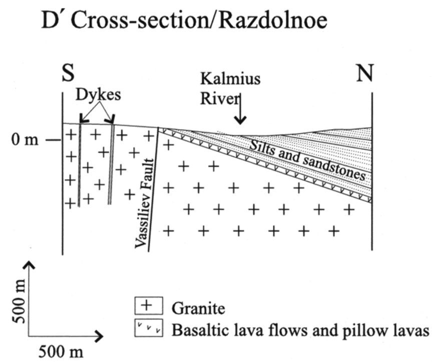


Fig. 8. Geological cross-section from the eastern part of the study area (cross-section C). See Figure 2 for a precise location.

Basin (Fig. 1). The structural setting of the southern Donbas region has been investigated in detail over three field seasons. This work has led to the compilation of a regional map and a series of cross-sections which illustrate the structural setting. The sediments, which are predominantly NNE-dipping (average dip 10–15°), directly overlie a Precambrian basement (granites, granodiorites, migmatites and amphibolites; see Makoukina 1961). A series of cross-sections clearly shows the half-graben structures which developed across the area in response to Devonian and subsequent Viséan rifting (Figs 4, 6, 8 & 9). According to our fieldwork, the typical rift-induced pattern of half-grabens was produced by the northern fault system at a regional scale. Detailed palaeostress analysis (using micro-fault inversion) confirms the normal nature of the faulting. This micro-tectonic study has also recognized a NNE–SSW tensional trend under which the WNW–ESE-trending normal marginal faults have moved (Saintot *et al.* 1999) and no evidence of oblique movements was recorded at the times of rifting phases (i.e. Late Devonian and Early Viséan).

In the western part of the study area there is clear evidence of several normal movements along the approximately E–W-trending faults of the northern part of the region (along which the half-graben developed, Figs 4 & 6). In Figure 6 a clear angular unconformity exists between the sandstones which dip 15°N and the overlying limestones which dip 8°N. This unconformity is even more pronounced on a second cross-section



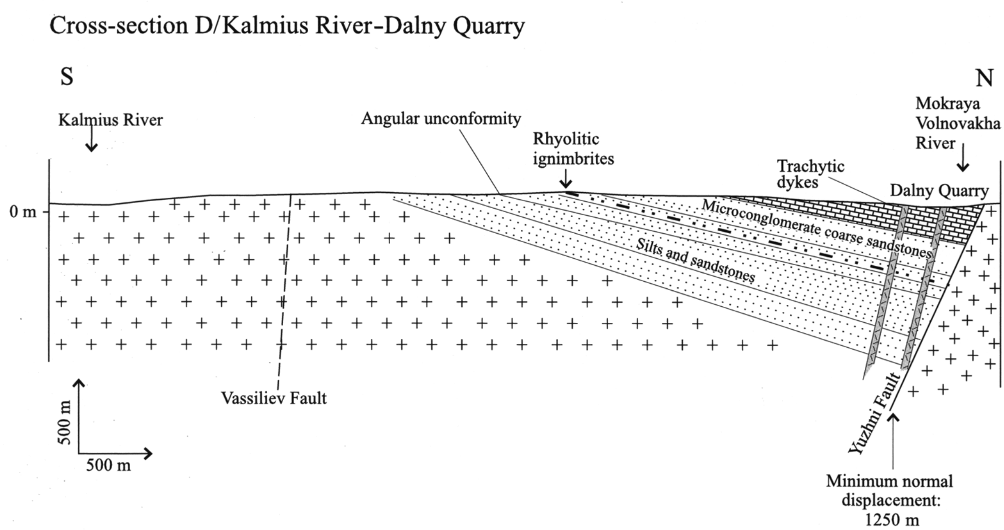


Fig. 9. Geological cross-sections from the Kalmius River–Dalniy Quarry area (cross-section D) and the Razdolnoe area (cross-section D′). See Figure 2 for a precise location.

