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Late Paleozoic Transgressions in the Greater Caucasus (Hun Superterrane, Northern Palaeotethys): Global Eustatic Control

Trangresiones del Paleozoico tardío en el Gran Caucaso (Superterreno Hun, PaleoTesis Norte)

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Abstract

The Greater Caucasus (a mountain chain in southwesternmost Russia, northern Georgia, and northwestern Azerbaijan) is a Gondwana-derived terrane, which was included into the Hun Superterrane (late Silurian-Late Devonian) and then docked at the Laurussian margin of the Palaeotethys Ocean close to the terranes of the present Alps (Late Devonian-Middle Triassic). The Upper Paleozoic sedimentary complex, up to 20,000 m thick, provides a good record to discuss global eustatic changes. Three transgressions are reported in this region, which occurred in the Lochkovian, Frasnian-Famennian, and Changhsingian. The first of them embraced the northern part of the Greater Caucasus, the second was larger and covered this region entirely, whereas the third occurred in its western part only. All of them corresponded evidently to global eustatic rises, and, therefore, their explanation does not require an implication of the regional tectonic activity. These regional transgressive episodes are also known from the Southern and Carnic Alps, Arabia, and Northern Africa. A correspondence between the Late Permian marine sedimentation in the Greater Caucasus and non-marine sedimentation in Spain is established. Thus, they were of planetary extent and the present global eustatic curve is confirmed. The regional transgressions resulted in carbonate deposition, biotic radiations, and reefal growth.

Key words: palaeogeography, Late Paleozoic, sea-level changes, eustasy, Greater Caucasus, Paleotethys, Russia

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INTRODUCTION

The global sea-level fluctuated strongly during the Late Paleozoic. Well-known earlier reconstructions of this eustasy were made by Vail et al. (1977), Johnson et al. (1985), Ross & Ross (1985), Hallam (1992), Ronov (1994), and Hallam & Wignall (1999). Recently, Haq & Al-Qahtani (2005) have proposed a new global eustatic curve. In general, it is concluded that the sea level dropped gradually during the Late Paleozoic, with a pronounced drop in the mid-Permian. Second-order fluctuations were superposed on this trend. Unfortunately, our knowledge of the Paleozoic sea-level changes remains limited (see also Miller et al. 2005). For example, the present curve of Haq & Al-Qahtani (2005) suggests a major regression in the end-Permian, although Hallam & Wignall (1999) argued for a major transgression at this time. By the same token, sea-level changes at the Frasnian-Famennian transition are not clear (Racki 2005). Verification of proposed eustatic curves as well as their details, improvement, and justification is possible only by careful comparison of numerous regional data from across the world. An example from the Jurassic demonstrates that such analysis would significantly contribute both to the identification of the global sea-level changes and their explanation (Hallam 2001). A reconstruction of the global eustatic curve is an enormously difficult task and some doubts are even expressed as to its existence (e.g., McGowran 2005). However, such a curve will be an important key to explain the changes in world palaeogeography, sedimentary environments, and biotic evolution. In the regional record, we document the global sea-level changes by the transgressive and regressive episodes. However, the latter may also reflect (and almost always do anyway!) the regional tectonic influences. Thus, our task is to differentiate between the eustatic- and tectonic-induced regional sea-level changes. Hallam (2001) made an intriguing conclusion, that transgressions are more evident when traced globally, than regressions, which reflect mostly the regional tectonic movements. In the author's opinion, this does not diminish the importance of world correlation of regressive episodes and unconformities, but emphasizes the need for global tracing of the regional transgressions.

The Greater Caucasus Terrane, presently included into the Alpine Mediterranean Belt and located in the Southwest of Eurasia (Fig. 1), provides an exceptional Late Paleozoic record. According to the present tectonic model (Tawadros et al. 2006; Ruban et al., 2007), this region was one of the Hunic terranes identified by Stampfli & Borel (2002), and therefore, its record is meanful for both the Afro-Arabian margin of Gondwana and Variscan Europe. This article is the first, which attempts to give a comprehensive, although brief synthesis of knowledge on the Late Paleozoic transgressions, which occurred in the Greater Caucasus.

