Paleostress field reconstruction and revised tectonic history of the Donbas fold and thrust belt (Ukraine and Russia)

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[1] In the WNW-ESE Donbas fold belt (DF), inversion of 3500 microtectonic data collected at 135 sites, in Proterozoic, Devonian, Carboniferous, and Cretaceous competent rocks allowed reconstruction of 123 local stress states. Accordingly, four successive paleostress fields reveal the tectonic evolution of the DF. At the numerous sites that have been affected by polyphase tectonics, the chronology between local paleostress states (also paleostress fields) was established using classical criteria (crosscutting striae, pre- or post-folding stress states, stratigraphic control). The oldest event is an extensional stress field with NNE-SSW σ3. It corresponds to the rifting phases that generated the basin in Devonian times and its early Visean reactivation. Later, the DF was affected by a transtension, with NW-SE early Visean reactivation. Later, the DF was affected that generated the basin in Devonian times and its European Craton.

INDEX TERMS: evolution of the southern margin of the East European Craton.

[2] The Donbas fold and thrust belt (DF) is the strongly inverted and compressively deformed part of the Dnieper-Donets Basin (DDB), a Late Devonian rift basin located on the southwestern part of the East European Craton (EEC), in eastern Ukraine and in southern Russia (Figure 1a). Further to the southeast the DF joins the contiguous, deformed southern margin of the EEC (Karpinsky Swell). The width of the original rift basin (shaded in Figure 1a) varies between 60–70 km in the northwestern and 140–160 km in the southeast. Thicknesses of the Late Palaeozoic and younger sedimentary succession increase from only about 2 km in the northwestern to about 23 km in the DF and most of this is of Carboniferous age [e.g., Chekunov et al., 1993; Chekunov, 1994]. Devonian rifting was accompanied by major magmatic activity and the uplift of the Ukrainian Shield and the Voronezh Massif, forming a large radius arch that is transected by the DDB [Gavrish, 1989; Wilson and Lyashkevich, 1996].

[3] There are profound along strike variations in the degree of basin “inversion” in the DDB, ranging from severe in the DF to practically none in the Dnieper segment [Chirvinskaya and Sollogub, 1980; Stephenson et al., 2001]. There is a major Permian unconformity, with increasing thicknesses of eroded strata inferred to the south-east, and it has long been regarded that basin “inversion” was related to Permian Variscan/Uralian orogenesis on the margins of the EEC [e.g., Milanovsky, 1992] or to the activity of an asthenospheric (mantle) diapir [Gavrish, 1985, 1989; Chekunov, 1994]. Several kilometers of mainly Carboniferous strata have been eroded in the DF, especially on its southern margin [e.g., Stovba and Stephenson, 1999]. What is known about the subsurface structure of the DF has been based on surface exposure, shallow boreholes and deep sounding profiles [Stovba and Stephenson, 1999], to be presently augmented by new deep seismic reflection data still being interpreted [Roy-Chowdhury et al., 2001].

1. Introduction
and, similarly, the mechanisms leading to uplift and compressional deformation in the inverted (DF) part of the DDB are also problematic. The geological setting of the DF is complicated by the increasing proximity of the basin axis to the inferred southern edge of the EEC and its presumed relationship with contemporaneous basin development on the southern margin of the EEC. Recently, Stovba and Stephenson [1999] presented seismic reflection data in the southeastern DDB (Donets segment) documenting that Late Palaeozoic reactivation (syn-Variscan/Uralian) were (trans)lational rather than compressional, while those at the end of the Triassic and at the end of the Cretaceous were compressional in nature. Post-rift reactivations increase toward the southeast and Stovba and Stephenson [1999] surmised that they must be even more profound in the DF than in the uninvited part of the DDB. Accordingly, Stovba and Stephenson [1999] concluded that the main phases of shortening in the DF were Cimmerian (Late Triassic-Jurassic) and Eo-Alpine (end Cretaceous). This was in contradiction to firmly entrenched, long-held concepts prevalent in all published materials on the DF that it forms a part of an incipient (trans)lational rather than compressional, with thin coal and limestone beds. The very thick Middle Carboniferous strata consist of arenaceous-marine sediments with thin coal and limestone beds. The very thick Middle and Upper Carboniferous strata consist of arenaceous argillaceous rocks interbedded with coal and limestone beds.

2. Geology of the Donbas Fold Belt

2.1. Basin Fill

[6] The Devonian to Carboniferous sedimentary succession, deeply buried in other parts of the DDB, is well exposed in the DF (Figure 1b). These sequences have been studied from surface outcrops and coal prospecting boreholes to depths of 1–3 km [Levenshtein, 1963; Popov, 1963; Maidanovich and Radzivill, 1984].

[7] Extensively observed Middle/Late Devonian to Early Carboniferous rocks crop out on the southwestern margin of the DF where they unconformably overlie crystalline basement of the Azov Massif (Figures 1a and 1b). E-W-trending grabens and half-grabens developed possibly as early, but certainly by the end of the Frasnian, and a continental (fluvial, lacustrine) succession was established. Early synrift activity was accompanied by the extrusion of basalts. The Middle/Upper Devonian succession comprises thick clastic and carbonate sediments with interbedded volcanics (1800 m thick, depending on local variations in the quantity of volcanics). The thickness of the Devonian sediments within the axial zone of the DF is thought to be as great as 5 km [Garkalenko et al., 1971; Borodulin, 1974] (Figure 1c).

[8] A broad carbonate platform was established across the region from the latest Devonian, until the late early Visean (1000 m thick, Figure 1d). Otherwise, Carboniferous basin development occurred in a post-rift regime of frequent relative sea-level oscillations leading to continuous rhythmic sedimentation of alternating shallow-water, littoral, off-shore and continental facies, including erosional hiatuses [e.g., Levenshtein, 1963; Dvorjanin et al., 1996; Izart et al., 1996] (Figure 1d). Late Visean sediments consist mainly of thick sandy-clay deposits interbedded with thin coal and limestone beds. The very thick Middle and Upper Carboniferous strata consist of arenaceous argillaceous rocks interbedded with coal and limestone beds (Figures 1b, 1c, and 1d).

[9] No Permian sediments are preserved within the DF itself but Early Permian sediments do occur nearby, along the northwestern margin of the DF in its transition to the DDB. They comprise coastal-continental and occasionally shallow-marine facies, represented by monotonous sand-shale series with sparse interbeds of limestones, coals and salt layers. Magmatic rocks of Early Permian age occur in the southwestern DF [Alexandre et al., 2003] (Figure 1d).

[10] Very little Mesozoic sediment is preserved within the DF. In the nearby part of the DDB the Mesozoic succession consists of marine and continental sediments described as “close to” platform type [Eisenberg, 1988]. In the western part of the northern margin of the DF, Triassic sediments up to 150–200 m thick are preserved in a narrow strip [Belov, 1970]. Jurassic sediments are absolutely absent in the same area [Popov, 1963; Chirvinskaya and Sollogub, 1980].

Figure 1. (opposite) Presentation of the Donbas fold belt. (a) Location of the Donbas in the regional East European structural framework [from Stovba and Stephenson, 1999]. (b) Geological map of the Donbas from Nalivkin [1983]. (c) Geological cross section constructed from surface geology and shallow boreholes plus depth-converted DOBRefection-2000 seismic profile and southernmost part of the seismic profile extension in 2001, located on Figure 1b [from Maystrenko et al., 2003]: MA (60°–80°): Main Anticline and dips of beds of the two limbs; SS: Southern Syncline; SA: Southern Anticline; NS: Northern Syncline; NA: Northern Anticline. (d) Stratigraphic column of the DF with lithologies, thicknesses, magmatic events with new Ar-Ar absolute ages [from Alexandre et al., 2003], and tectonic evolution from Stovba and Stephenson, 1999. (Stratigraphic limits according to the International Stratigraphic Chart, International Union of Geological Sciences: International Commission on Stratigraphy and Commission of the Geological Map of the World, UNESCO, available at http://www.cgmw.org.)
the southwestern part of the DF, Early and Late Triassic and Late Jurassic magmatic rocks crop out [Chekunov and Naumenko, 1982; Alexandre et al., 2003] (Figure 1d). On the southern margin of the DF the Mesozoic is represented by up to 500 m of Upper Cretaceous marls and chalks, unconformably overlying either Palaeozoic rocks or crystalline basement (Figure 1b). Angular unconformities within the Mesozoic succession, in particular at the Triassic/Jurassic and Jurassic/Cretaceous boundaries, have been documented on the northwestern margin of the DF [Konashov, 1980; Eisenverg, 1988]. Palaeogene ( sands, clays, marls) and Neogene ( sands with clayey interbeds) units, unconformably overlying the Upper Cretaceous and older rocks [Eisenverg, 1988].

