GEOLOGICA CARPATHICA, 2024, 75, 2, 133–154 https://doi.org/10.31577/GeolCarp.2024.08

Upper Cenozoic conglomeratic formations as a rock record of the turning points in the evolution of the Carpathian orogen and its foreland

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(Manuscript received December 23, 2023; accepted in revised form June 12, 2024; Associate Editor: Michal Šujan)

Abstract: This paper presents the upper Cenozoic post-collisional terrestrial conglomeratic formations of the Eastern and South-Eastern Carpathian foreland subjected to sedimentological and geomorphological analyses including an in-depth review of previous investigations. These formations embrace gravels as the prominent component, together with sands and muds. The unit is up to 190 m thick. The conglomeratic formations are represented by erosional remnants in the Eastern Carpathians, whilst they occur as a continuous body (the Cândești Formation) in the South-Eastern Carpathians. The Eastern Carpathian formation ranges in age from the late Serravallian to early Tortonian stages, whilst the Cândești Formation extends from the late Early to the late Middle Pleistocene. These formations are separated by 10 Myr, supporting the view the Carpathians and their foreland from North to South evolved diachronously. Their affinity is determined by the sedimentary environment of alluvial fans within the wedge-top depozone of the foreland. This environment arose from simultaneous occurrence of rare intervals of intense floods and presence of highly erodible rocks in fan-supply catchments. Most of the fans are of hyperconcentrated flow-dominated type, whilst part of those within the Eastern Carpathians are of debris-flow-dominated type. The fans' origin was provided by uplifted orogen and stable or subsiding foreland, giving both high gradient and orographic precipitation. The fans accumulated following a time lag after the post-collisional orogen uplift. The facies and architectural differences between the formations are associated with specific sedimentation and regional tectonics. In the Eastern Carpathians thin conglomeratic formation was deposited above the regional angular unconformity by predominant progradation in conditions of the tectonic quiescence and low accommodation space. In the South-Eastern Carpathians the fluvial fine-grained sedimentation gradually passed into the thick conglomeratic Cândești Formation during the attenuated subsidence of the Focșani Depression. As the accommodation space decreased, the aggradation of the formation gave way to its progradation.

Keywords: Eastern Carpathians, South-Eastern Carpathians, foreland, upper Cenozoic, conglomerates, alluvial fans

Introduction

Coarse clastic sediments and rocks are characteristic of the majority of orogenic areas, of which the Carpathians are no exception. They represent repetitive elements of the Oligocene to Quaternary age of the Eastern Carpathians (EC) and South-Eastern Carpathians (SEC). Most of these elements are exposed in rare outcrops or represented in boreholes, being the secondary accumulations within the orogen thrust-fold structures. Within these structures, the upper Cenozoic conglomeratic formations (CF) that are not intensively deformed, are the most accessible for detailed study. The conglomeratic formations comprise a wide assemblage of clastic rocks, from clay to boulders, but their notable features are the poorly sorted, well-rounded, very coarse pebbles and fine cobbles (– 5 to −8 φ). The researchers have considered the CF to be of different fluvial origin and age (from the Tortonian to the Pleistocene) in separate locations (Teisseyre 1933, 1938; Samodurov 1957; Tsys 1962; Raskatov 1966; Liteanu 1967; Murgeanu et al. 1967, 1968a, b; Saulea et al. 1967; Ionesi 1971; Grasu et al. 2002; Vaschenko et al. 2003, 2007; Andreyeva-Grigorovich et al. 2008; Andreescu et al. 2013). This is also associated with the objective difficulty to date them.

The aim of the study is to provide a reinterpretation of the origin and age of the conglomeratic formations, based on new sedimentological and geomorphological observations, supported by a comprehensive review of the current knowledge. This allows the common features of the CF to be emphasised and, at the same time, provides the opportunity to remove certain contradictions in the interpretation of their origin. It also offers the possibility to achieve a clearer picture of the sedimentary and tectonic events that resulted in the marked change from the initially sand to mud-dominated basin filling, driven by the abrupt input of extensive coarse clastic deposits.

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Geomorphic, tectonic and paleogeographic setting

The CF deposits are found in three areas in a 600 km long strip adjoined to Carpathians (Figs. 1, 2). They are scattered over the EC foothills (e.g. Vaschenko et al. 2003, 2007); in contrast, forming continuous and thick accumulations along the SEC flank (e.g. Andreescu et al. 2013), the surface of which gently slopes from the mountains towards the Romanian Plain. The CF units occur as remnants and continuous strata that range in thickness from a few metres to 190 m. They also occur as gravel placers on the surface and 'lag' deposits flooring the foothill valleys.

Tectonically the area under consideration belongs to the external eastern part of the pro-foreland fold-thrust belt of the Alpine–Carpathian–Dinaridic system of orogens. In accordance with general ideas (e.g., Schmid et al. 2008; Nakapelyukh et al. 2018; Roger et al. 2023; with the references therein) this belt appeared when Tisza–Dacia and Alps–Carpathian–Pannonian microplates moved to the northeast and east into the Carpathian embayment of oceanic to thinned continental crust. This is accompanied by collision with the East-European Platform, roll-back of a subducting eastern European lithospheric slab, slab-detachment and extension in the overriding plate. The internal thrusts of the noted belt were emplaced in the Late Cretaceous time whereas the external nappes are of the Neogene age. The thrusting ceased in the Late Miocene (11–12 Ma), followed by the crustal uplift and exhumation.