(Fig. 4) where the sandstones dip 20-25°N and the overlying limestones dip 10-15°N. The development of this unconformity is consistent with published seismic interpretations (Stovba *et al.* 1995, 1996; Stovba & Stephenson 1999). The presence of the unconformity would suggest that tilting to the north occurred both during the Late Famennian (i.e. between the deposition of the youngest late Famennian sandstones and the overlying late Famennian limestones) and after carbonate sedimentation along the northern WNW–ESE fault zone (Fig. 2). In addition, along the northernmost part of cross-section B (Fig. 4), the dip of the limestones (20–25°N) is more than that of the underlying strata (10°N), suggesting the presence of a fault along which another half-graben developed. This interpretation would be consistent with the general structural pattern which developed in the region (see Fig. 4).

In the easternmost part of the region (Fig. 9), a weak angular unconformity exists within the Middle–Late Devonian sediments. There is, however, no angular unconformity at the base of the overlying limestone. In constrast to the western area (see Figs 4 & 6), tilting occurred earlier during Devonian times (possibly in the Givetian to Frasnian) rather than during the Late Famennian.

The cross-sections show that there is abundant evidence of tectonic activity throughout the entire region. This was probably related to normal reactivation of the basin-bounding faults (e.g. the Yuzhni Fault Zone). This phase of tectonic activity would correspond with the Early Viséan phase of rift reactivation which was postulated on the basis of seismic evidence (cf. Stovba and Stephenson 1999). If this is the case, then we can assume that this phase of tectonic tilting produced an efficient slope on which the upper Lower Viséan chert units could have slumped. Moreover, synsedimentary normal faults have been observed within these upper Lower Viséan chert units, suggesting evidence of contemporaneous tectonic activity.

The chronology between the trachytic volcanism (described above from dykes and domes) and the tilting of units, i.e. normal movement along the marginal fault zone, is constrained assuming that the dykes were vertical when they developed (see cross-sections with tilted dykes). Crustal fracturing facilitated magma ascent and eruption in a subaerial environment which was in the process of karstification. Thus, subaerial trachytic volcanism was synchronous with karstification. Tilting related to normal movement along the basin-bounding fault, i.e. the Yuzhni Fault Zone which forms the northern boundary of the half-graben in Figure 9, occurred after uplift (and emergence) and was synchronous with the slumping of the chert-rich unit.

Discussion

Along the southern margin of the Donbas fold belt, the analysis of the interlinked tectonic, sedimentary and magmatic history has revealed the evolution for this area from Middle Devonian through to Early Viséan times (Fig. 10). In some ways this evolution mirrors that of the Dniepr–Donets region to the northwest, but in others it is clearly different. A significant difference is that, within the study area, a synrift event – without any pre-rift event as in the Pripyat–Dniepr–Donets system – occurred at the onset of the Devonian. This initial rifting event was active and not passive inasmuch as it began with a period of uplift and associated volcanism, both being related to mantle-plume activity (Wilson & Lyashkevich, 1996), rather than simple passive extension of the lithosphere. This initial active synrift event is evidenced by numerical models of the region (Van Wees *et al.* 1996), but has not been previously demonstrated by field studies.

Thus, the region was subjected to a series of tectonic events, beginning in Middle Devonian times, which led to the formation of a series of interrelated half-grabens. The development of integrated tectono-sedimentary successions within these half-grabens (*sensu* Leeder & Gawthorpe 1987), however, was complicated by the series of related magmatic events. Within these half-grabens, the volcanic succession comprises *c.* two-thirds of all recorded strata within the basins. This fact must be kept in mind when attempting to interpret the tectonic and sedimentological evolution of the region.

Analysis of the sediments which occur in fault-bounded half-graben basins has been the subject of much research (e.g. Coward *et al.* 1987). In particular, Leeder & Gawthorpe (1987) used the occurrences of discrete sediment packages and predictable sequences to propose tectono-sedimentary models. The model which most closely resembles the present area is a continental basin with axial drainage which evolved over time into a carbonate-dominated shelf succession.