2.2. Basin Deformation

[11] Deep WNW-ESE faults form the southern and northern boundaries of the DF [Sollogub et al., 1977] (Figures 1b and 1c). Folds in the central part of the DF trend west-northwestward, and are fairly tight, in some places overturned. The Main Anticline (MA) is the largest and most pervasive fold in this zone (Figures 1b and 1c). It is an almost symmetric structure with steeply dipping limbs (60°–80°; Figure 1c), complicated by faults as thrusts (or oblique thrusts), as oblique normal and strike-slip faults developed at its hinge [Lutuguin, 1956], in which dextral movement has been recognized [Maidanovich and Radzivill, 1984; Belichenko et al., 1999; Privalov et al., 2000]. The MA is bordered by two gentle synclines and anticlines (Figures 1b and 1c). Localized folds and thrust faults (as mesoscale structures) are developed in the northern zone of the DF. Carboniferous strata typically dip at angles of 30°, locally to more than 40°. Many folds are tilted northward. Thrust faulting is more regionalized; some thrusts can be traced for many tens of kilometers. Thrust surfaces commonly dip 40°–60° southward. A set of major north-vergent thrusts characterizes the present-day structure of the northern margin of the DF [Belokon, 1975; Mikhailov and Borodulin, 1976] (Figure 1c). Stratigraphic offsets on thrust faults can be substantial (1000 to 2000 m); the maximum is 4000 m [Popov, 1963]. According to Popov [1963] and Zhykalyak et al. [2000], thrusting was episodic, with movements occurring during major tectonic phases at the end of Palaeozoic, in the Mesozoic, and the Cenozoic. To the northwest, toward the uninverted part of the Donets segment, offsets on thrusts decrease and fade out. In the southern zone of the DF, E-W trending minor folds prevail. Near the city of Donetsk (cf. Figure 1b) is a zone of transverse structures [Popov, 1963]. The gentle WNW-ESE folds are overprinted by a widely developed system of strongly asymmetric folds striking northwest, which over long distances possess the characteristics of flexures. Thrust or reverse faults are south vertical. Faults and block structures occur mainly in a corridor between the DF and Azov Massif. Because most of the area is covered with Upper Cretaceous sediments, they are well studied only in the southwestern DF, where Devonian and Lower Carboniferous sediments crop out (Figure 1b). This zone has the appearance of an undulating, northward dipping monocline broken by series of moderately large faults that produce its block structure. Offsets are 400 m and more; dip angles are to 40°–70°, mostly to the southwest (opposite the dip direction of sedimentary horizons). Individual blocks strike mainly to the northwest, subparallel to the main linear folds of the DF.

[12] The main structures of the DF, in particular the MA, are believed to continue eastward, below Mesozoic platform cover in the area of the Karpinsky Swell [e.g., Popov, 1963; Belov, 1970; Garetsky, 1972] (cf. Figure 1a). The morphology of the eroded surface and the gentle structure of the overlying Jurassic and Cretaceous sedimentary cover are thought to reflect inheritance of the strike and character of the folds observed in the DF. It has also been speculated that the Donbas folds and marginal faults can be traced through the Karpinsky Swell to the Caspian Sea [e.g., Belov, 1970; Garetsky, 1972] (cf. Figure 1a).

[13] The characteristics and the tectonic events forming the DF are controversial [cf. Stovba and Stephenson, 1999; Stephenson et al., 2001]. Initial rifting was clearly Devonian in age, but the rifting regime as extensional or transtensional is not known. From observations of the orientations of dykes, mainly from the Azov Massif and thought to be related to the Devonian rifting phase [Muravt, 1972], as well as fractures in the Devonian volcanics of the DF, Korchemagin and Yemets [1987] determined a NNE extensional axis. This was followed by profound post-rift subsidence in the DF during the Carboniferous, as it did in the adjacent DDB, interrupted by rifting phases in late early Visean and in Serpukhovian times [cf. Stovba et al., 1996; van Wees et al., 1996; Stovba and Stephenson, 1999] (Figure 1d). Evidence of the former is amply manifested near the southern margin of the DF as renewed faulting, rapid development of local topographic variations, syn-sedimentary deformation of late early Visean strata [cf. McCann et al., 2003] and an intra-Visean thickening of beds [Garkaleno et al., 1971] (Figure 1c). The DF was uplifted in the Early Permian, especially its southern margin. It is classically thought that the Donbas “fold and thrust belt” was formed during the Permian [Popov, 1936, 1939, 1963; Pogrebitsky, 1937; Stepanov, 1937; Gavrish, 1989] in response to stresses related to the Uralian-Caucasian Variscan orogeny [Milanovsky, 1992]. In this framework, Korchemagin and Yemets [1987] inferred from observations of slickensides a compressive paleostress field, with NNE-SSW compressional axis, which they considered to be related to Permian fold development (though they stipulated a low reliability for the absolute dating of this compressive stress field). However, recent studies have shown that fault deformation of this age is normal in style and that the uplift occurred under transtension-extension phase accompanying a post-rift reactivation [Stovba and Stephenson, 1999]. Gavrish [1985] and Chekanov [1994] argued that the Permian uplift could have been due to a mantle diapir. Synchronous Early Permian volcanic activity in the Scythian Platform (cf. Figure 1a for location) lend supporting evidence [Alexandre et al., 2003]. The occurrence of Cimmerian tectonic compression is observed where
Mesozoic sediments exist [Popov, 1963; Konashov, 1980]; it is recorded through the offsets of Triassic beds along the northern thrust planes [Popov, 1963; Sobornov, 1995]. At the end of Cretaceous times, the DF displays inversion with development at this time of localized folds commonly associated with thrust development [Stovba and Stephenson, 1999; Saintot et al., 2003]. Deformation of this age was recognized in earlier studies (“orogenic” phase according to Popov [1936, 1939]; Stepanov [1937]) but was thought to involve relatively minor reactivation of structures formed during the main fold belt forming events in the Late Palaeozoic. Korchemagin and Yemets [1987] determined a corresponding Alpine strike-slip stress field with a NNW compressional stress axis. Finally, a Paleocene orogenic phase is reported [Popov, 1936, 1939, 1963; Stepanov, 1937].

3. Paleostress Field Reconstruction in the Donbas Fold Belt

3.1. Method and Data: Using Paleostress Analyses to Constrain Tectonic Evolution

[14] The method is based on the kinematics of small-scale brittle structures collected in the field. The kinematic data are inverted to compute stress tensors as described in detail elsewhere [Angelier, 1990, 1994]; such paleostress field analyses have previously been applied in detail in various areas [e.g., Angelier et al., 1985; Sévrier et al., 1985; Bergerat, 1987; Mercier et al., 1987, Hippolyte and Sandulescu, 1996].

[15] A similar study, based on inversion of brittle microtectonic objects, was possible in the DF because most of its exposed stratigraphic succession (with the exception of the Cenozoic) is dominated by lithologies (predominantly limestones and sandstones) that are competent for brittle deformation. About 3500 small-scale brittle tectonic data were observed in the DF at 135 sites, in Proterozoic, Devonian, Carboniferous, Permian and Mesozoic rocks. Fault slip data sets have allowed computation of 82 local stress tensors (with characteristics listed in tables corresponding to stress maps in subsequent sections). Stylolithic peak, diaclace and tension gash orientations, where available, have been used to determine the position of one of the principal stress axes. In places, conjugate systems of shear joints and of en echelon tension gashes also provided information about the attitude of the three principal stress axes. In total, the microtectonic data inversion has resulted in the reconstruction of 123 stress states.

[16] Local stress states are then considered according to the attitudes of stress axes in order to reconstruct successive paleostress fields under which brittle deformation occurred (only approximately since amplitudes of block rotations around vertical axis are unknown in the DF). The relative ages of local stress states in sites exhibiting polyphase stress histories was established using classical criteria such as successive striations on the same fault plane, crosscutting relationships between fractures, determination of pre- or post- folding stress states (assuming that one of the principal stress axis is vertical—parallel to lithostatic pressure—when faulting occurred), and the age of affected stratigraphic units.