Since the late Langhian age the East Carpathian pro-foreland basin emerged as part of the eastern Para-Tethys Sea and developed in conditions of the thrusts' propagation and foreland asymmetric subsidence (Artyushkov et al. 1996; Leever et al. 2006). The nappes overrode the deposits of the basin almost completely or partly. The basin was filling down, changing its configuration and from the Tortonian age retreated in the SSE direction towards the present Black Sea (De Leeuw et al. 2020). In the SEC the foreland basin includes the anomalously deep (10–12 km) Focşani Depression, which subsidence accelerated from the beginning of the Pliocene when rest of the basin adjoined to the East Carpathians was already overfilled and partly occupied by the deltaic Balta Fm (Matoshko et al. 2016). In terms of the foreland stadiality the CF refer to the late (Molasse) stage (Crampton & Allen 1995) or overfilled phase with "molasse" style of sedimentation (Catuneanu 2019).

The CF occupy the area of the external nappes and partly the noted basin beyond the fold-thrust belt. According to Artyushkov et al. (1996) conglomerates up to 100–200 m thick were forming near the nappe front during ~1 Myr after each convergence phase. The Miocene-age units within the basin and Paleogene–Miocene rocks of the nappes in the EC, together with the Quaternary deposits in the SEC, underlie the CF. These deposits are exposed at the modern surface and are in part overlapped by the younger Quaternary subaerial strata.

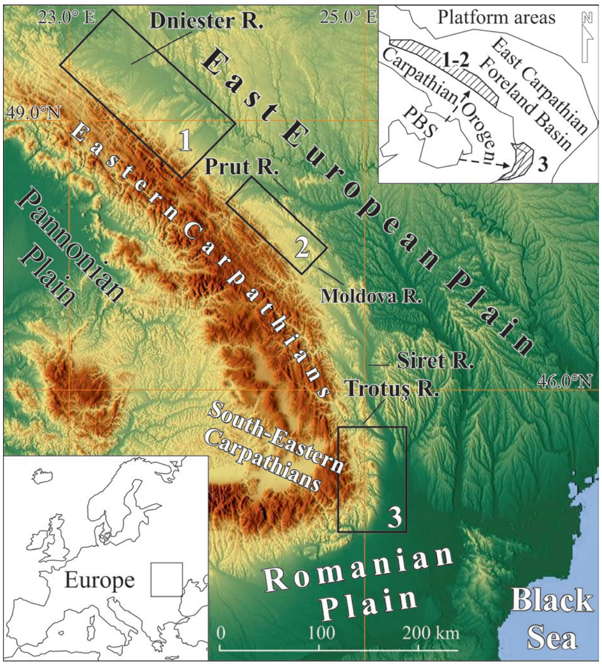

Fig. 1. General location map of the studied areas: 1–2: Eastern Carpathian area; 1 – Upper Dniester district, 2 – Upper Siret and Suceava–Moldova districts, 3 – South-Eastern Carpathian (Cândeşti Formation) area. PBS – Pannonian Basin System. The arrows show approximate directions of thrust transport.

Materials and methods

A number of literatures on CF, written in several languages and accumulated over more than 130 years of study (e.g., Athanasiu & Preda 1928; Teisseyre 1933, 1938; Murgeanu et al. 1967, 1968 a, b; Raskatov 1966; Vialov 1966; Liteanu 1967; Saulea et al. 1967; Ionesi 1971; Ionesi et al. 1971; Vashchenko et al. 2003, 2007; Necea et al. 2005, 2013; Oprea-Gancevici 2010; Jacyshyn 2010) has been reviewed. This review united the territories of Ukraine–Romania and their different regions for the first time. The old sites were located and revisited, as well as a number of new sites have been examined. The results of the contemporaneous sedimentological, lithological and geomorphological case studies of the CF in Romania (Grasu et al. 2002) and Ukraine (Jacyszyn et al. 2011) have been included in this synthesis.

The principal data were obtained during field study of the extensive (up to 1–2 km long and 20–100 m high) outcrops, embracing a significant part of the formation, together with small exposures (up to 3–5 m) from which facies details could be determined (Fig. 2). Occurrences of gravel at the terrain surface between exposures helped to correlate the CF units. The core descriptions of all available boreholes were also examined to further evaluate the general picture of the vertical successions in the rock sequences. The remote sensing materials, large-scale topographic maps and digital images of SRTM (USGS 2023), processed using GIS tools for relief analysis, were also examined. The same data and GPS navigation were applied to determine the accurate location of