In the study area, initial basin infill occurred following the deposition of an Eifelian–Givetian clastic, mainly continental sequence. Subsequent fault activity led to the initiation of a series of half grabens. The Middle Devonian fluviatile deposits contain mainly material derived from the underlying basement, providing evidence of its exposure and active erosion at this time. The

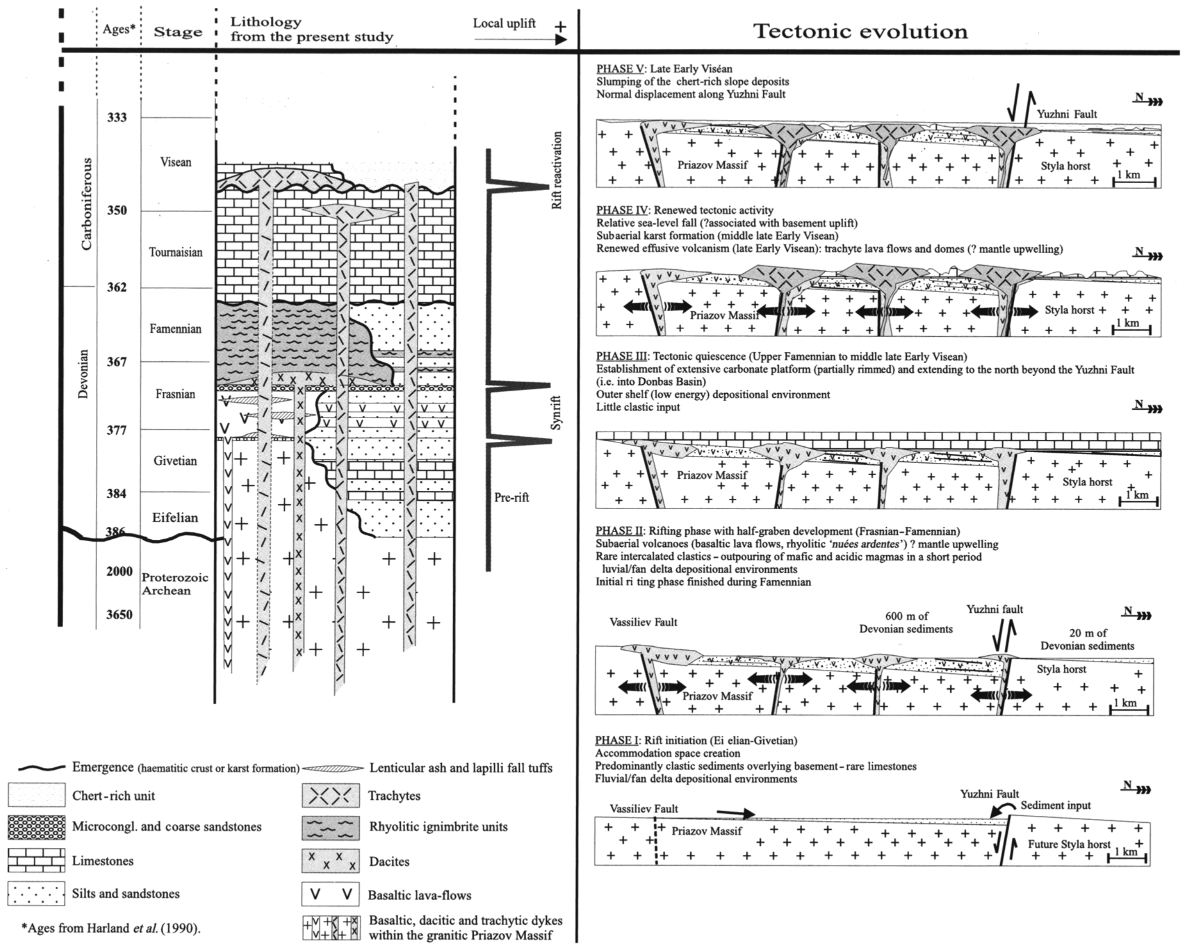


Fig. 10. Schematic stratigraphic section from the southern margin of the Donbas (left-hand side), together with a series of schematic cross-sections illustrating the tectono-magmatic–sedimentary evolution of the region from Eifelian to late Early Viséan times.

northerly Yuzhni Fault was the main graben-bounding fault, with activity along the Vassilievka Fault being more restricted. This latter fault was one of a series of parallel faults which developed along the southern margin of the basin, and within the Priazov Massif. These faults provided a conduit for ascending magmas in the region, and also account for the numerous dykes of doleritic basalts, dacites and trachytes within the Proterozoic basement. (Indeed, removal of a large part of this half-graben system is also a possiblity that must be considered.) During this initial phase of rifting, volcanic activity was not completely absent, as evidenced by rare volcanic lithic fragments, and even a small channel cut into an underlying basaltic unit. This would support the suggestion that crustal-scale fracturing was present right from the initial stages of rifting. Normal faulting was restricted to localized movement along the Yuzhni Fault, leading to the development of a weak intra-Devonian angular unconformity.