[17] The measured local stress state may reflect a combination of effects related to development of major structures as well as to the prevailing regional (tectonic) paleostress field. The former can be considered as “internal” stresses (as in a fold limbs during fold formation, or also along strike-slip fault zones). They can differ from the regional stress field not only in terms of stress trajectories but also by stress regime. Accordingly, the reconstructed paleostress stress states are presented in subsequent sections in the framework of their respective structural settings.

3.2. Southwestern Margin of the Ukrainian Donbas (Figure 2)

[18] The southwestern margin of the DF (and more the Ukrainian shield and the Azof Massif) was strongly uplifted in Permain times and, consequently, the underlying Proterozoic crystalline rocks and the Devonian to Lower Carboniferous syn-rift succession is exposed, giving access to the earliest history of basin formation. The Devonian to Lower Carboniferous stratigraphic succession in this area is characterized by extrusive rocks and continental clastics with intercalated volcanoclastic units, followed by a thick carbonate platform sequence (Figure 1d). The characteristics of all local stress states are listed in Table 1.

3.2.1. Observed Paleostress Events

[19] The oldest paleostress state observed in rocks of the DF southwestern margin is a tensional stress field with $\sigma_3$ trending NNE-SSW, recorded at nine sites in the Azov Massif and in Devonian-Lower Carboniferous rocks from normal faults and a very dense network of tension gashes (Figure 2a). A block structure characterizes the area with sedimentary layers dipping 10°-20° northward or eastward. Tilting is related to normal movements along WNW-ESE trending faults such as the Yujni Fault [cf. McCann et al., 2003]. The NE-SW trending major fault zones (Figure 2) can be interpreted as being related to fault block development on the rift margin during the initial rifting phase. At site 9, the stress tensor reveals a transessional regime along a set of NW-SE to NNW-SSE trending faults (see stereoplot on Figure 2a). Nevertheless, the general extension axis of the stress field trends perpendicular to the main WNW-ESE rift-related faults (Vassiliev, Volnovakha or Yujni Faults, see Figure 2), which might therefore have acted as purely normal faults (i.e., without oblique movements). This paleostress state likely corresponds to the Late Devonian stretching phase that initiated rift basin formation and to its reactivation in the Visean. Tournaisian-early Visean rocks are affected by this event and syn-sedimentary normal faults have been found in a late early Visean unit with related thickening of beds (site 105).

[20] A paleostress state with a NW-SE trending compressional axis was identified at 11 sites (Figure 2b). Strike-slip as well as reverse faults developed under this stress field and many stylolitic peaks display the NW-SE $\sigma_1$ axis. At eight sites, the $\sigma_1$ axis strikes E-W (Figure 2b, bottom left corner). No relative chronology could be established between these two compressional trends. However, numer-
Table 1. Characteristics of Local Stress States Corresponding to the Stress Fields Presented on Figure 2a

<table>
<thead>
<tr>
<th>Site Number/Localities</th>
<th>Age of Rocks and Lithologies</th>
<th>N-S to NE-SW Extension on Figure 2a</th>
<th>Strike-Slip Regime on Figure 2b</th>
<th>Compressive Regime on Figure 2c</th>
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<tbody>
<tr>
<td>Site Number/Localities</td>
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<tr>
<td>9/South of Razdolnoe Vassilievka fault zone</td>
<td>Archean granite/Devonian volcanic rocks</td>
<td>NW-SE trending dykes, tension</td>
<td>Syn-tight normal faulting</td>
<td>NW-SE dextral en echelon tension</td>
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<tr>
<td>10/Dalniy Quarry</td>
<td>Tournaisian-Visean limestones</td>
<td>tension gashes and joints</td>
<td>Conjugate system of shear</td>
<td>tension gashes</td>
</tr>
<tr>
<td>11/Nikolaevka Village</td>
<td>Archean granites/Devonian deposits</td>
<td>nearly E-W trending dykes and</td>
<td>Conjugate system of shear</td>
<td>associated tension</td>
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<tr>
<td>39/Maf Khaya -Styla</td>
<td>Sandstones (Visian?)</td>
<td>normal faults</td>
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<tr>
<td>41/Styla Lake</td>
<td>Visean limestones</td>
<td></td>
<td>Conjugate system of shear</td>
<td>associated tension</td>
</tr>
<tr>
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<td>conjugate sets of en echelon</td>
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<td>109/Sukhaya Volnovakha</td>
<td>Volcanic plug</td>
<td>associated tension gashes and</td>
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<td>Site Number/Localities</td>
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<td>tension gashes</td>
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<td>Archean granites/Devonian deposits</td>
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<td>tension gashes</td>
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<tr>
<td>39/Maf Khaya -Styla</td>
<td>Sandstones (Visian?)</td>
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**Table 1**: Characteristics of Local Stress States Corresponding to the Stress Fields Presented on Figure 2a
Table 1. (continued)

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<th>( \sigma_2 )</th>
<th>( \sigma_3 )</th>
<th>( \alpha )</th>
<th>% RUP</th>
<th>Q</th>
<th>Ch.</th>
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<td>NW-SE dextral en echelon tension gashes plus E-W striking stylolitic planes</td>
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<td>210 69</td>
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<td>180 12</td>
<td>272 07</td>
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<td>167 28</td>
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<td>N-S striking fractures dipping 60° W</td>
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* N, number of fault slip data to compute the stress tensor (using inversion method “INVD”) [Angelier, 1990]. Dir. and Pl., trends and plunges of principal stress axes in degrees; \( \phi = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3) \); \( \alpha \), average angle between observed slip and computed shear, in degrees (acceptable with \( \alpha < 22.5^\circ \)). RUP, criterion of quality ranging from 0% (calculated shear stress parallel to actual striae with the same sense and maximum shear stress) to 200% (calculated shear stress maximum, parallel to actual striae but opposite in sense), acceptable results with RUP < 75%. Q, quality as 3, high quality; 2, reliable; 1, poor quality. Ch., chronology between local stress events (in polyphase sites); asterisk signifies a back-tilted stress event.

*1 = 31/Razdolnoe Dolginskaya suite- Famennian N-S striking fractures
*10 = Dalniy Quarry Tournaisian- Visean limestones N-S trending tension
*9 = Razdolnoe Vassilievka Archean granite/Devonian
*69 = Komcomolskoe Visean limestones 2 NE-SW tension gashes
*68 = Zhogolevsky quarry Tournaisian Visean

3.2.2. Chronology and Summary of Paleostress Stress Trends on the Donbas Southwestern Margin (Figure 3)

[23] The succession of paleostress fields described above was established from field observations as follows. At site 9, successive grooves developed on the same fault plane: the first under a NNE-SSW extension and the second under a strike-slip regime with NW-SE \( \sigma_1 \) axis. At site 11, the attitude of normal faults is clearly pre-tilting whereas all other recorded stress events are post-tilting. At site 68, reverse faults that developed under the NE-SW compression crosscut tear faults that developed under the strike-slip regime with NW-SE \( \sigma_1 \) axis. At site 69, NE-SW directed tension gashes cut across the strike-slip fault planes related to the strike-slip paleostress regime with NW-SE trending \( \sigma_1 \) axis. At site 108, one set of tension gashes developed before tilting under the NNE-SSW extensional trend.

[24] Thus, two stress fields are well recorded on the southwestern DF margin and they undoubtedly both represent important tectonic events. The first is the extensional stress field affecting rocks of Devonian to early Visean age (in particular, the competent Tournaissian-early Visean limestones) and presumably related to intracratonic rifting during which the DDB initially developed. The second major tectonic phase is characterized by a strike-slip paleostress field with a NW-SE trending compressional axis (deviated to E-W between active fault zones). Strike-slip faults as well as reverse faults developed, suggesting that the paleostress field was a transpressive one.

3.3. Russian (Eastern) Part of the Donbas (Figure 4)

[25] In the easternmost, Russian part of the DF, the Main Anticline structure attenuates somewhat, with limbs dipping 45° – 60°. The Konstantinovsky Fault zone lies along its hinge and is a right-lateral strike-slip fault according to associated fault patterns at its western termination (NNE-SSW striking normal fault pattern north of its trace and E-W striking reverse fault pattern south of it; Figures 4b and 4d). The northern and southern margins are respectively characterized by northward directed thrusts (the Almazny Fault zone) and southward directed thrusts (the Persianovsky Fault zone [Pogrebnov et al., 1985]; Figure 4).