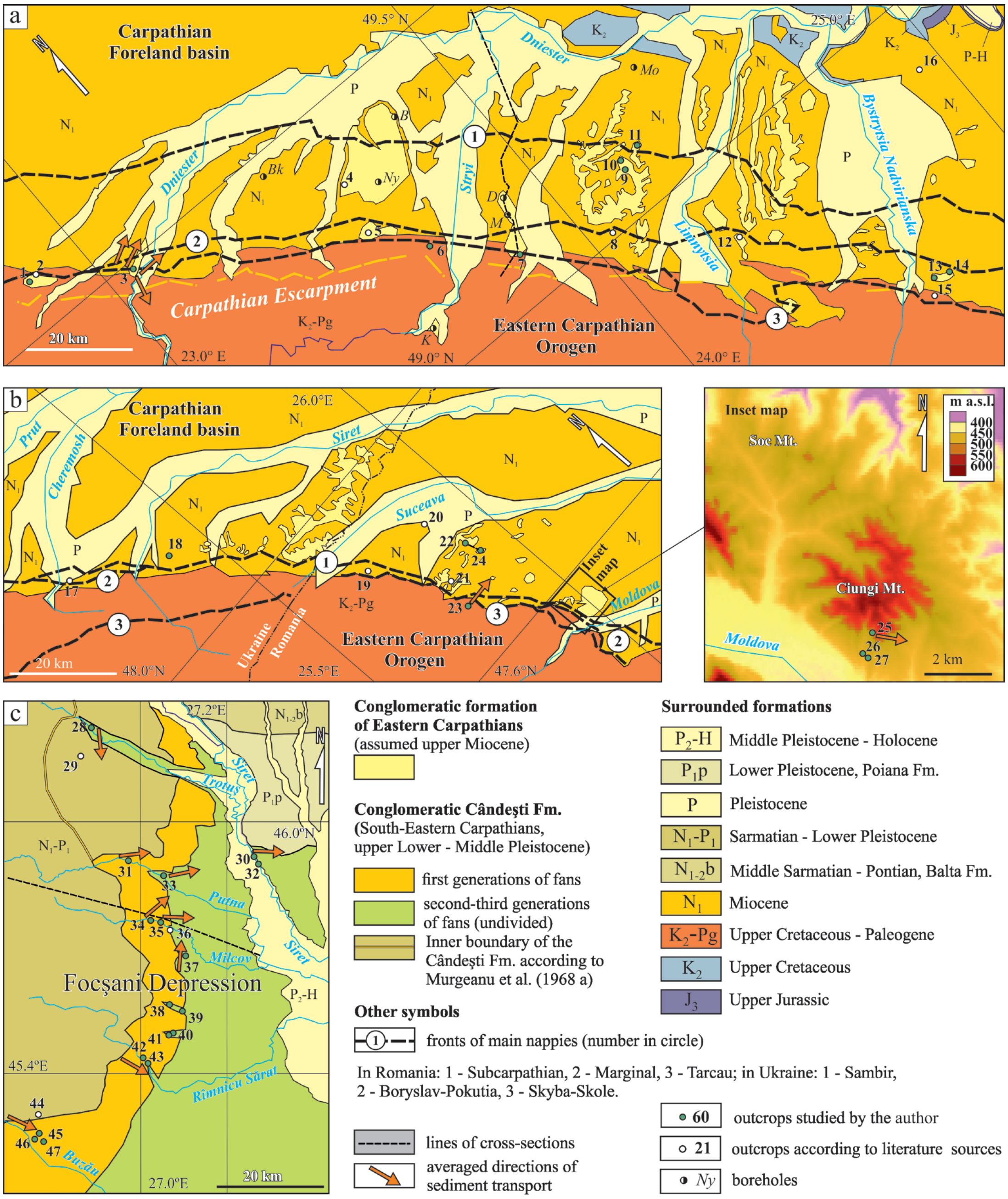

Fig. 2. Detailed location maps of the Eastern and South Eastern Carpathian conglomeratic formations and their geological surrounding linking to main tectonic structures according to published sources (Barbu et al. 1963, 1966; Murgeanu et al. 1967, 1968a, b; Vaschenko et al. 2003, 2007; Gerasimov et al. 2004; Matoshko et al. 2016) and author's data. **a** — Upper Dniester district; **b** — Upper Siret and Suceava–Moldova districts; **c** — South-Eastern Carpathians area (Cândeşti Formation). Note that position of the sole thrust front to the South of the Trotus R. mouth is unclear and cannot be shown on the map. For the locations see Fig. 1. Outcrops studied by the author, mentioned in the text and linking to the numbers: 1 – Radych H. 1; 2 – Radych H. 2 (Andreyeva-Grigorovich et al. 2008); 3 – Torganovychi; 4 – Mykhailevychi (Andreyeva-Grigorovich et al. 2008); 5 – Dobrogostiv (Raskatov 1966); 6 – Dovgoluka–Vulychne; 7 – Bolekhivska H.; 8 – Zalisia H. (Raskatov 1966); 9 – Bolokhiv 1; 10 – Bolokhiv 2; 11 – Zavadka; 12 – Krasna H. (Raskatov 1966); 13 – Verkhnii Maidan 1; 14 – Verkhnii Maidan 2; 15 – Loieva–Strimba (Teisseyre 1933; Samodurov 1957); 16 – Nadorozhna (Raskatov 1966); 17 – Vyzhnytsia; 18 – Migovo; 19 - Voitinel–Remezău (Miclăuş et al. 2011); 20 – Osoi H. (Ionesi 1971); 21 – Clit (Grasu et al. 2002); 22 – Burla; 23 – Solca 1, 2; 24 – Bodnăreni–Arbore; 25 – Tigani Scarp (Grasu et al. 2002); 26 – Ciungi 1; 27 – Ciungi 2; 28 – Viişoara; 29 – Pralea (Athanasiu & Preda 1928); 30 – Măligureni; 31 – Burca; 32 – Ionăşeşti; 33 – Sârbi; 34 – Pituluşa; 35 – Odobesti; 36 – Beciu (Necea et al. 2013); 37 – Cârligele–Vîlcele; 38 – Dragosloveni–Terchești; 39 – Dumbrăveni; 40 – Obrejiţsa; 41 – Liesti–Obrejiţsa; 42 – Podgoria 1; 43 – Podgoria 2; 44 – Câldâruşa–Vlâdeni (Van Baak et al. 2015, located by present author); 45 – Cernăteşti; 46 – Valea Nucului–Valea Puţului; 47 – Valea Puţului. The borehole names in Fig. 2a denote the nearest settlements, mentioned in cross-section and linking to the letters: B – Bilche; Bk – Bykiv; D – Dovge; K – Korchyn; M – Morshyn; Mo – Monastyrets; Ny – Nyzhni Gai. Carpathian Escarpment in Fig. 2b coincides with the Sambir–Subcurpathian nappe front. The loess formation partly overlapping the CF is not shown.

the sites investigated (see Appendix). Unfortunately, the insufficient number of exposures in this large area, the rarity of differently orientated exposures, and the inaccessible steep and loose cliffs restricted the possibilities of detailed lithofacies analysis.