GEOLOGICAL SETTING

The Greater Caucasus pertains to a large mountain chain, which is located in the Southwest of Eurasia and connected with the other mountains originated during the Alpine phase of orogenic activity. The Greater Caucasus embraces southwesternmost Russia, northern Georgia, and northwestern Azerbaijan. The Paleozoic sedimentary complexes, whose total thickness is up to 20,000 m, are known in the central part of this territory. They are exposed both in numerous little outcrops and in the continuous sections along the river valleys. Deposits of all three Upper Paleozoic systems are known in the Greater Caucasus (Fig. 1). The Devonian is dominated by volcanics and volcanoclastics, and carbonates are known in the Upper Devonian. The Mississippian is composed of shales, sandstones, and rare carbonates, whereas the Pennsylvanian is represented by non-marine coal-bearing strata. The Cisuralian-

Guadulupian is a typical molasse, whereas the mid-Permian corresponds to a major hiatus. Only in the Changhsingian, did marine sedimentation recommence, when accumulation of sandstones and shales was followed by a remarkable episode of carbonate sedimentation.

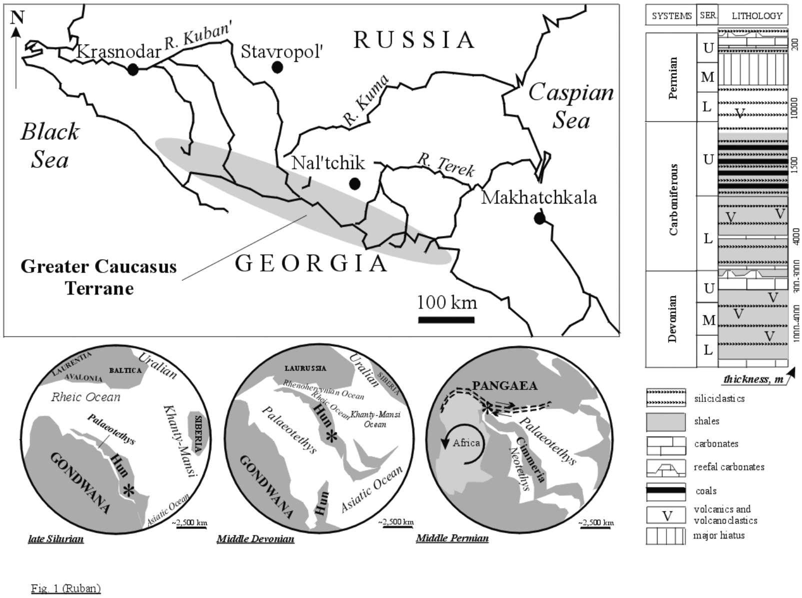

Fig. 1. Location of the Greater Caucasus Terrane and a composite lithologic section of the Paleozoic deposits exposed in the northern Greater Caucasus (after Ruban (2006) with additions). The base palaeomaps are simplified from Stampfli & Borel (2002).

The Greater Caucasus Terrane has been identified by Gamkrelidze (1997) and later by Tawadros et al. (2006) and Ruban et al. (2007). The latter authors have also developed a new model, which describes the evolution of the Greater Caucasus (Fig. 1). Until the mid-Silurian, this terrane was a part of the Afro-Arabian margin of Gondwana, i.e., it lay on the southern periphery of the Prototethys and Rheic oceans. In the Ludlow, a breakup occurred along the Gondwanan margin and a ribbon of terranes, called the Hun Superterrane, was formed (Stampfli & Borel 2002; Stampfli et al. 2002; von Raumer et al., 2002, 2003). A new ocean, i.e., the Palaeotethys, was originated in between the Hun Superterrane and Gondwana. This ocean grew in size, whereas the oceans, located between the Hun and Laurussia, were closed. Thus, the Greater Caucasus Terrane together with the other Hunic terranes moved northward. In the Late Devonian, it reached the Laurussian margin. The right-lateral displacements along the major shear zone (Arthaud & Matte 1977;

Swanson 1982; Rapalini & Vizán 1993; Stampfli & Borel 2002; Ruban & Yoshioka 2005) led to the stacking of the Greater Caucasus Terrane somewhere close to the Carnic Alps and Bohemia. In the Late Triassic, the direction of the displacements along the above-mentioned shear zone changed to left-lateral (Swanson 1982; Rapalini & Vizán 1993; Ruban & Yoshioka 2005). The Greater Caucasus Terrane then rapidly reached its present position at the south of Baltica plate.