[26] Kinematic observations of brittle structures have been made in Carboniferous rocks as well as Cretaceous rocks (sites 20, 61, 65). Analysis of 900 brittle structures (including many tension gash sets) has allowed the infer-
Figure 2. Paleostress field succession in the southwestern zone of the Donbas fold belt recorded in Proterozoic crystalline rocks, Devonian and Early Carboniferous volcanic and sedimentary succession. (a) Extensional stress trends relative to the rifting event. (b) A strike-slip and compressive regime with NW-SE trending $\sigma_1$ and E-W $\sigma_3$ stress axis deviation. (c) Compressional and strike-slip regime with NE-SW trending $\sigma_1$. (d) Some E-W directed tensional stress axis trends along the E-W striking major fault zones. Key for stereoplots: Schmidt’s projection, lower hemisphere; bedding planes as broken lines, fault planes as thin lines, striae as small arrows (inward directed = reverse, outward directed = normal, couple of thin arrows = strike-slip); computed stress axes as 5-, 4- and 3- branch stars ($\sigma_1$, $\sigma_2$ and $\sigma_3$, respectively); direction of compression: inward directed large arrows, direction of extension: outward directed large arrows. As background: extract of geological map of the Ukrainian Donbas fold belt [Donetsk State Regional Geological Survey, 1995].
ence of four successive paleostress fields based on differing directional trend, described below from the oldest to the youngest tectonic events. The characteristics of all local stress states are listed in Table 2.

3.3.1. Observed Paleostress Events

[27] Numerous tension gashes have been developed at 4 sites in Middle and Upper Carboniferous rocks under a stress regime with a NW-SE trending $\sigma_3$ axis (Figure 4a). At site 67, a set of NE-SW orientated tension gashes is also related to right-lateral strike-slip movement along a NNE-SSW to N-S striking main fault zone (cf. stereoplot Figure 4a; NE-SW trending stylolites and associated planes are also observed near the fault).

[28] Inversion of tension gashes and strike-slip faults measured in Middle and Upper Carboniferous sediments at six sites gives a strike-slip regime characterized by a N-S trending $\sigma_3$ axis (Figure 4b).

[29] Six sites have recorded, through the development of strike-slip fault systems and related tension gashes, the occurrence of a strike-slip regime with a NE-SW trending $\sigma_3$ axis (Figure 4c). This paleostress event affected Middle and Upper Carboniferous strata.

[30] The most unambiguously recorded paleostress field was seen at 17 sites and is characterized by a N-S to NE-SW trending $\sigma_1$ axis (Figure 4d). Reverse faults as well as strike-slip faults developed under this regime. The strike-slip stress field is also indicated by the synchronous development of N-S trending tension gashes and N-S striking stylolitic peaks (as at sites 51, 59 and 64). In contrast to the three older inferred paleostress fields, this stress event also affected Upper Cretaceous rocks (sites 61, 65).

3.3.2. Chronology and Structural Importance of the Reconstructed Paleostress Stress Trends in the Russian Donbas (Figure 5)

[31] The relative ages of local stress states, therefore between paleostress fields, is summarized in Figure 5. Pre- and post-tilting events were recognized at sites 50 and 51. Other relative ages were inferred from crosscutting relationships between brittle structures and successive striations on the same fault mirror. Therefore, from one site to another, it was possible to give a well-constrained chronology between the determined stress fields as presented in the previous paragraphs, except between the two intermediate stress fields (the strike-slip regimes with NE-SW $\sigma_3$ and with N-S $\sigma_3$).

[32] The youngest identified stress event occurred after Late Cretaceous times whereas the three others evidently occurred prior to the Late Cretaceous times but after the Late Carboniferous. Moreover, the three older events (Figures 4a, 4b, and 4c) mainly developed tensional features in contrast to the youngest (Figure 4d) that developed compressive structures such as reverse fault systems.

[33] Significantly, the last stress event is fully compatible with fold trends observed in the area. The trend of the shortening axis responsible for the fold development is systematically parallel to the maximum principal reconstructed stress axis trend. For instance, at several sites, the $\sigma_3$ stress axis has been found to be parallel to the dip direction of the tilted bedding planes. At site 65, in Upper Cretaceous chalks, the reconstructed N-S trending $\sigma_1$ stress axis is perpendicular to the spectacular localized close anticline axis (that could correspond to an anticline formed above a thrust plane). Moreover, at site 50, strike-slip stress

**Figure 3.** Chronology between local stress states in the southwestern margin of the Donbas (corresponding to paleostress field succession on Figure 2). Caption for stress axis trends (arrows and triangles) as for Figure 2.
states occurred before and after the tilting of bedding planes and were followed immediately by a purely compressional stress event.

Accordingly, it is inferred that the youngest stress event is responsible for the inversion of initially normal faults and that it formed the compressive structures (such as localized folds and associated thrust planes) at a time no older than the end of the Late Cretaceous.

A final remark concerns the right-lateral strike-slip movement that occurred along the Konstantinovsky Fault (as indicated by the observed associated fault pattern described above). Two of the inferred paleostress fields could have generated the observed displacement and structural pattern, given their principal stress axes orientations relative to the fault orientation: the strike-slip regime with a NE-SW trending \( \sigma_3 \) axis (Figure 4c) and the N-S to NE-SW compressional regime (Figure 4d). These are the two best documented paleostress fields in terms of the measured number of related brittle structures and the number of affected sites. No major structures seem to have been developed under the effects of the two other inferred paleostress fields (NW-SE trending \( \sigma_3 \) axis, Figure 4a; N-S trending \( \sigma_3 \) axis, Figure 4b). These two stress fields are likely indicative of relatively minor tectonic events, leading only to the reactivation of inherited major fault zones.

3.4. Northern Zone of the Ukrainian Donbas (Figure 6): Record of a N-S to NW-SE Compression as the Youngest Tectonic Event

The northern zone of the Ukrainian DF corresponds to the inverted northern margin of the initial rift basin. The traces of the NW-SE striking major fault zones in this area (Svero-Donestky, Mariievsky and Almazny faults) present a lenticular shape (Figure 6). The area is characterized by localized close folds developed upon shallow thrust planes. The inversion of the northern margin has affected Upper Cretaceous strata.

Most observations were of brittle structures in Carboniferous rocks but measurements were also taken from Permian (site 91) and Cretaceous aged strata (site 95). The 13 reconstructed local stress states (listed in Table 3) can be grouped to form one stress field (Figure 6) related to folding and thrusting processes along the northern margin of the basin. For instance, at seven sites (sites 86, 96, 97, 103, 120,
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<th>Pl.</th>
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<th>Pl.</th>
<th>ϕ</th>
<th>α</th>
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<th>Q</th>
<th>Ch.</th>
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###Strike-Slip Regime with NW-SE Extensional Trend as on Figure 4a

###Strike-Slip Regime with a N-S Extensional Trend as on Figure 4b

###NE-SW Extensional Trend in Strike-Slip Regime as on Figure 4c

###N-S to NE-SW Compression as on Figure 4d

**Notes:**
- NNW-SSE and NWW-ESE tension gashes
- associated tension gashes
- tension gashes trending N120 to N150
- NNW-SSE to NW-SE tension gashes and fibers
- tension gashes and fibers
- reverse fault planes: trace of NE-SW compression
- associated jointing
- pre- and post tilting left lateral strike-slip faults in NNE-SSW comp
- associated tension gashes and stylolites
- stylolites and reverse slips parallel bedding planes: N-S compression
- N-S striking stylolites
- left-lateral strike-slip faults and tension gashes: NNW-SSE comp. axis; pre-tilting stylolites, shear joints and post tilting stylolites: N-S comp. axis
- tension gashes and stylolites: N-S comp. axis

**Comments:**
- NE-SW striking tension gashes
- NE-SW tension gashes
- right-lateral strike-slip faults and associated stylolites and tension gashes
- tension gashes trending N090
- shear joints developed in strike-slip event with E-W comp.axis
- associated tension gashes
- associated tension gashes
- tension gashes and fibers
- reverse fault planes: trace of NE-SW compression
- associated jointing
- pre- and post tilting left lateral strike-slip faults in NNE-SSW comp
- associated tension gashes and stylolites
- stylolites and reverse slips parallel bedding planes: N-S compression
- N-S striking stylolites
- left-lateral strike-slip faults and tension gashes: NNW-SSE comp. axis; pre-tilting stylolites, shear joints and post tilting stylolites: N-S comp. axis
- tension gashes and stylolites: N-S comp. axis
129, 130), the maximum principal stress axis (i.e., the pressure axis) is parallel to the direction of shortening.