The general approach to the characterisation of the sedimentary structures (Collinson et al. 2006), the identification of alluvial facies as a whole (Boggs 2006; Miall 2006), the alluvial fan facies specifically (DeCelles et al. 1991; Blair & McPherson 1994) and the modified Udden-Wentworth's grain-size classification (Blair & McPherson 1999) were adopted for both lithofacies and sedimentary architectural investigations. These methods were modified to take into account the specific object of the study. A number of principal publications which interpreted coarse clastic rocks in terms of sedimentary processes and environment (e.g., Costa 1984; Todd 1989; Sohn 1997; Pierson 2005), in addition to those noted above, were selected to support the interpretation of study's results. Among them, the general classification of the sediment-water flows (Costa 1984, table 4) served as the guide for interpretation of the transportation/deposition modes of facies and their associations.

Summary of previous research

South-Eastern Carpathians

Mrazec and Teisseyre (from Liteanu 1967) were the first to describe the gravels of this area in 1901 and referred them to the youngest Levantinian deposits (part of the Lower Pleistocene in the contemporaneous international stratigraphic division; see details in Results 'Massive clast-supported conglomerate'). They indicated the typical site of these gravels near modern-day Cândeşti (the site 'Valea Nucului – Valea Puţului', mentioned in present paper, is located nearby) and their occurrence as a tract along the SE and Southern Carpathians. Liteanu (1967), assigned them a more formal stratigraphic status terming them the 'Cândeşti Beds'. He noted their heterochrony within the Levantinian stage and that they included sands and clays. These beds were later mapped by Murgeanu et al. (1967, 1968b). Short summaries of the general geological characteristics of these deposits are presented in modern publications by Necea et al. (2005); Leever et al. (2006); Matenco et al. (2007); Andreescu et al. (2013) and Necea et al. (2013). In the last two papers they were incorporated into a regional stratigraphical scheme under the name 'Cândeşti Formation'. The noted authors implied fluvial genesis of the Cândeşti gravels but Andreescu et al. (2013) spoke out for their alluvial origin and Necea et al. (2013 with references therein) called them 'proximal alluvial fans'.

Eastern Carpathians

In the Ukrainian part of the EC, Teisseyre (1933) was the first to discover thick pebble-dominated strata at a very high position, apart from the river terraces, and associated it with the accumulation surface termed the 'Loieva Level' (site – Loieva–Strimba, Fig. 2a). Later Teisseyre (1938), Samodurov (1957) and Raskatov (1966) recorded some new gravel exposures at this level. Teisseyre (1938) described the area of this level within the EC foothills. Raskatov (1966) also found the 'higher' pebble level at the summit of the Krasna Hill ('Krasna Level') and some other uplands. The differentiation of the Loieva and Krasna levels is an obvious misunderstanding, since hypsometrically the localities from which they were originally named are, in fact, the same. It should also not be confused with the site Krasna (near the village with the same name).

Some researchers (Samodurov 1957; Demediuk 1976) combined the muds and sands overlain by gravel into one formation. During the geological mapping Vaschenko et al. (2003, 2007) distinguished clay-gravel strata within the interfluves of the foothill rivers, substantially expanding the area of the Loieva Level. The assumptions concerning the origins of the Loieva–Krasna levels differed significantly and included: ancient river formation (Raskatov 1966), onshore fans (Vialov 1966; Artyushkov et al. 1996), deltas (Tsys 1962), complex terrestrial formation (Vaschenko et al. 2003, 2007), and alluvial deposits of the river terraces (Gerasimov et al. 2004). All of these interpretations were based on the general lithological and geomorphological descriptions of the separate sites.

During the last 15 years new research has been undertaken at several new sites (Jacyszyn 2010; Jacyszyn et al. 2011). It includes additionally detailed studies of the hypsometric position of the alluvial deposits, grain-size and petrographic composition of gravel, as well as the reconstruction of paleocurrents from the orientation of clasts. These researchers revived the 'terraces concept', placing the Loieva Level within a single staircase of terraces, however they failed to offer any new serious evidence in support of their interpretation.

Most researchers considered these strata to be attributable to the Pliocene or 'Eopleistocene' which, in modern stratigraphy, probably implies that they are of Lower Pleistocene age. In particular, Artyushkov et al. (1996) noted that deposition of the youngest conglomerates started 3 Ma (Piacenzian age) and continued through the Pleistocene epoch.

In the Romanian part of the SEC, the first evidence described similar to the Ukrainian material was Ciungi Hill (hereafter – Ciungi for sites from this locality). Martiniuc (from Grasu et al. 2002) described these deposits in 1948 as being composed of an alternating sequence of gravel and sand with inclusions of clay, interpreting them to be of deltaic origin. Later authors supported this hypothesis and expanded the area of the conglomerates to include a foothill strip between the mountain margin and the Suceava and Moldova rivers (Barbu et al. 1963, 1966; Ionesi 1971; Ionesi et al. 1971). They distinguished their complexes at several altitudes, with several gravel horizons divided by clay-sand deposits. The microfauna within the clays sandwiched between two conglomerate strata in Burla (Ionesi 1971) indicated their Lower Volinian (i.e. the late Serravallian–early Tortonian) age.

In the most recent study of the Clit and Ciungi sites they have been subjected to detailed lithofacies analysis (Grasu et al. 2002). Debris flow facies, consisting of coarse gravel, have been distinguished in the Clit and interpreted as belonging to an alluvial fan. In the Ciungi four groups of facies dominated by pebble-cobble units, as well as minor sand and mud facies, have been described. As a whole all of them were associated with channel deposits of an alluvial fan environment.