METHOD: A CONCEPTUAL FRAMEWORK

According to Catuneanu (2006), transgression is defined as a landward migration of the shoreline. It is strongly recommended to make a distinction between transgressions and deepening pulses. Although they are linked in some cases, their true relationships are very complicated (see also Catuneanu 2006). Not in all cases does a transition to facies formed at greater depth, recorded in the sedimentary succession, mark a landward shift of the shoreline. Transgression may have a number of mechanisms, which are distinct in isolated (where are, therefore, not influenced by the global eustasy) and open or half-open basins (Fig. 3).

The Greater Caucasus Terrane was embraced during the Late Paleozoic by marine basins, directly related to the oceans extant at those times. Thus, they were open basins. Consequently, we need to examine two possible explanations of documented transgressions. If they corresponded to the global eustatic events recorded by the curve of Haq & Al-Qahtani (2005), they were eustasy-dominated. When such correspondence is not found, this means tectonic factors were more significant. Alternatively, this also may indicate inaccuracies in the global eustatic curve. The Late Paleozoic transgressions, which occurred in the Greater Caucasus, were recorded thanks to the careful analysis of data on the spatial distribution of deposits of a particular age. These data are contained particularly in the comprehensive overviews by Paffengol'ts (1959), Milukho-Maklaj & Miklukho-Maklaj (1966), Kizeval'ter & Robinson (1973), Obut et al. (1988), Kotlyar et al. (1999, 2004) and Gaetani et al. (2005). The periods, when marine facies became the most wide-spread, were the times of transgressions. Each regional transgressive episode is characterized here in a similar way, i.e., its age, area, sedimentary environments, and possible controls are considered.

A RECORD OF THE REGIONAL TRANSGRESSIONS

Three transgressive episodes may be recorded in the Late Paleozoic history of the Greater Caucasus, namely the Lochkovian, the Frasnian-Famennian, and the Changhsingian episodes.

The Lochkovian regional transgressive episode (D1-RTE) is recorded in the northern part of the Greater Caucasus (Fig. 3). Carbonates with shale interbeds of the upper member of the Manglajskaja Formation, up to 10 m thick, are known there (Obut et al. 1988). In some sections, the lowermost Devonian is represented by shales, siltstones, sandstones with carbonate lenses, whose total thickness reaches 100-150 m. The age of the above-mentioned strata is established precisely with conodonts, and these deposits also contain bivalves trilobites, and tentacultes. The Lochkovian trangsression is evident in the valley of the Malka River, where the Devonian deposits, including basal sandstones with gravels, overlie the lower Silurian strata with an evident disconformity (Obut et al. 1988). Thus, it seems that transgression was directed eastward. At the beginning of the Devonian, the Greater Caucasus Terrane together with the other Hunic terranes moved northward (Stampfli & Borel 2002; Tawadros et al. 2006; Ruban et al., 2007). The absence of the Lochkovian deposits in the southern part of the Greater Caucasus Terrane may be explained by the inclusion of the latter into a large island, which existed along the central axis of the Hun Superterrane. If so, the documented transgression occurred from the Rheic Ocean. This episode was relatively short, because already in the late Early Devonian the sea became restricted in the Greater Caucasus, and volcanoclastic deposition started. D1-RTE undoubtedly corresponded to global eustatic rise (Fig. 3). Even if any regional tectonic activity might have affected the relative sea-level, such influences were minor.

The Frasnian-Famennian regional transgressive episode (D3-RTE) is recorded in the entire Greater Caucasus (Fig. 3). This transgression started in the early Frasnian or even in the end-Givetian, when siliciclastic deposits of the Semirodnikovsksja Formation (its total thickness reaches 1,700 m) were deposited on the Early-Middle Devonian complex composed of volcanics and volcanoclastics. However, the transition between under- and overlying deposits was gradual, and the amount of conglomerates increases upwards (Kizeval'ter & Robinson 1973). These strata are overlain by the carbonates, including reefal limestones, of the Pastukhovskaja Formation, up to 3,000 m thick. The age of these strata is established as Famennian (Kizeval'ter & Robinson 1973). In the southern part of the Greater Caucasus, the upper Frasnian-Famennian Kirarskaja Formation consists of sandstones and shales with limestone interbeds. Stratigraphic relationships between the Devonian formations are not well-justified, and therefore, it becomes difficult to evaluate the direction of this transgression. In the Late Devonian, the Greater Caucasus Terrane was docked at the Laurussian margin, as well as other so-called European Hunic terranes, although a narrow remnant of the Rhenohercynian Ocean remained open between the latter and Laurussia (Stampfli & Borel 2002; Tawadros et al. 2006). It should be further investigated, whether the opening Palaeotethys or closing Rhenohercynian Ocean embraced the studied terrane. This transgression ended at the Devonian/Carboniferous boundary, because it is marked by an erosional surface (Kizeval'ter & Robinson 1973). During D3-RTE, the global eustatic level fluctuated (Fig. 3), and this regional transgression may be related to the pronounced eustatic rise, which occurred in the late Frasnian-early Famennian (Haq & Al-Qahtani 2005). It is necessary to point out the existing misinterpretation of eustatic changes at the Frasnian-Famennian boundary (Hallam & Wignall 1999; Kalvoda 2002; Racki 2005). Earlier eustatic rise in the late Givetian-early Frasnian may have initiated a marine incursion in the studied territory and deposition of conglomerates. Their long accumulation in the Frasnian may be easily explained by the remarkable global regression (Haq & Al-Qahtani 2005), which did not permit a transition to sandstones and shales or carbonates. Thus, as in the previous case, the regional transgressive episode and associated events corresponded well to the global eustasy, and implication of any regional tectonic activity is not necessary.