It is interesting to note that a deviation of the principal stress axis trajectory occurred in this area. Close to the NW-SE trending northern limit of the basin (Figure 6), $\sigma_1$ trends N-S whereas inside the basin, it trends NW-SE. Moreover, toward the center of the basin, the strike of the thrust planes becomes perpendicular to the $\sigma_1$ axis (sites 86, 120). The thrust planes were reformed under this NW-SE compression. The stress trends highlight the right-lateral movement that occurred along the northern margin during its inversion, also producing the lenticular shape of the traces of the fault pattern.

It also suggests that stress trends recorded within the sedimentary basin are different than the regional stress trends (at the plate tectonic scale). Such a phenomenon could be explained by a rheological contrast between the crystalline host and sedimentary infill of the basin. In the present case, a N-S compressive trend could have been the trace of the regional trend. At polyphase site 120, a pre-tilting strike-slip stress tensor was inferred. It could be the legacy of the N-S compressive stress field that affected the basin just prior to folding and thrusting (that occurred under NW-SE directed compression in this zone). Alternatively, the right-lateral reverse movement along the northern major boundary faults may have produced the deviation of stresses with “secondary” thrust planes newly forming inside the basin (such as the Nikanorovsky Fault, between sites 86 and 120 on Figure 6). A deviation of $\sigma_1$ stress axis trajectory along the Almazny Fault zone, from sites 91, 129 and 101 to sites 127 and 121 (Figure 6), is also noted: the trajectory is N-S north of the Almazny fault and E-W to the south.

3.5. Paleostress Trends in the Main Anticline Zone of the Donbas (Figure 7)

The records of three different stress fields have been recognized in the vicinity of the MA of the DF. The characteristics of all local stress states are listed in Table 4.

### 3.5.1. Observed Paleostress Events

The oldest inferred stress field corresponds to the one directly related to the active development of the MA (Figure 7a). Brittle structures observed at 12 sites are clearly associated with the anticline development (Figure 8). They include sets of tension gashes (perpendicular to the bedding planes with trends parallel to the fold axis - sites 80, 81, 115, 110, 111 - and perpendicular to the fold axis - sites 115, 119), conjugate systems of strike-slip faults, and shear joints developed on the limbs of the MA (sites 87, 134).

The second paleostress field that has been recognized on the MA is well documented at 6 sites (Figure 7b) and corresponds to the regional tectonic event affecting Upper Cretaceous strata in the DF (as well illustrated on Figure 6). Under this action of this NW-SE compression, the MA fault zones (located at the hinge of the anticline) acted as right-lateral strike-slip faults (Figure 7b).

The last event is poorly recorded (at 4 sites along the MA; Figure 7c) and indicates a strike-slip structural regime with an E-W trending $\sigma_1$ axis. Any tectonic interpretation
The relative ages of the first (formation of the MA; Figure 7a) and third mentioned paleostress fields (Figure 7c) is clearly demonstrated. The younger stress field (strike-slip regime with an E-W trending \( \sigma_1 \) axis) is clearly present both pre- and post-development of a local brachyanticline that developed locally along the MA (e.g., site 87). Elsewhere (specifically, in the DF-DDB transition zone northwest of the MA), brachyanticlines can be seen to have developed as a result of salt tectonics that deformed the sedimentary succession up to and including Upper Cretaceous strata [Balukhovsky, 1959a, 1959b; Popov, 1963]. Paleostress indicators observed in this area (Figure 9) include a conjugate system of shear joints in Upper Cretaceous chalks (sites 24 and 25) and in Permian strata (site 26) as well as a very dense network of N20–N30 striking vertical joints (10–15 cm space interval) in an Upper Cretaceous chalk quarry at site 117. These brittle structures developed during active salt anticline growth during the Eo-Alpine tectonic period [Stovba and Stephenson, 1999]. It follows that similar structures related to the poorly recorded last stress event in the MA may have formed analogously.

### 3.5.2. Chronology Between the Stress Fields in the Main Antcline Zone

At site 87, the relative ages of the first (formation of the MA; Figure 7a) and third mentioned paleostress fields (Figure 7c) is clearly demonstrated. The younger stress field (strike-slip regime with an E-W trending \( \sigma_1 \) axis) is clearly present both pre- and post-development of a local brachyanticline that developed locally along the MA (e.g., site 87). Elsewhere (specifically, in the DF-DDB transition zone northwest of the MA), brachyanticlines can be seen to have developed as a result of salt tectonics that deformed the sedimentary succession up to and including Upper Cretaceous strata [Balukhovsky, 1959a, 1959b; Popov, 1963]. Paleostress indicators observed in this area (Figure 9) include a conjugate system of shear joints in Upper Cretaceous chalks (sites 24 and 25) and in Permian strata (site 26) as well as a very dense network of N20–N30 striking vertical joints (10–15 cm space interval) in an Upper Cretaceous chalk quarry at site 117. These brittle structures developed during active salt anticline growth during the Eo-Alpine tectonic period [Stovba and Stephenson, 1999]. It follows that similar structures related to the poorly recorded last stress event in the MA may have formed analogously.

### Table: Site Numbers and Successive Stress States

<table>
<thead>
<tr>
<th>Site Number</th>
<th>Successive Stress States</th>
<th>Compression Event with N-S to NE-SW Trending ( \sigma_1 )</th>
<th>Strike-Slip Regime with N-S Trending ( \sigma_1 )</th>
<th>Strike-Slip Regime with N-S Trending ( \sigma_1 )</th>
<th>NW-SE Trending ( \sigma_2 ) in a Strike-Slip Regime</th>
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</tr>
</tbody>
</table>

### Figure 5
Chronology between local stress states in the eastern ending of the Donbas (corresponding to paleostress field succession in Figure 4). Caption for stress axis trends (arrows and triangles) as for Figure 2.

### Figure 6
Paleostress trends in the northern zone of the Donbas: the record of a N-S to NW-SE compression as the last tectonic event. Caption for stress axis trends (arrows and triangles) and keys for stereoplots as for Figure 2. As background: Extract of geological and structural map of the Donbas [Pogrebno et al., 1985].
### Table 3. Characteristics of Local Stress States Corresponding to the Stress Fields Presented on Figure 6a

<table>
<thead>
<tr>
<th>Site Number/Localities</th>
<th>Age of Rocks and Lithologies</th>
<th>N</th>
<th>$\sigma_1$</th>
<th>$\sigma_2$</th>
<th>$\sigma_3$</th>
<th>$\phi$</th>
<th>$\alpha$</th>
<th>% RUP</th>
<th>Q</th>
<th>Ch.</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>86/Svetlodarsk</td>
<td>Upper Carb. limestones</td>
<td>32</td>
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<td>02</td>
<td>240</td>
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<td>9</td>
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<td>Campanian limestines</td>
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<td>10</td>
<td>067</td>
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<td>287</td>
<td>74</td>
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<td>7</td>
<td>29</td>
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<tr>
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<td>15</td>
<td>37</td>
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<tr>
<td>101/Pervomaisk</td>
<td>Upper Carb. limestones</td>
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<td>279</td>
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<td>135</td>
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<td>Bashkirian limestones</td>
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<td>25</td>
<td>0.6</td>
<td>4</td>
<td>26</td>
</tr>
<tr>
<td>121/Guardinski</td>
<td>Bashkirian - Moscovian sandstones</td>
<td>18</td>
<td>138</td>
<td>00</td>
<td>228</td>
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<td>71</td>
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</tr>
<tr>
<td>130/Yurevka</td>
<td>Bashk. Mosc. sandstones</td>
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<td>090</td>
<td>73</td>
<td>190</td>
<td>03</td>
<td>281</td>
<td>16</td>
<td>0.7</td>
<td>6</td>
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</table>

*Definitions, abbreviations, and sources as for Table 1.