The impossibility of identifying the location of the sites reported has meant that some of the information in some earlier publications could not be related to the deposits under consideration in this short review. In particular, this includes the strips along both flanks of the Prut valley in Ukraine and to several sites to the southeast of the Ciungi in Romania. Despite the assumptions made in some publications that the general characteristics of the different CF exposed in both Ukraine and Romania, this concept failed testing. Some of the controversial issues relevant to the present study are discussed in the results and discussion below.

Thus, the state of our knowledge about CF is as follows. In the area adjacent to the Eastern and South-Eastern Carpathians, there are specific conglomerates and sand-clays similar to them in occurrence. Many of them are exposed and lie close to the surface. The main areas of their distribution are established. However, lithological descriptions of the CF are scarce and lithofacies characteristics are absent. There is no consensus on what these CF mean in terms of depositional system, tectonism and climate. The estimates of their ages range widely, from the Serravallian stage to Pleistocene series.

Petrography of gravel clasts and provenance

Among the collection of papers reviewed above, petrographic studies of coarse clastic deposits should be considered separately. These investigations were not carried out systematically, but they are sufficient to provide a relatively clear definition of the provenance from which the clastic material of the CF originated.

The harder varieties of flysch rocks are predominant in the conglomeratic formations of the EC. In the Dniester Piedmont district, these rocks are represented (in order of distribution) by sandstones, marls and slates, while clasts of cherts, menilites and limestones are rare. According to Andreyeva-Grigorovich et al. (2008), the Radych conglomerate is composed of flysch-derived pebbles of the menilite-type black shales, cherts and quartz–glauconite non-calcareous sandstones. According to the author's observations and Vialov's (1966) opinion, the sandstone dominates in the clast assemblages there. Apart from flysch, single rounded fragments of igneous rocks, quartzites and quartz are also found (Samodurov 1957). Most rocks are derived from the outer orogenic zone; they are highly erodible on the one hand yet they are able to persist in gravel under intense abrasion in transport, on the other.

In the Suceava–Moldova district, Grasu et al. (2002) recorded only flysch clasts: sandstones, menilites, bituminous marls, mudstones, limestones, slates, quartz, and noted the absence of elements that originated from the inner Carpathian 'Crystalline-Mesozoic Zone'. Slightly different results are cited by Oprea-Gancevici (2010) for this district. Among the cobbles, he identified, 70 % were flysch rocks, 14 % quartzites, 12 % Mesozoic sedimentary rocks, 2 % metamorphic rocks and 2 % of Sarmatian (Serravallian–Tortonian) sandstones. At some sites the clasts petrographic assemblages are represented by a single rock type (e.g., Solca, uniform shale) or one type with a minor constituent of others (Clit, 90 % of local sandstone; Grasu et al. 2002). This composition can be attributed to the relatively close proximity to their source area and to the restricted small contributing gravel catchment.

In the SEC, among the gravels, Necea et al. (2005) distinguished Mesozoic and Paleogene rocks from the Carpathians nappes of the local area. Thus, the results of the petrographic studies confidently indicate a single Carpathian provenance for the clastic rocks and deposits considered here, derived specifically from the outer orogenic regions for the EC and inner orogenic regions for the SEC.

Results

Lithofacies

A starting point for this research was the spectrum of clastic rocks and sediment (conglomerates, sands and muds with minor gravel contents) having a similar occurrence and noted in previous studies. It turned out that they are represented by a number of associated gravel, sand and mud facies, separated by variation in grain-size, sedimentary structures and fabric of the conglomerates (Table 1). The facies identified herein and their reference codes follow Miall (2006). The facies interpretation, presented here, aims to reconstruct the most general characteristics of transportation and deposition of the sediments. Given that the previous and present studies showed no traces of *in-situ* fossils in the CF strata, marine and lacustrine environments are excluded from consideration in this summary. These beds also lack pyroclasts and therefore the volcanic environment is also excluded from this account. The glacial environment is excluded for paleogeographical reasons (see Discussion 'Climate-storms-floods complicity').

Gravel facies: massive matrix-supported (Gmm), massive clast-supported (Gcm), stratified clast-supported (Gh)

The gravel facies in exposures are almost all conglomerates with various degrees of packing and lithification. Their general predominant attributes are: very coarse pebble−cobble composition (− 5 to −8 φ) and poor to moderate sorting. There are rare finds of tree trunks directly in the conglomerates. The roundness of clasts is varied – from breccia to well-rounded ones – depending mostly on petrographic composition of rocks and probably less on the distance of transportation. Meanwhile, the presence of rounded fragments is noticeable

Table 1: Facies of conglomeratic formations (codes are according to Miall 2006).