The Changhsingian regional transgressive episode (P3-RTE) is recorded in the western part of the Greater Caucasus (Fig. 3). This transgression started with the deposition of sandstones, shales, and carbonates of the Kutanskaja and Nikitinskaja with a total thickness exceeding 50 m (Miklukho-Maklaj & Miklukho-Maklaj 1966). But its peak was reached, when the carbonate-dominated Urushtenskaja Formatio with a thickness of more than 100 m, was formed. This formation also includes reefs. All these strata overlie unconformably the Carboniferous-Permian molasse and other older sedimentary complexes. The age of these deposits is now evaluated with brachiopods, foraminifers and other fauna as the late Chaghsingian (Kotlyar et al. 1999, 2004; Gaetani et al. 2005). However, there is some evidence for a conformable contact with the underlying deposits locally (Miklukho-Maklaj & Miklukho-Maklaj 1966). It seems that transgression occurred from the southwest, because the Early-Middle Permian marine environments were established only there. In the Late Permian, the Greater Caucasus Terrane was amalgamated with the Laurussian margin of Pangaea, and it was located somewhere close to the terranes of the present Alps (Tawadros et al. 2006; Ruban et al., 2007). The central and eastern parts of the Greater Caucasus, where the Lopingian deposits are absent, were evidently included into the continental land mass. This transgressive episode was the shortest. It was ended already by the earliest Triassic, because an unconformity is established at the base of the Triassic sedimentary complex (Miklukho-Maklaj & Miklukho-Maklaj 1966; Gaetani et al. 2005). P3-RTE corresponded to a low-amplitude, but still globally-recognizable eustatic rise (Fig. 3). Moreover, recent studies argue that this global transgression strengthened at the Permian/Triassic boundary (Hallam & Wignall 1999; Wignall 2004; Racki & Wignall 2005; Erwin 2006), which is not reflected on the curve of Haq & Al-Qahtani (2005). This well explains marine deposition until the earliest Triassic in the Greater Caucasus. The age of the Abagskaja Formation, which overlies the Urushtenskaja Formation and consists of limestones of about 25 m thickness (Miklukho-Maklaj & Miklukho-Maklaj 1966), is considered as latest Permian-earliest Triassic, because of an extremely impoverished faunistic complex (Miklukho-Maklaj & Miklukho-Maklaj 1966). This biotic crisis is explained by the devastating mass extinction occurring directly at the Permian/Triassic boundary. This gives us the age of the Abag Formation. Thus, the end-Permian regional transgression occurred from the Palaeotethys Ocean thanks to the global eustatic rise. No tectonic forces are necessary to explain this regional episode, although the origin of the Alpine-type structures in the entire Proto-Alpine Region and associated extension (Krainer 1993) might have reinforced transgression.