The first tectonic pulse was accompanied by subsidence and the establishment of a fluvial depositional system across the area with possible local lacustrine carbonates. Basaltic eruptions were significant, and the outpouring of such large volume of basaltic magma at this early stage of basin formation in the southern Donbas region suggests that the first feeder dykes extended as far as the mantle and were probably related to major crustal fractures. The lack of pyroclastic edifices suggests that the basaltic volcanism was fissural. Furthermore, the absence of intercalated clastic deposits within the entire basaltic unit would suggest a rapid magmatic outpouring, possibly coeval with a significant but short-lived decrease in the subsidence rate during eruptive phases.

Following cessation of the volcanic activity, fluvial depositional environments were re-established across the region. A subsequent phase of volcanic activity involved the emplacement of dacitic dykes and domes sporadically across the region followed by outpourings of rhyolitic nuées ardentes. This second volcanic period was synchronous with the deposition of the fluvial succession further to the east (Fig. 5), with tectonic instability extending until the end of this phase of volcanism. There is a clear angular unconformity between the mainly Devonian synrift succession and the overlying ?Upper Famennian-lowermost Viséan carbonate platform sediments which suggests that the onset of subsidence across the region was not uniform, and involved a degree of tilting of the strata.

Volcanic activity in Devonian times was mainly subaerial, as evidenced both from the sediments and also from features within the volcanics themselves. There is also evidence of some restricted and localised subaqueous eruption (e.g. lacustrine ?pillow lavas). Eruptions were fissural and the duration of each volcanic episode was always quite short (*c.* 1Ma). This would suggest not only the short-lived opening of the magma conduits but also that magma generation was accompanied by brittle crustal deformation whereby the resulting faults facilitated the passage of magmas to the surface.

A subsequent pulse of tectonic activity was recorded for the late Early Viséan, leading to the uplift and exposure of the previously deposited carbonates and localized karst formation. This marked the end of the Early Carboniferous period of basin subsidence. Renewed tectonic activity was accompanied by the sub-aerial extrusion of a trachytic magma. Associated trachytic dykes cross-cut the early Early Viséan carbonates. This phase of trachytic volcanism closely followed, or was perhaps even coeval with, the period of limestone karstification interpreted here in terms of basement uplift, rather than sea-level fall. It should also be noted that the Upper Famennian to Lower Viséan limestones are missing in the south-central part of the Styla Block. The block presumably remained continental and tectonically stable throughout this period. The precise sequence of events which follows is unclear – largely as a result of the restricted exposure of the slumped chert deposits. If these deposits are, indeed, marine in origin (which has been suggested by Russian colleagues), then the phase of renewed subsidence, and possible sea-level rise (Ross & Ross 1988) led to the deposition of this chert-rich siliciclastic unit which overlies the platform carbonates. This period of basin subsidence may also have exhibited a degree of tectonic instability, leading to the formation of major slide deposits produced as a result of destabilization of the sediments and their downslope movement as a result of gravity sliding. However, while other mechanisms for the destabilization of this sediment body must also be considered, the broad regional extent of the deformation would suggest that the underlying cause was one which could cause instability on a broad scale. Given that trachytic volcanism – and uplift – occurred before normal faulting (i.e. tilting of all units), the Viséan phase of rifting is in good agreement with an active rifting model. A second interpretation, however, must also be provided. This is where the chert-rich deposits are interpreted as having been deposited within a lacustrine environment. This would, therefore, mean that the period of emergence, signalled by the karstification of the limestones, continued with the establishment of a lacustrine environment within the region. As already noted, the areal extent of these lakes is impossible to determine due to the restricted outcrop. However, the fine-grained nature of the sediment, together with the general geological setting would support such an interpretation. Silica-rich deposits have been reported from a number of lakes (Talbot & Allen 1996). The source of silica in these environments is both from silica-rich organisms within the lake, but also in association with hydrothermal activity, or as a product of alkaline conditions (e.g. Eugster 1980).