Facies, codes	Dominated and maximal grain-size range	Sedimentary structures and fabric details for conglomerates	Thickness m	Numbers of sites in Fig.2 where facies are studied	Interpretation (modes of transport and deposition)
Conglomerate massive matrix-supported Gmm	−3 to −8 φ, up to −10 φ	absent, random clasts orientation	>4.0	1, 9, 10, 13, 14	plastic debris flow
Conglomerate massive clast-supported Gcm	−5 to −8 φ	absent, random clasts orientation	>3.0	3, 6, 7, 11, 22, 23	pseudo-plastic debris flow; hyperconcentrated flow
Conglomerate stratified clast-supported Gh-a	−4 to −8 φ, up to −9 φ	faint parallel stratification; imbrication; inverse grading; strings of larger clasts parallel to unit bounds	0.2–16.0	23, 25, 26, 27, 30, 31, 32, 33, 34, 35, 38, 39	hyperconcentrated flow; traction carpet
Conglomerate stratified clast-supported Gh-b	−3 to −7 φ, up to −9 φ	crude parallel and lenticular stratification	0.1–8.0	28, 32, 37, 43, 47	
Sand stratified Sl	3 to 1 φ	low-angle cross-stratification	0.2–2.0	25, 27, 34, 35, 39, 43	hyperconcentrated flow; water flow; suspended load
Sand massive and stratified Sm	4 to 1 φ	absent; faint parallel stratification	0.4–2.0	23, 25, 27, 28, 34, 35, 37, 39, 47	
Mud stratified Fl	>4 φ	faint parallel stratification	0.05–5.0	27, 30, 31, 34, 35, 39, 40, 41, 42, 47	
Mud massive Fsm		absent	0.4–12.0	11, 25, 30, 32, 33, 34, 35, 37, 42, 43, 47	hyperconcentrated flow; mudflow

at most sites. The obvious criterion for subdividing coarse deposits is the absence or presence of sedimentary structures, i.e., those of massive or stratified facies. The first of these includes the matrix- and clast-supported units.

Massive matrix-supported conglomerate (Gmm)

Description: Massive matrix-supported conglomerate (Gmm), visually 0.5 to 3–4 m thick, have been followed for several tens of metres in length. These units consist of pebbles and cobbles (−3 to −8 φ) in a mud and sand matrix, with a bimodal (gravel/sand-fines) grain-size distribution (Fig. 3a, b). The boulders (fine to medium, −8 to −10 φ) are frequent.

Interpretation: These facies can be related to a wide circle of the gravity sediment mass-movement and water-sediment flows. However, taking into account the signs of fluvial transport (rounded clasts and poorly sorting), only water-sediment flows remain relevant to the interpretation, and among them only a visco-plastic debris flow can deposit such facies, related to alluvial fans (Costa 1984). This is confirmed by observations of active modern debris flows (e.g., DeCelles et al. 1991; Blair & McPherson 1994; Brenna et al. 2020). According to Costa (1984) the debris flows are represented by a series of waves or surges, with periods ranging from a few seconds to several hours. They deposit 10^{-2}–10^{-1} m debris.

Massive clast-supported conglomerate (Gcm)

Description: Massive clast-supported conglomerate (Gcm) is somewhat coarser than the Gmm facies (−5 to −8 φ) but lacks boulders (Fig. 3c, d). These materials are tightly packed; the interstices are filled by sand and mud with different proportions, and grain-size trends are absent. However, the parallel orientation of the larger flat clasts is observed locally.

Interpretation: These facies can result from high-density (or hyperconcentrated) flow, driving gravelly traction carpets along stream beds (Costa 1984; Todd 1989; DeCelles et al. 1991) or a pseudoplastic (viscous-dominated) mechanism (Miall 2006).

Stratified clast-supported conglomerate (Gh)

Description: Stratified clast-supported conglomerate (Gh) is poorly to moderately sorted, tightly packed; the interstices being filled by sand and mud in differing proportions. This facies displays faint (Gh-a) to clear crude parallel stratification (Gh-b). The grain size decreases, sorting improves, and packing becomes weaker from the Gh-a to Gh-b facies.

The Gh-a facies is a conglomerate with 0.2–16.0 m thick units composed of coarse pebbles–fine cobbles (−4 to −8 φ) and solitary fine boulders (up to −9 φ). According to Grasu et al. (2002) the frequency distribution of the similar gravel constituents of the Ciungi is polymodal, its median varying between 1.5 and 6.9 cm. The Gh-a variety possesses blurred elements of the sedimentary structures: faint parallel stratification, distinguishable only from 10^1–10^2 m distance from exposure. In places, this stratification is emphasised by inverse grading (Fig. 3e). When studying the close-ups, weakly pronounced imbrication can be seen (Fig. 3f) together with separate short parallel strings of larger clasts (Fig. 3g). At the same time this variety can often look massive when observed in close-up.

The Gh-b variety is the finer conglomerate (from medium pebble to fine cobble, −3 to −7 φ), 0.1–8.0 m thick. It has a much clearer crude parallel and gently lenticular stratification (Fig. 3g).

Interpretation: The stratification and better sorting relative to the massive conglomerates suggest that the stratified conglomerates originated from traction carpet bedload of