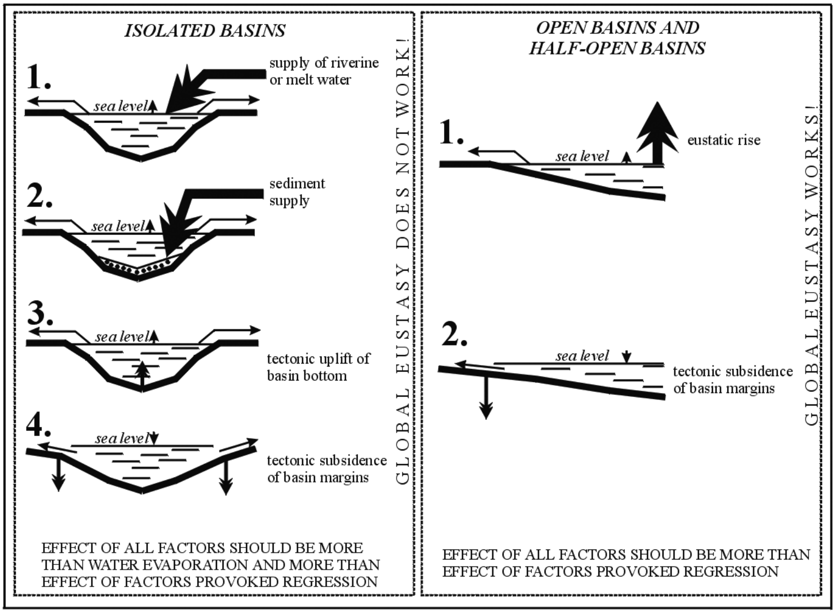

Fig. 2 (Ruban)

Fig. 2. Mechanisms of transgressions in isolated and open and half-open basins. In each particular case, the mechanism of transgression may be complicated and may have included elements from two or more idealized models.

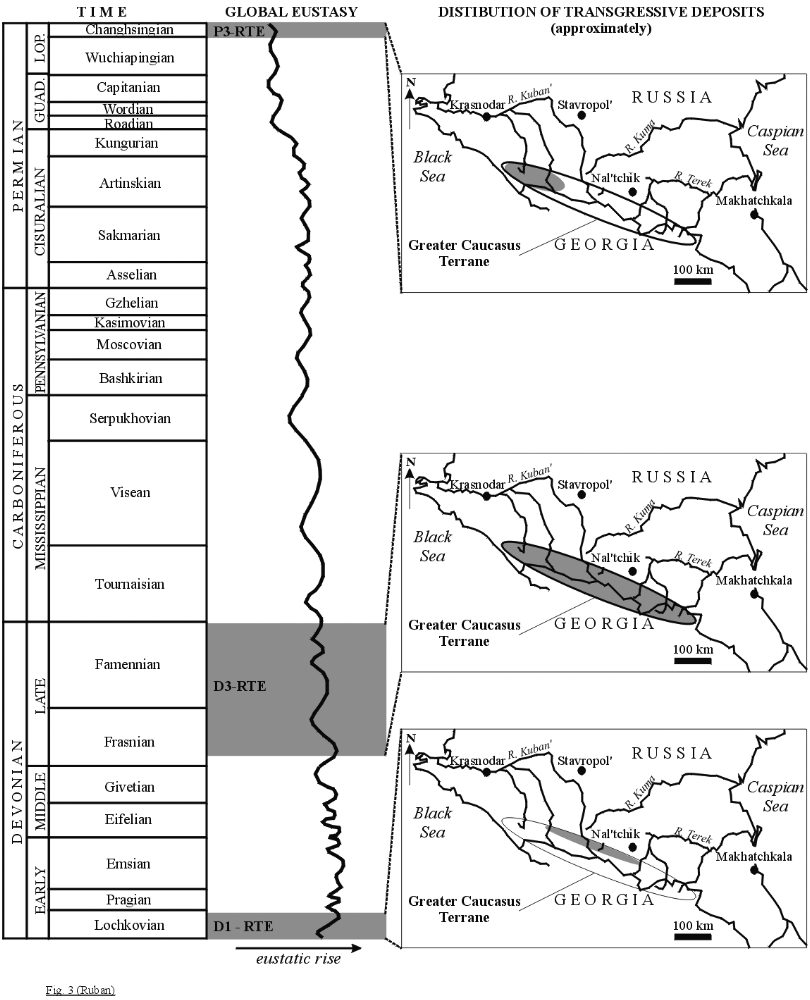

Fig. 3. The Late Paleozoic global eustatic curve (after Haq & Al-Qahtani 2005) and regional transgressive episodes (highlighted by gray). Abbreviations: Guad. - Guadalupian, Lop. - Lopingian. See text for explanation of the regional transgressive episodes.