### Table 4. Characteristics of Local Stress States Corresponding to the Stress Fields Presented on Figure 7a

<table>
<thead>
<tr>
<th>Site Number/Localities</th>
<th>Age of Rocks and Lithologies</th>
<th>$\sigma_1$</th>
<th>$\sigma_2$</th>
<th>$\sigma_3$</th>
<th>$\phi$</th>
<th>$\alpha$</th>
<th>% RUP</th>
<th>Q</th>
<th>Ch.</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
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<td>339</td>
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<tr>
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<td>Bashkirian sandstones</td>
<td>6</td>
<td>090</td>
<td>73</td>
<td>190</td>
<td>03</td>
<td>281</td>
<td>16</td>
<td>0.7</td>
<td>6</td>
</tr>
<tr>
<td>7/Mius River</td>
<td>Moscovian-Kasimovian sandstones</td>
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<td>006</td>
<td>19</td>
<td>243</td>
<td>58</td>
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<td>25</td>
<td>0.6</td>
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<tr>
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<td>Bashkirian sanstones</td>
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<td>325</td>
<td>12</td>
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<td>7</td>
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<tr>
<td>81/Okhvotka</td>
<td>Bashkirian sanstones</td>
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<tr>
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<td>Bashkirian sandstones</td>
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<td>02</td>
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<td>85</td>
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<td>05</td>
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</tr>
<tr>
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<td>Moscovian siltsstones</td>
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<td>090</td>
<td>73</td>
<td>190</td>
<td>03</td>
<td>281</td>
<td>16</td>
<td>0.7</td>
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<td>Moscovian sandstones</td>
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<td>119/Arendreeva</td>
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<td>125</td>
<td>77</td>
<td>0.5</td>
<td>7</td>
</tr>
<tr>
<td>120/Malo Ivanovka</td>
<td>Bashkirian limestones</td>
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<td>296</td>
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</tr>
<tr>
<td>129/K. Popasnaya</td>
<td>Bashkirian sandstones</td>
<td>7</td>
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<td>85</td>
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<tr>
<td>130/Yurevka</td>
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<td>339</td>
<td>84</td>
<td>0.6</td>
<td>9</td>
</tr>
</tbody>
</table>

*Definitions, abbreviations, and sources as for Table 1.
brachyanticline at this site (pre- and post-tilting stress tensors; see stereoplots on Figure 7c).

At site 120 (presented on Figure 6), the pre-tilting stress tensor could correspond to the stress field related to MA formation (Figure 7a) with the post-tilting tensor belonging to the last compressional event (Figure 7b).

No chronology exists between the NW-SE compression (Figure 7b) and the strike-slip field with the E-W pressure axis (Figure 7c).

3.6. Southern Zone of the Donbas (Figure 10)

The southern zone of the DF is characterized by two gentle WNW-ESE folds (syncline and anticline). Twenty-five stress states (with characteristics listed in Table 5) were documented in this area, corresponding to three separate paleostress fields.

3.6.1. Observed Paleostress Events (Figure 10)

The oldest paleostress field in this area is a strike-slip regime with an E-W trending $\sigma_1$ axis, constrained at five sites (Figure 10a). It affected (Serpukhovian and Bashkirian) Carboniferous rocks but not Upper Cretaceous cover. This stress field is unique and difficult to interpret because the trend of the maximum principal stress axis is nearly parallel to the trends of the general WNW-ESE structural grain of the area (fold axes and major fault zones–Mushketovsky and Prodolniy faults). Nevertheless, in this area, faults trending nearly N-S and dipping eastward (Meridionalny, Dulinsky, and Markovsky faults) or westward (Ilovaisky and Frantsusky faults) could have moved as reverse faults under such a stress field.

Middle Carboniferous as well as Upper Cretaceous sediments both record a strike-slip regime with a NW-SE trending $\sigma_1$ axis (Figure 10b) at 10 sites. At sites 71 and 47, the local calculated stress tensors clearly show a transpressive regime with $\psi$ ratio equal to 0.1. It is inferred that, under this stress field, the major WNW-ESE trending faults in this area were oblique thrusts and that the N-S trending faults were left-lateral strike-slip faults (as micro-faults at site 47, where both fault trends are present). In Upper Cretaceous strata, this paleostress regime clearly generated thrust planes at small as well as at large scales, accompanied by folding.

At seven sites, a strike-slip to compressive paleostress regime with a nearly N-S trending $\sigma_1$ axis is indicated (Figure 10c). Thrusts (and, locally, folds) developed in Upper Cretaceous rocks under this stress regime, as under the previous one.

3.6.2. Chronology of Paleostress Fields in the Southern Zone of the Donbas

The first strike-slip regime (with E-W $\sigma_1$; Figure 10a) is characterized by back-rotated local stress states at polyphase sites (15, 22, 28, 36), whereas the other stress states are post-tilting of bedding planes at the same sites (Figures 10b, 10c, and Table 5). Moreover, in this zone, this event affected only Carboniferous and not Upper Cretaceous rocks whereas the last two events both developed thrusts within Upper Cretaceous strata. At site 14, the NW-SE transpression occurred prior to the tilting of beds whereas the observed NNE-SSW compression occurred after tilting. Both regimes can, nevertheless, be considered as part of the same regional tectonic phase that
produced thrusts and folds at the end of the Cretaceous in the DF.


The paleostress analyses for each of the separate zones of the DF can be extrapolated to determine the paleostress history of the DF as a whole. The main problem is to fix the absolute chronology of the various documented stress fields. The general lack of Triassic-Jurassic-Lower Cretaceous strata outcropping in the DF itself, means that a complete stratigraphic control on the timing of tectonic events is not possible. Nevertheless, several hypotheses concerning the stress history of the DF can be made based on the presented paleostress results and data from the published literature.

Figure 11 (synthesis of the more detailed Figures 2, 4, 6, 7, 9, and 10) summarizes the successive paleostress fields that have been reconstructed in the various structural zones of the DF. In each zone, the relative chronology between the successive stress regimes has been established (section 3) with the exception of the two youngest events observed in the western MA zone (Figures 7b and 7c) and the two intermediate ones in the eastern part of the DF (Figures 4b and 4c). Figure 12 illustrates the trends of successive stress fields.

The extensional stress field (Figure 11) inferred to be related to the initial rifting phase of basin development is only recorded on the southern DF margin where it is well-documented in exposed Devonian and Lower Carboniferous rocks (Figure 2a). Because the younger sedimentary succession was not affected by this event, there is little doubt concerning its relationship with the earliest tectonism of the basin development (e.g., intracratonic rifting). The minimum stress axis lies NNE-SSW perpendicular to the rift axis (as also inferred by Korchemagin and Yemets [1987]; Figure 12a). Under this NNE-SSW extension, the main WNW-ESE rift faults might have acted as purely normal faults whereas transtensional deformation could have occurred along the major NE-SW oblique faults of the southwestern margin of the DF (cf. Figure 2).

The second stress regime, recognized in two zones, is a strike-slip regime with a NW-SE trending $\sigma_3$ axis...
(Figures 11 and 12b). It has been argued that this regime is linked directly to growth of the MA (Figure 7a). But in the eastern part of the DF (Figure 4a), this stress event is more widely recognized and could be the legacy of a regional stress field. This event, in view of its age relative to the other stress regimes (it is the oldest one in the MA zone and in the eastern DF) and the age of the affected rocks, could correspond to an early Permian transtensional event recognized by Stovba et al. [2003]. Preliminary interpretation of new deep seismic reflection data across the MA (the first such data) indeed suggests the presence of a salt body deep in its core [Roy-Chowdhury et al., 2001].

The third set of paleostress states (Figures 11 and 12c) is formed by a group of observations that indicate strike-slip deformation with an E-W \( \sigma_1 \) trend that is parallel bedding slips. No reverse faults (as micro-tectonic objects), thereby indicating a compressional stress component in this strike-slip regime, were found during the course of the present study. Given the inferred normal component of fault displacement and according to the reconstructed paleostress orientations, it follows that the Permian “folding” phase in the DF occurred in a transtensional stress regime with a NW-SE trending \( \sigma_3 \) axis. In such a case, salt tectonics could have controlled WNW-SEE trending anticline and syncline development (including especially the MA) in the DF, just as it has in the adjacent DDB [cf. Stovba and Stephenson, 1999, 2002; Stovba et al., 2003; Saintot et al., 2003]. Preliminary interpretation of new deep seismic reflection data across the MA (the first such data) indeed suggests the presence of a salt body deep in its core [Roy-Chowdhury et al., 2001].
Figure 10. Paleostress states in the southern zone of the Donbas fold belt recorded in Carboniferous and Upper Cretaceous series. Caption for stress axis trends (arrows and triangles) and keys for stereoplots as for Figure 2. As background: Extract of geological map of the Ukrainian Donbas fold belt [Donetsk State Regional Geological Survey, 1995].
NW-SE striking $\sigma_1$ (Figure 2b and cf. section 3.2.1). In the eastern (Figure 4b) and southern (Figure 10a) zones of the DF, however, there is evidence of a well-defined single stress field of more regional significance. This could be related to Triassic-Jurassic Cimmerian tectonics, including thrusting in the easternmost DF [Popov, 1963; Konashov, 1980; Stovba and Stephenson, 1999] and Karpinsky Swell [Sobornov, 1995]. Following, this stress field can be reconstructed using the observations of the eastern and southern zones (Figures 4b and 10a) and also, those of the southwestern and the MA zones (Figures 2b and 7c). In this scheme, E-W compressive trends would not correspond to an Eo-Alpine deviation of stresses in the southwestern zone and would not be related to brachyanticline formation in the MA zone. Such a strike-slip stress regime might have concentrated left-lateral deformation on the major NW-SE striking faults and generated transpressional deformation along the margins of the DF (including the reverse component of the well known thrusting of the Cimmerian deformation).