Throughout the synrift period, the observed changes in magma composition and magma volume from one volcanic episode to another (i.e. basalt–dacite–rhyolite–trachyte) indicates that partial melting and magma mixing involved successively different levels of the mantle and crustal lithosphere. In addition, both partial melting and/or rock-source volume decreased with time. For both the rhyolitic and dacitic events, crustal fracturing was short lived and less deep than during the period of basalt eruption (i.e. down to middle/lower crustal levels only). The cause of this shift in the source of partial melting within the lithosphere, together with the rather transient fracturing of the continental crust, could be related to lithospheric thinning at this particular stage of Donbas Basin formation. Such a conclusion would be supported by the work of Van Wees *et al.* (1996), who noted that Late Devonian–Early Carboniferous stretching resulted in a lithosphere thickness of *c.* 50 km. It should also be noted that the eruption of trachytes, which marks the last phase of volcanic activity in the study area, occurred in Early Viséan times, when rift reactivation took place across the entire Dniepr–Donets–Donbas system (Stovba & Stephenson 1999).

The relationship between volcanism and sedimentation is interesting since each eruptive cycle is preceded by a period of clastic deposition, the sediments of which record relief formation and erosion. However, the majority of the lithic fragments within the fluviatile successions were predominantly derived from the underlying crystalline basement rather than from the volcanic successions. This would suggest that the drainage basins supplying these rivers were concentrated within the shield area. Thus, sediment was mainly derived from these crystalline uplands, with limited percentages being derived from the coeval volcanics within the half grabens themselves. An additional point, however, is that the predominance of large rivers would suggest that transport distances were also greater – another factor leading to a decrease in volcanic material.

Conclusions

A series of tectonic pulses has been recorded in the Palaeozoic evolution of the southern Donbas region. Alekseev *et al.* (1996) noted that prior to the onset of crustal extension in the Dniepr–Donets area, the Ukrainian Shield and Voronezh High were covered by marine Middle Devonian sediments (not recorded in the present study area). The onset of crustal extension was contemporaneous with magmatism (Wilson & Lyashkevich 1996) and uplift which was highest to the south of the present southern margin of the Donbas Basin and on the Styla Block. Initial sedimentation involved the establishment of a localized fluvial and lacustrine depositional environment within the first half-grabens which developed between the main basin-bounding faults (Eifelian–Early Givetian). The first phase of synrift activity (during Eifelian times) was probably associated with some initial basaltic extrusion. The Late Givetian–Early Famennian period was one of pronounced tectonic, volcanic and sedimentary activity. Tectonic activity involved both the upper crust – with the formation of a series of half-grabens and the resultant subdivision of the region into several blocks extending in an E–W direction, and the lower crust – with the formation of crustal-scale fractures which enabled the mantle and crustal magmas to ascend to the surface. The resultant volcanic pile covered the central part of the study area, while fluvial deposition was restricted to the margins. Each volcanic episode was preceded by a phase of renewed uplift, which may have been localized (i.e. doming resulting from magmatic upwelling). The end of this major volcanic period marked the onset of a phase of basin subsidence, and the establishment of carbonate deposition across the region. Subsidence variations led to the formation of a slight angular unconformity between the clastic Middle–Upper Devonian succession and the overlying carbonate platform sediments.

A subsequent uplift phase led to the exposure of the entire platform and resultant karstification. This period of karst formation was coeval with the subaerial extrusion of trachytes (also subjected to extreme chemical weathering). It should be emphasized that the numerous trachytic dykes not only occur within the underlying limestones, but also within the basaltic pile and the Priazov Massif – up to 50 km to the south of the present-day margin of the Donbas. The widespread nature of this event attests to a fissural genesis, and also for significant fracturing of the upper crust (trachytes are most likely derived from lower crustal melts: Chalot-Prat, unpublished data) during intra-Early Viséan times. In addition, this spatially extended partial melting of the crust may be linked to basalt intrusion, which supports the idea that emersion was related to thermal uplift of the lithosphere rather than a relative or eustatic sea-level fall.