DISCUSSION

It is important to look for analogues of the Late Paleozoic transgressions, which occurred in the Greater Caucasus. In the Austrian and Italian Southern and Carnic Alps, the Lochkovian and Lopingian transgressions were remarkable events (Krainer 1993; Schönlaub & Histon 1999; Venturini 2002). Although the evidences for Late Devonian shoreline shifts are less clear, a transgressive surface at the base of the lithostratigraphic unit named "calcari a goniatiti e climenia", whose age is late Frasnian - early Famennian, has been documented in the Carnic Alps (Venturini 2002). In Arabia, transgressions occurred in the end-Silurian, twice in the Famennian, and in the entire Lopingian (Sharland et al. 2001; Haq & Al-Qahtani 2005). In Northern Africa, the Famennian transgression is evident, whereas there was a significant regression in the Lochkovian and at the Permian/Triassic boundary; however, a transgression is known in the Lopingian (Guiraud et al. 2005). Thus, in spite of observed time differences, which may be caused by improperly understood stratigraphic framework and imperfect correlations both in the Greater Caucasus and other regions or by regional tectonic influences, we can suggest that the Late Paleozoic transgressions in the Greater Caucasus were analogous to those in other regions. This indicates their global extent, and also supports the curve of Haq & Al-Qahtani (2005).

It is very intriguing that the latest Permian transgression in the Greater Caucasus coincided with the onset of the Buntsandstein sedimentation in Spain, particularly in the Cordillera Ibérica and the Cordillera Costero-Catalana (Vera 2004). Such a correspondence can be explained by the relation of the Late Permian sedimentation (either marine or non-marine) in the Carnic Alps, Spain, and the Greater Caucasus to the beginning of extension, which embraced at least entire Southern Europe (see e.g., Krainer 1993; Stampfli & Borel 2002).

All three Late Paleozoic transgressions in the Greater Caucasus were expressed by carbonate deposition (see above). Moreover, D3-RTE and P3-RTE occurred at times of reefal growth on the periphery of carbonate platforms (Ruban 2005, 2006). In the Late Devonian, rimmed shelf was attached to the Hun island, whereas in the Lopingian, a carbonate platform of the same type was attached directly to the continental margin of Pangea. Although Khain (1962) argued for tectonic control of the Late Devonian and Late Permian reef distribution in the Greater Caucasus, we may now postulate that the appearance of these reefs might have resulted directly from the eustatically-driven transgressions. By the same token, all reported regional transgressive episodes undoubtedly coincided with biotic radiations. Available data on brachiopods suggest their Early and Late Devonian and Late Permian diversifications (Ruban 2006). The same events are also known in the regional evolution of other fossil groups like bivalves, trilobites, corals, and bryozoans (Paffengol'ts 1959; Miklukho-Maklaj & Miklukho-Maklaj 1966; Nalivkin & Kizel'vater 1973; Obut et al. 1988; Kotlyar et al. 1999, 2004). Transgressions led to the appearance of relatively shallow-marine environments on shelves, which were favorable for marine fauna. When reefal communities grew up, this accelerated biotic radiations as this was previously hypothesized by Ruban (2006).

CONCLUSIONS

Three regional transgressive episodes are known in the Late Paleozoic history of the Greater Caucasus Terrane. They were the Lochkovian, Frasnian-Famennian, and Changhsingian transgressions. Attempted comparison of them with the present global eustatic curve (Haq & Al-Qahtani 2005) and data from other regions (Alpine Europe, Northern Africa, and Arabia) suggests that all these transgressions were eustatically-controlled, and the role of regional tectonic activity to explain them seems to have been insignificant. However, the specifics in the regional tectonic evolution may explain why other transgressions documented by the global curve did not appear in the studied region. Thus, a conclusion that transgressions can be traced globally, analogous to the changes in the global sea level made by Hallam (2001) for the Jurassic, can now also be inferred for the Late Paleozoic. It is also possible to state, that the global eustatic curve of Haq & Al-Qahtani (2005) is confirmed with the data from the Greater Caucasus Terrane.

Another important conclusion from the study is that all three regional transgressions were expressed by carbonate deposition and biotic radiations, and two of them also coincided with reefal growth.

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REFERENCES

Arthaud F. & Matte P. 1977: Late Paleozoic strike-slip faulting in southern Europe and northern Africa: Result of a right-lateral shear zone between the Appalachians and the Urals. Geol. Soc. Amer. Bull. 88, 1305-1320.

Catuneanu O. 2006: Principles of Sequence Stratigraphy. Elsevier, Amsterdam.

Erwin D.H. 2006: Extinction: How Life on Earth Nearly Ended 250 Million Years Ago. Princeton Univ. Press, Princeton.