The two youngest stress events (Figure 11) are well defined in the whole of the DF and both affected Upper Cretaceous rocks. They both display compressive as well as strike-slip stress tensors and are closely related to the last folding and thrusting stage that affected the DF. Associated anticlines are commonly developed above shallow thrust planes (e.g., site 28, in Cretaceous rocks at sites 14 and 35 on Figure 10b, at site 120 on Figure 7b, at site 68 on Figure 2c). These two stress fields clearly are reflecting Eo-Alpine tectonism in the DF. In detail, this tectonic phase can be characterized by two successive stress fields: a strike-slip-compressional regime with NW-SE trending $\sigma_1$ [cf. Korchemagin and Yemets, 1987] immediately followed by second strike-slip-compressional regime with N-S to NE-SW trending $\sigma_1$. Maps on Figures 12d and 12e show the orientations of the maximum stress axis corresponding to these two stress fields. The first of these Eo-Alpine stress fields is well documented in the MA zone and north of it (Figure 12d) but there is no evidence of the second event in either of these areas. In all other zones, both are well developed.

The present study, taking into account all reconstructed paleostress trends and geometries of major structures, supports a new tectonic model for the evolution of the Donbas fold and thrust belt. Basin development began as a result of rifting during Late Devonian-Early Carboniferous times synchronous with that observed in the DDB to the northwest. The marginal faults of the DF acted as normal faults at that time. According to conventional views, the DF area, in particular its southern margin, was uplifted in the Permian. However, in significant contradiction to conventional models, the WNW-ESE trending “folds” of this age developed within a transtensional [?] stress regime that has not been strongly recorded in the DF as a whole. The occurrence of a Cimmerian tectonic phase not been unequivocally recorded in the DF although, the strike-slip stress field with an E-W compressional axis could be an indication of Cimmerian tectonics. A major Eo-Alpine (latest Cretaceous-Palaeocene) tectonic phase consists, in detail, of two successive strike-slip-compressional stress fields. Only one of them is recorded and developed significant structures in the northern zone and along the MA of the DF. This tectonic event is responsible for widespread thrusting and folding. Both events are otherwise widely documented throughout the DF. Right-lateral movement along the northern margin of the DF and the MA fault zones is well documented and is consistent with the geometry of the Eo-Alpine tectonic stress field.

5. Relationships of Paleostress Fields in the Donbas With Geodynamics Setting

The observed DF paleostress orientations, and the tectonic events they signify, allow some geodynamic infer-
ences to be made about plate boundary processes affecting the southern part of the East European Craton (EEC) since the Late Palaeozoic. The DF, in this respect, lies in a key location as regards both Variscan/Uralian (Hercynian) and later Tethyan belt growth and evolution of the European continent.

The Donbas belongs to a paleorift system that developed in the present-day southern EEC during Late Palaeozoic times that included the Dnieper-Donets (DDB) paleorift and Peri-Caspian Basin [e.g., Zonenshain et al., 1990], as schematically illustrated in Figure 13a. It has been suggested that this system represented a failed arm of a more extensive network of intracratonic rifts within a larger craton that ultimately accommodated continental breakup and development of the present southern margin of the EEC in the Devonian [e.g., Zonenshain et al., 1990; cf. Shatsky, 1964]. Recently, various basin subsidence modeling studies of the DDB paleorift [Kusznir et al., 1996a, 1996b; Starostenko et al., 1999] and the implications of these studies as regards the intensity and character of Late Devonian magmatism [Wilson and Lyashkevich, 1996; Wilson et al., 1999] suggested that mantle plume activity played a role in Late Devonian
Figure 13. Schematic reconstructions with paleostress fields in the Donbas at the scale of the plate tectonics. (a) Geodynamical setting of the East European Craton in Middle Devonian times and PDDB rifting [after Zonenshain et al., 1990; Puchkov, 1997; Nikishin et al., 2001]. (b) Geodynamical setting of the East European Craton in Visean times and DDB rift reactivation [after Zonenshain et al., 1990; Puchkov, 1997; Brown et al., 1998; Nikishin et al., 2001]. (c) Geodynamical setting of the East European Craton in Early Permian times and transtensional environment along the DDB [after Ziegler, 1990; Puchkov, 1997; Nikishin et al., 2001]. (d) Geodynamical setting of the southern edge of the East European Platform in Late Cretaceous/Early Paleocene and tectonic inversion of the Donbas [after Philip et al., 2000; Saintot, 2000; Nikishin et al., 2001]. Couple of divergent arrows as $\sigma_3$ trend in extensional or strike-slip stress fields (from our study); couple of convergent arrows as supposed $\sigma_1$ trend in compressional or strike-slip stress fields. Abbreviations: EEC, East European Craton, PDDB, Prypiat-Dnieper-Donets Basin; Karp., Karpinsky Swell.
Late Palaeozoic EEC is very speculative. Obviously the location of this arc complex relative to the affinity in the present Central Range of the Great Caucasus example, there are remnants of a Middle to Late Devonian margin of the EEC and not to the Uralian geodynamics. For related to subduction located in the present-day southern Seravkin et al. 1989; eastward directed according to recent studies [e.g., Fokin et al., 2001]. However, Ural-related subduction at this time (Late Devonian to Early Carboniferous) was eastward directed according to recent studies [Vazeva et al., 1989; Seravkin et al., 1992; Puchkov, 1997; Chemenda et al., 1997; Brown et al., 1998, 1999]. In this context, rifting of the DDB as a response to “back arc” processes would be related to subduction located in the present-day southern margin of the EEC and not to the Uralian geodynamics. For example, there are remnants of a Middle to Late Devonian subduction related arc complex of unknown directional affinity in the present Central Range of the Great Caucasus [Belov, 1981; Milanovsky, 1991; Nikishin et al., 2001], but obviously the location of this arc complex relative to the Late Palaeozoic EEC is very speculative.

These rather poorly constrained considerations of the Late Palaeozoic geodynamic setting of the DF are summarized in Figure 13a. What is important from the present paleostress study in this context is that rifting occurred (presumably in response at least in part to an “active” driving mechanism) with an extensional axis roughly perpendicular to the inferred Rheic Ocean margin of the EEC and roughly subparallel to the developing Uralos orogenic belt.

After a period of distinct tectonic quiescence in the latest Devonian and earliest Carboniferous, tectonic extension was clearly reactivated in the late early Visean in the DDB [Stovba and Stephenson, 1999] and in the DF [this study; McCann et al., 2003]. The orientation of the extensional axis was similar to that of the main Late Devonian rifting stage (Figure 13b).

During Early Carboniferous times, according to many authors [e.g., Belov, 1981; Zonenshain et al., 1990; Adamia, 1991; Milanovsky, 1991; Ustaomer and Robertson, 1997], the “southern” margin of the EEC was part of an accretionary belt with widespread thrusting and folding (e.g., “Scythian Orogen” [Nikishin et al., 2001]). However, the compressive far-field effects of this are not in evidence in the DDB paleostress history (or in the preserved geological record of the DDB). Thus, for some reason the DF was isolated from the effects of Late Palaeozoic convergence processes involving the presumed orogenic consolidation to its south. One implication is that accretion of Scythian terranes to the EEC did not occur at this time; if it did, then it occurred in an extremely “soft” manner without the transmission of significant compressional stress into the EEC [Stephenson et al., 2001].