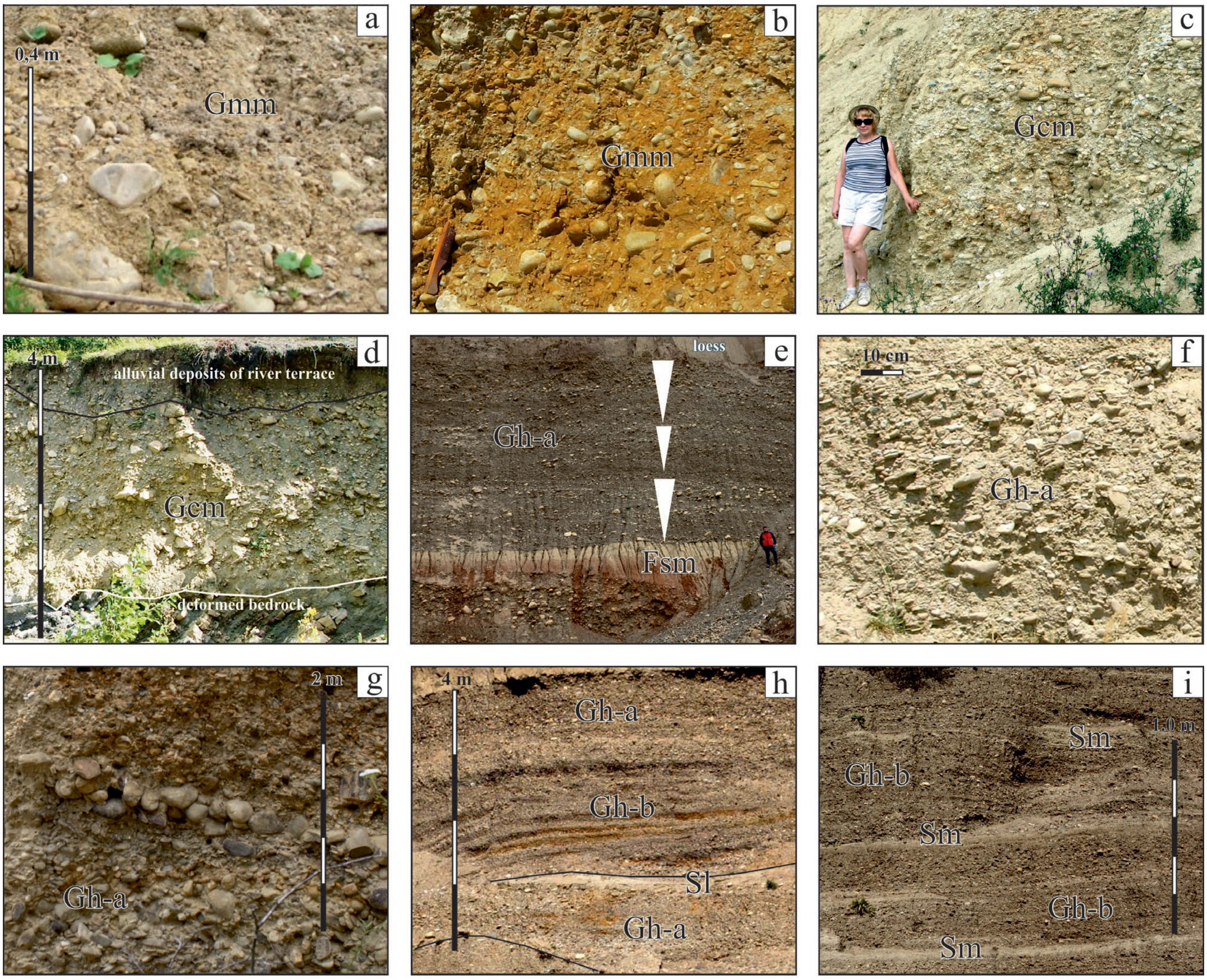

Fig. 3. The coarse-grained facies. The massive conglomerates: **a** — rich clayey matrix supported cobble and fine boulders of uniform sandstones (Gmm facies, Bolokhiv 1); **b** — gravel (from medium pebble to coarse coble) with quite various roundness in very fine sand matrix (Gmm facies, Bolekhivska H.); **c** — clast-supported very coarse pebble and fine cobble (clasts of different composition but sandstones are predominant) with tight packing and general random orientation of clasts (Gcm facies, Burla); **d** — clast-supported breccia of slabby shale clothes the surface of deformed bedrock grey mudstone (Gcm facies, Solca 2). The stratified conglomerates: **e** — faint parallel stratification and repeated inverse grading, separated the unit of clast-supported conglomerate on three sub-units (Gh-a facies, Sârbi); **f** — weakly pronounced imbrication in clast-supported conglomerate (Gh-a facies, Dumbraveni); **g** — string of cobble in clast-supported bulk of coarse and very coarse pebble (Gh-a facies, Tigani Scarp); **h** — different manifestation of faint parallel stratification (Gh-a, Gh-b facies, Ionăşeşti); **i** — regular planar parallel stratification in clast-supported medium-coarse pebbles (Ch-b facies) underlined by alternation with lenses of silty sand (Sm facies, Viişoara). Height of men is: ~1.80 m (Fig. 3h), woman – 1.64 m; length of comb is 14 cm. The black lines are erosional contacts, white line – non-erosional contact, white cones – inverse grading. The letter designations hereafter are facies codes (Table 1).

hyperconcentrated (sediment-laden) flow (Costa 1984; Todd 1989; Sohn 1997; Pierson 2005) which Blair & McPherson (1994) called 'the flow of supercritical conditions. The alternative interpretations, of water-current flow (DeCelles et al. 1991; Nemec & Postma 1993; Miall 2006) and debris-flood origins (Church & Jakob 2020; Brenna et al. 2020), can be excluded owing to the lack of cross-bedding and other signs of bedforms in the first case and the lack of an open-framework texture and distinct common orientation of clasts in the second.

Sand facies: stratified (Sl) and massive (Sm)

Description: There are low-angle cross-bedded (Sl) and massive to faint parallel stratified (Sm) sands (Fig. 4a, b), 0.2–2.0 m thick. They are unconsolidated and poorly to moderately sorted. The sand is very fine to medium (4 to 1 φ); massive sand is often silty and clayey. According to Grasu et al. (2002) the grain-size frequency distribution of sands in the Ciungi is polymodal. The sands occasionally include solitary 'floating' gravel clasts, their clusters and lenses. This gravel is