During the late Early Viséan, two possibilities are suggested for the formation of the siliceous unit. The first involves the development of a lacustrine environment while the second involves a marine transgression which flooded much of the karstified surface and the subsequent deposition of the unit. This phase would correspond with the continuing synrift activity in the region (which began with the trachytic volcanism). Deposits from the overlying carbonates, which are Late Viséan in age, may correspond with the true beginning of the post-rift phase, with related subsidence, of basin development. This siliclastic unit exhibits a variety of wet-sediment deformation features, including slumping, folding and faulting, which can be interpreted as being related to the instability of the basement as a result of renewed activity along the Yuzhni Fault.

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References

AISENVERG, D. E., LAGUTINA, V. V., LEVENSTEIN, M. L. & POPOV, V. S. 1975. *Field Excursion Guidebook for the Donets Basin.* C. R. 8th Congress on International Strat. Geol. Carboniferous, Moscow, 1975, 360 pp.

ALEKSEEV, A. S., KONONOVA, L. I. & NIKISHIN, A. M. 1996. The Devonian and Carboniferous of the Moscow Syneclise (Russian Platform): stratigraphy and sea-level changes. *Tectonophysics,* **268**, 149–168.

CHEKUNOV, A. V., KALUZHNAYA, L. T. & RYABCHUN, L. I. 1993. The Dnieper–Donets Paleorift, Ukraine: deep structures and hydrocarbon accumulations. *Journal of Petroleum Geology,* **16**, 183–196.

COWARD, M. P., DEWEY, J. F. & HANCOCK, P. L. (eds) 1987. *Continental Extensional Tectonics.* Geological Society Special Publication, **28**, 139–152.

EUGSTER, H. P. 1980. Lake Magadi, Kenya, and its precursors. *In:* NISSENBAUM, A. (ed.) *Hypersaline Brines and Evaporitic Environments.* Developments in Sedimentology, **28**. Elsevier, Amsterdam, 195–232.

FREUND, A. & ROSI, M. (eds) 2001. *From Magma to Tephra: Modelling Physical Processes of Explosive Volcanic Eruptions.* Elsevier, Amsterdam, 318 pp.

Geological Map of the USSR and Adjoining Water-covered Areas, Scale 1:2 500 000, Minesterstvo Geologii SSSR, BSEGEI, 1983.

Geological Map of the Ukrainian Donbas Foldbelt, Scale: 1:200 000, Artemovsk Geological Survey, Ukraine, 1995.

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 University Press, Cambridge, 263 pp.

KUSZNIR, N. J., STOVBA, S. M., STEPHENSON, R. A. & POPLAVSKII, K. N. 1996. The formation of the northwestern Dnieper–Donets Basin: 2D forward and reverse synrift and post-rift modelling. *Tectonophysics,* **268**, 237–255.

LEBAS, M. J., LE MAITRE, R. W., STREICKEISEN, A. & ZANETTIN, B. 1986. A chemical classification of volcanic rocks based on the total alkali–silica diagram. *Journal of Petrology,* **27**, 745–750.

LEEDER, M. R. & GAWTHORPE, R. L. 1987. Sedimentary models for extensional tiltblock/half graben basins. In: COWARD, M. P., DEWEY, J. F. & HANCOCK, P. L. (eds) *Continental Extensional Tectonics.* Geological Society Special Publication **28**, 139–152.

MAKOUKINA, G. O. 1961. Petrography of the dyke effusive complex of the SW Donbas. *In:* TKACHUK, L. G. (ed.) *IGS Proceedings,* Petrography, Mineralogy and Geochemistry, Issue **15**, 172 pp. (in Ukranian).

NIKISHIN, A. M., ZIEGLER, P. A., STEPHENSON, R. A., CLOETINGH, S. A. P. L., FURNE, A. V., FOKIN, P. A., ERSHOV, A. V., BOLOTOV, S. N., KOROTAEV, M. V., ALEKSEEV, A. S., GORBACHEV, V. I., SHIPILOV, E. V., LANKREIJER, A., BEMBINOVA, E. YU. & SHALIMOV, I. V. 1996. Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution. *In:* STEPHENSON, R. A., WILSON, M., DE BOORDER, H. & STAROSTENKO, V. I. (eds) *EUROPROBE: Intraplate tectonics and basin dynamics of the eastern European Platform. Tectonophysics,* **268**, 23–63.