Gaetani M., Garzanti E., Pollino R., Kiricko Y., Korsakhov S., Cirilli S., Nicora A., Rettori R., Larghi C. & Bucefalo Palliani R. 2005: Stratigraphic evidence for Cimmerian events in NW Caucasus (Russia). Bull. Soc. géol. France 176 283-299.

Gamkrelidze I.P. 1997: Terranes of the Caucasus and Adjacent Areas. Bull. Georgian Acad. Sci. 155, 75-81.

Guiraud R., Bosworth W., Thierry J. & Delplanque A. 2005: Rhanerozoic geological evolution of Northern and Central Africa. J. African Earth Sci. 43, 83-143.

Hallam A. 1992: Phanerozoic Sea-Level Changes. Columbia Univ. Press, New York.

Hallam A. 2001: A review of the broad pattern of Jurassic sea-level changes and their possible causes in the light of current knowledge. Palaeogeogr., Palaeoclimatol., Palaeoecol. 167, 23-37.

Hallam A. & Wignall P.B. 1999: Mass extinctions and sea-level changes. Earth-Sci. Rev. 48, 217-250.

Haq B.U. & Al-Qahtani A.M. 2005: Phanerozoic cycles of sea-level change on the Arabian Platform. GeoArabia 10, 127-160.

Johnson J.G., Klapper G. & Sandberg C.A. 1985: Devonian eustatic fluctuations in Euramerica. Geol. Soc. Am. Bull. 96, 567-589.

Kalvoda J. 2002: Late Devonian-Early Carboniferous Foraminiferal Fauna: Zonations, Evolutionary Events, Paleobiogeography and Tectonic Implications. Folia Facultatis Scientiarium Naturalium Universitatis Masarykianae Brunensis 39, 1-213.

Khain V.E. 1962: Reefs and tectonics. – In: Obut A.M. (Ed.): A significance of biosphere in the geological processes. - Questions on interconnection between paleontology and tectonics. Gosgeoltehizadt, Moskva, 162-170 (in Russian).

Kizeval'ter D.S. & Robinson V.N. 1973: The Greater Caucasus. In: Nalivkin D.V., Rzhosnickaja M.A. & Markovskij B.P. (Eds.): Stratigraphy of the USSR. Devonian, 1. Nedra, Moskva, 220-227 (in Russian).

Kotlyar G.V., Nestell G.P., Zakharov Y.D. & Nestell M.K. 1999: Changhsingian of the Northwestern Caucasus, Southern Primorye and Southeastern Pamirs. Permophiles 35, 18-22.

Kotlyar G.V., Zakharov Y.D. & Polubotko I.V. 2004: Late Changhsingian fauna of the Northwestern Caucasus Mountains, Russia. J. Paleontol. 78, 513-527.

Krainer K. 1993: Late- and Post-Variscan Sediments of the Eastern and Southern Alps. In: von Raumer J.F. & Neubauer F. (Eds.): Pre-Mesozoic Geology in the Alps. Springer, Berlin, 537-564.

McGowran B. 2005: Biostratigraphy: Microfossils and Geological Time. Cambridge Univ. Press, Cambridge.

Miklukho-Maklaj A.D. & Miklukho-Maklaj K.V. 1966: Crimean-Caucasian alpine fold region. In: Likharev B.L. (Eds.): Stratigraphy of the USSR. Permian. Nedra, Moskva, 391-402 (in Russian).

Miller K.G., Kominz M.A., Browning J.V., Wright J.D., Mountain G.S., Katz M.E., Sugarman P.J., Cramer B.S., Christie-Blick N. & Pekar S.F. 2005: The Phanerozoic Record of Global Sea-Level Change. Science 310, 1293-1298.

Obut A.M., Morozova F.I., Moskalenko T.A. & Tchegodaev L.D. 1988: Graptolits, conodonts and stratigraphy of the Silurian, Lower Devonian of the Northern Caucasus. Nauka, Novosibirsk (in Russian).

Paffengol'ts K.N. 1959: Geological Overview of the Caucasus. Izdatel'stvo AN Armjanskoj SSR, Erevan (in Russian).

Racki G. 2005: Toward understanding Late Devonian global events: few answers, many questions. In: Over D.J., Morrow J.R. & Wignall P.B. (Eds.): Understanding Late Devonian and Permian-Triassic Biotic and Climatic Events: Towards an Integrated Approach. Elsevier, Amsterdam, 5-36.