The “eastern” margin of the EEC in the Late Palaeozoic is much better known than the “southern” margin. In this case, compressional deformation structures began to be formed in the southern Urals at late Famennian-Early Carboniferous times with the accretion of Magnitogorsk volcanic arc [e.g., Zonenshain et al., 1990; Puchkov, 1997, Brown et al., 1998, 1999; Brown and Spadea, 1999] and the development of the Emba Branch of the Variscan Uralian orogenic belt [Puchkov, 1997]. According to Brown et al. [1998, 1999] and Brown and Spadea [1999], the arc-continent collision ended during the Tournaisian, as marked by the collapse of the accreted arc and deposition of carbonates on top of it. Thus, it appears as though the “eastern” margin of the EEC was tectonically quiescent when the Visean extensional reactivation of the DDB and Donbas rift occurred.

Further to the east, however, convergence between the EEC and Kazakhstan plates continued, with the subduction of an oceanic plate (the “Paleo-Uralian Ocean” [Puchkov, 1991, 1997]) beneath Kazakhstan (or beneath previously accreted outboard terranes at its margin [Puchkov, 1997]). Figure 13b shows a schematic reconstruction of the paleotectonic setting of the time, including a speculative location of the active subduction zone, prior to Middle Carboniferous to Late Permian-Early Triassic continental collision between the EEC and Kazakhstan [Puchkov, 1991, 1997]. Puchkov [1997] mentioned that the Kazakhstan plate rotated several degrees anti-clockwise at Visean-earl Bashkirian times, just preceding the “rigid” continental collision. The effect of this was to create a tensional regime in the southern part of the Urals, with “the formation of sedimentary and magmatic complexes atypical of collision or subduction” [Puchkov, 1997, p. 226]. Given that this tensional event in the southern Urals is contemporaneous with reactivation of extension in the interior of the EEC plate, as documented in the DDB-DF rift, it is permissible to suggest that there may be some connection. What may be important in this regard is that the residual effects of Late Devonian rifting (thermal perturbation and incompletely relaxed ambient extensional stresses) in the DDB-DF system at this time will have remained very significant. It follows that the additional imposition of even relatively small extensional stresses, such as those produced by rotation of the Kazakhstan plate relative to the EEC, could be sufficient for extensional reactivation of the DDB-DF.

In Early Permian times, the Donbas was in a trans-tensional stress regime (Figure 13c), similar to the wide-spread extensional-trans-tensional regime that characterized most of north central Europe and the Tornquist-Tessyere Zone at the same time [Ziegler, 1990]. According to Ziegler [1990], Late Carboniferous-Early Permian dextral translation between Africa(-Gondwana) and Laurasia was responsible for the development of regional wrench fault systems in which pull-apart basins formed, including the Permian basins of north central Europe [cf. van Wees et al., 2000]. This system of regional right-lateral faults defined a diffuse plate boundary between Africa and Laurasia. The widespread tectonic instability that developed in northwestern Europe during the Late Carboniferous-Early Permian is often ascribed to the “post-orogenic extensional collapse” of the Variscan Orogeny [cf. Ziegler, 1990]. However, the fact that the DF (and DDB), located within the EEC and obviously not part of the Variscan belt, appear to share...
this regional tectonic setting suggests otherwise. As such, the DF paleostress observations support the results of modeling studies that suggest that gravitational instability of the Variscan Orogeny would be insufficient in itself to generate the intensity and degree of post-Variscan extension observed in north central Europe [Henk, 2000]. Thus, the observed Early Permian state of stress of the DF developed within a widespread geodynamic setting that was related to large-scale plate boundary forces affecting much of the European lithosphere at that time.

[68] Contemporaneously, in a background of global relative sea level fall [Harland et al., 1990], the Donbas, especially its southern margin and the adjacent Azov Massif and Ukrainian Shield, was affected by a very rapid, important and absolute uplift [Stephenson et al., 2001]. Given the observed transtensional stress regime, this can be related to regional strike-slip movements at adjacent sub-plate boundaries. Unfortunately, there exists very little real constraint on what the paleotectonic situation in the region just to the (present-day) south of the DF was at this time. As such, it can only be speculated that such movements could have occurred between consolidated terranes to the south and the EEC itself, and/or between accreted terranes and the subducting Paleo-Tethys oceanic plate, and/or along suture zones between different terranes (cf. Figure 13c). In any case, the implication of such a model is that the southern boundary of the EEC could be considered as a broad transcurrent plate boundary at this time, similar as proposed by Arthaud and Matte [1977] 25 years ago from the western Variscan system (the Appalachians) to the Urals. (It is also noted that, at Wordian times, Gaetani et al. [2000] infer a right-lateral strike-slip boundary between western Europe (-Laurasia) and Africa (-Gondwana), as the continuation of the Paleo-Tethyan subduction zone the plate boundary.)

[69] Mesozoic tectonic stress regimes affecting the DF were related to the active southern (Tethyan) margin of the EEC [e.g., Stampfl et al., 1991; Dercourt et al., 1993; Ricou, 1996] since significant plate-scale tectonic activity in the southern Urals had ceased by the end of the Permian [Zonenshain et al., 1990; Nikishin et al., 1996; Puchkov, 1997]. The third stress regime observed in the DF (strike-slip; Figure 12c) is poorly constrained in age but could be a record of Triassic-Jurassic Eo-Cimmerian tectonics. Rift systems on the southern margins of the EEC have been inverted at the end of the Triassic, during the closure of the Paleo-Tethys ocean [Sengör, 1984; Nikishin et al., 2001]. According to the stress orientations determined for the Donbas, the (speculative) Eo-Cimmerian event, deformation would have been (left-lateral?) tranpressional indicating oblique convergence.

[70] The regional tectonic setting in which the youngest, tranpressional, stress regime of the DF was formed, during Eo-Alpine deformation at end of Cretaceous-beginning of Tertiary times is shown in Figure 13d. Tectonic inversion of sedimentary basins elsewhere in Europe, such as in the Polish Trough [Stephenson et al., 2003], occurred at the same time [cf. Ziegler, 1990]. The exact mechanism by which plate boundary stresses are transmitted into plate interiors to produce intraplate inversion structures remains unclear [e.g., Ziegler et al., 1998] but, in any case, it seems highly likely that a common, plate-scale, process is responsible. Nevertheless, there is a change of trajectory of paleostress axes observed in various locations throughout Europe at this time, for example, from a slightly NW to north directed main compressional axis in the DF to a more NE directed compressional axis in the Polish Trough (Figure 13d) [Lamarche et al., 1999; Philip et al., 2000]. Finally, it is perhaps noteworthy that this was also the time when the extensional East Black Sea basin was developing [Finetti et al., 1988; Robinson et al., 1996]. Surrounding areas such as the western Caucasus are characterized by related transtensional structures of this age [Saintot, 2000; Saintot and Angelier, 2002]. Interestingly, the stress axis trends of the tranpressional strike-slip stress field of the DF are identical to those of the transtensional strike-slip stress field related to East Black Sea basin opening recorded in the western Caucasus (Figure 13d) [Saintot, 2000; Saintot and Angelier, 2002].

6. Conclusion

[71] The main conclusions that can be drawn from the inferred paleostress history of the DF as reported here are: (1) The “inversion” of the DF, with concomitant shortening and development of compressional structures such as thrusts and folds can be reassigned to the Late Cretaceous. This is clearly not a Permian event as has been conventionally reported in the literature. (2) A strike-slip regime, probably transtensional in style, affected the DF during the Permian, contemporaneously with a regional sea level drop and uplift of the southern margin of the DF. (3) The Devonian and Early Carboniferous rifting phases characterizing the DF and Dnieper-Donets Basin occurred under an extensional stress regime without evidence of any strike-slip component along the WNW-ESE marginal faults of rift system. (4) The earliest set of WNW-ESE striking “folds” (the MA and open synclines and anticlines south and north) observed in the present-day structure of the DF have developed under a strike-slip regime with probable transtensional deformation, in conjunction with salt tectonics in the Permian time. (5) The second set of folds is characterized by various strikes of fold axes and is commonly associated with shallow thrusting. It corresponds to Eo-Alpine compressional and strike-slip stress fields, clearly recognized in Carboniferous and Cretaceous rocks. (6) A strike-slip regime with an E-W compressional axis can characterize a Cimmerian phase of deformation; if not, deviation of stress trends of the Eo-Alpine stress field. (7) The timing and style of tectonic events recorded by kinematic indicators in the DF permit important new constraints to be placed on the otherwise poorly known geodynamic history of the southeastern margin of Europe.

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