OWEN, R. B. & CROSSLEY, R. 1992. Spatial and temporal distribution of diatoms in sediments of Lake Malawi, central Africa, and ecological implications. *Journal of Palaeolimnology,* **7**, 55–71.

POLETAEV, V. I. *ET AL.* 1988. The guidebook to excursion to the Lower/Middle Carboniferous boundary deposits of Donbas. *In:* AISENVERG, D. YE. (ed.) *IGS Publication,* 92 p. (in Russian and English).

PUSCHAROVSKY, Y. M. 1947. Devonian sediments of southern outskirts of the Donets Basin. *In:* SHATSKY, N. S. (ed.) *Materials on Geology of the Donets Sediments of the Southern Outskirts of the Donbas.* MOIP Press, New Series, **9(13)**, 5–21 (in Russian).

RENAUT, R. W. & OWEN, R. B. 1988. Opaline cherts associated with sublacustrine hydrothermal springs at Lake Bogoria, Kenya Rift. *Geology,* **16**, 699–702.

ROSS, C. A. & ROSS, J. R. P. 1988. Late Paleozoic transgressive–regressive deposition. In: WILGUS, C. K., HASTINGS, B. S., KENDALL, C. G. ST. C., POSAMENTIER, H. W., ROSS, C. A. & VAN WAGONER, J. C. (eds), *Sea-level Changes: an Integrated Approach.* Society of Economic Paleontologists and Mineralogists, Special Publication **42**, 227–247.

ROTAI, A. P. 1944. *Geological Works in the Donets Basin during the Summer of 1928–1929.* Proceedings VGRO, 356 (in Russian).

SAINTOT, A., PRIVALOV, V., ZHIKALIAK, M., BREM, A. & EUROPROBE/INTAS 1999. Some kinematic indicators for the tectonic evolution of the Donbass fold-and-thrust belt (Ukrainian part). Joint meeting of EUROPROBE TESZ/PANCARDI/GEORIFT, Tulcea (Romania). *Romanian Journal of Tectonics and Regional Geology*, **77**, p. 83.

STEPHENSON, R. A. & EUROPROBE INTRAPLATE TECTONICS AND BASIN DYNAMICS DNIEPER–DONETS AND POLISH TROUGH WORKING GROUPS 1993. Continental rift development in Precambrian and Phanerozoic Europe: EUROPROBE and the Dnieper–Donets Rift and Polish Trough basins. *Sedimentary Geology,* **86**, 159–175.

STOVBA, S. M. & STEPHENSON, R. A. 1999. The Donbas fold belt: its relationships with the uninverted Donets segment of the Dniepr–Donets Basin, Ukraine. *Tectonophysics,* **313**, 59–83.

STOVBA, S. M., STEPHENSON, R. A. & DVORIANIN, E. 1995. Dniepr–Donets Basin, Ukraine: main observations from regional seismic profiles. *Comptes Rendus des de l'Académie des Sciences, Paris*, **321**, Série IIa, 1103–1110.

STOVBA, S., STEPHENSON, R. A. & KIVSHIK, M. 1996. Structural features of the Dniepr–Donets Basin, Ukraine from regional seismic reflection profiles. *Tectonophysics,* **268**, 127–147.

TALBOT, M. R. & ALLEN, P. A. 1996. Lakes. *In:* READING, H. G. (ed.) *Sedimentary Environments: Processses, Facies and Stratigraphy.* Blackwell Science, 83–124.

VAN WEES, J.-D., STEPHENSON, R. A., STOVBA, S. M. & SHIMANOVSKY, V. 1996. Tectonic variation in the Dniepr–Donets Basin from automated modelling of backstripped subsidence curves. *Tectonophysics,* **268**, 257–280.

WILSON, M. & LYASHKEVICH, Z. M. 1996. Magmatism and the geodynamics of rifting of the Pripyat–Dnieper–Donets rift, East European Platform. *Tectonophysics*, **268**, 65–81.

ZIEGLER, P. A. 1990. Collision-related intra-plate deformations in Western and Central Europe. *Journal of Geodynamics,* **11**, 357–388.