Racki G. & Wignall P.B. 2005: Late Permian double-phased mass extinction and volcanism: an oceanographic perspective. In: Over D.J., Morrow J.R. & Wignall P.B. (Eds.): Understanding Late Devonian and Permian-Triassic Biotic and Climatic Events: Towards an Integrated Approach. Elsevier, Amsterdam, 263-297.

Rapalini A.E. & Vizán H. 1993: Evidence of Intrapangaea movements in Gondwanaland. Comptes Rendus XII ICC-P 1, 405-434.

Ronov A.B. 1994: Phamerozoic transgressions and regressions on the continents: a quantitative approach based on areas flooded by the sea and areas of marine and continental deposition. Amer. J. Sci. 294, 777-801.

Ross C.A. & Ross J.R.P. 1985: Late Paleozoic depositional sequences are synchronous and worldwide. Geology 13: 194-197.

Ruban D.A. 2005: Rises of the macrobenthos diversity and the Paleozoic – Mesozoic rimmed shelves in the northern Caucasus. The First International Scientific Conference of Young Scientists and Students: "New Directions of Investigations in Earth Sciences". Abstracts. October 3-4, 2005. Baku, 113.

Ruban D.A. 2006: Diversity Changes of the Brachiopods in the Northern Caucasus: a brief overview. Acta Geol. Hungar. 49, 57-71.

Ruban D.A. & Yoshioka S. 2005: Late Paleozoic – Early Mesozoic Tectonic Activity within the Donbass (Russian Platform). Trabajos de Geología 25, 101-104.

Ruban, D.A., Al-Husseini, M.I. & Iwasaki, Y. 2007: Review of Middle East Paleozoic Plate Tectonics. GeoArabia 12, 35-56.

Schönlaub H.P. & Histon K. 1999: The Palaeozoic Evolution of the Southern Alps. Mitt. Österr. Geol. Ges. 92, 15-34.

Sharland P.R., Archer R., Casey D.M., Davies R.B.., Hall S.H., Heward A.P., Horbury A.D. & Simmons M.D. 2001: Arabian Plate Sequence Stratigraphy. GeoArabia Spec. Publ. 2, 1-371.

Stampfli G.M. & Borel G.D. 2002: A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamics plate boundaries and restored synthetic oceanic isochrons. Earth Planet. Sci. Lett. 196, 17-33.

Stampfli G.M. von Raumer J. & Borel G.D. 2002: The Palaeozoic evolution of pre-Variscan terranes: From peri-Gondwana to the Variscan collision. In: Martinez-Catalan J.R., Hatcher R.D., Arenas R. & Diaz Garcia F. (Eds.): Variscan Appalachian Dynamics: The building of the upper Paleozoic Basement. Geol. Soc. Am. Spec. Pap. 364, 263-280.

Swanson M.T. 1982: Preliminary model for an early transform history in central Antlantic rifting. Geology 16, 317-320.

Tawadros E., Ruban D., Efendiyeva M. 2006: Evolution of NE Africa and the Greater Caucasus: Common Patterns and Petroleum Potential. The Canadian Society of Petroleum Geologists, the Canadian Society of Exploration Geophysicists, the Canadian Well Logging Society Joint Convention. May 15-18, 2006. Calgary, 531-538.

Vail P.R., Mitchum R.M.Jr. & Thompson S. 1977: Global cycles of relative changes of sea level. In: Payton C.E. (Ed.): Seismic Stratigraphy - Application to Hydrocarbon Exploration. Amer. Assoc. Petrol. Geol. Memoir 26, 83-97.

Venturini C. (Ed.) 2002: Carta geologica delle Alpi Carniche 1:25,000. S.E L.C A., Firenze.

Vera J.A. (Ed.) 2004: Geología de España. SGE-IGME, Madrid.

von Raumer J.F., Stampfli G.M., Borel G. & Bussy F. 2002: Organization of pre-Variscan basement areas at the north-Gondwanan margin. Intern. J. Earth Sci. 91, 35-52.

von Raumer J., Stampfli G.M. & Bussy F. 2003: Gondwana-derived microcontinents – the constituents of the Variscan and Alpine collisional orogens. Tectonophysics 365, 7-22.

Wignall P.B. 2004: Causes of mass extinction. In: Taylor P.D. (Ed.): Extinctions in the History of Life. Cambridge Univ. Press, Cambridge, 119-150.