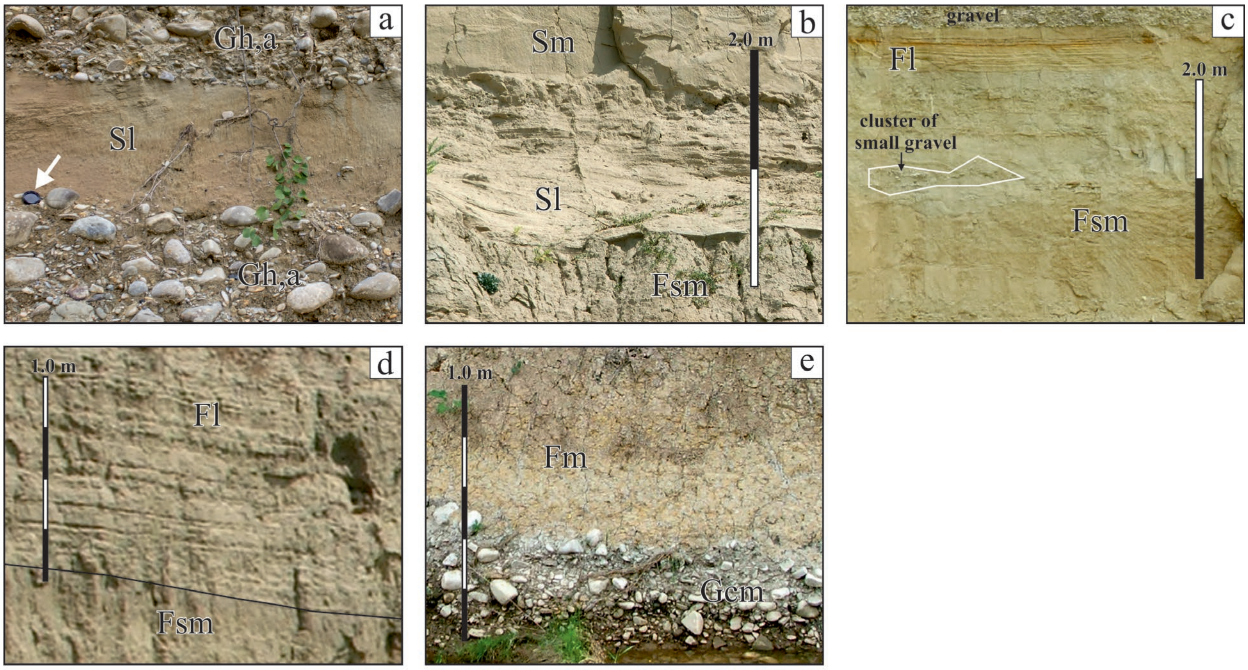

Fig. 4. The fine-grained facies: **a** — silty sand unit sandwiched between conglomerate units, clear cross lamination truncated at upper 'traction' contact (Sl facies, Pituluşa); **b** — alternation (from top to bottom): massive sand (Sm facies), cross-bedded sand (Sl facies) and massive silt (Fsm facies) with sharp, non-erosional contacts (Dragosloveni–Terchești); **c** — gradual transition downwards from thin bedded silt to its massive variety (Fl, Fsm facies, Podgoria 2); **d** — clear medium to thick planar parallel bedding in sandy silt resting with erosional contact upon massive silt (Fl, Fsm facies, Dumbrăveni); **e** — homogeneous massive clay lying upon massive clast-supported conglomerate with non-erosive contact (Fm, Gcm facies, Bolokhiv 1). The diameter of lens cap (Fig. 4a) is 5 cm.

usually finer (granules and fine-medium pebble) than in related conglomerates.

Interpretation: Since there are no signs of bioturbation and soft-sediment deformation, the massive sands (Sm) must reflect rapid deposition from suspension in the hyperconcentrated flow (Costa 1984; Todd 1989; Blair & McPherson 1994; Sohn 1997; Pierson 2005) and occasional faint parallel stratification can be related to the upper flow, plane bed regime (Miall 2006). The low-angle cross-bedded sands (Sl) are related to the nascent bedforms (Collinson et al. 2006) when depositing material from suspension re-engaged in movement by sporadic water flow.

Mud facies: stratified (Fl) and massive (Fsm)

Description: Mud facies (>4 φ) include faint parallel stratified (Fl) and massive (Fsm) varieties (Fig. 4c, d), 0.05–12.0 m thick. The silts are often sandy and include scattered gravel as in the sand facies. Rarely the Fl facies is represented by the uniform regularly bedded (1–4 cm thick) silts, lacking gravel, which sometimes fills minor incisions (Fig. 4d). The same silt may be obscured by a series of thin immature paleosols which developed along the bedding planes.

Interpretation: The transport/deposition modes of the silt facies are analogous to those of the sand facies, i.e., upper flow, plane bed regime (Fl) and rapid settling from suspension (Fsm) in interpretation of Collinson et al. (2006) and Miall (2006). Besides, the massive muds (Fsm) can be a product of mudflow.

Lithofacies associations

The facies associations include repeated vertical and lateral combinations of lithofacies. They allow to specify the depositional process (sediment flow type) suggested in previous section, and approaching a holistic view on the sedimentary environment. The only type of association that falls under the strict definition above, observed in a large number of sites, is succession Gh, Sl, Sm, Fl, Fsm. The hallmark of this succession is the frequent and well-developed alternation of the stratified conglomerate (Gh) and sand-mud facies (Sl, Sm, Fl, Fsm). The base of the conglomerates is mostly sharp planar or uneven. It often truncates the bedding of the underlying deposits at a low angle. This truncation reflects the inertial movement of the traction carpet immediately before its 'freezing'. This contact is referred to here as a 'traction contact'. The contact of the conglomerates with the overlying sand-mud facies is sharp and non-erosional.

All this suggests that the alternation consists of a number of couplets with coarser-grained and finer-grained components, as described by Blair & McPherson (1994). The thickness of the components varies within a wide range from a few centimetres to 12–16 m (Table 1), resulting in their highly variable ratio. They differ also in their regularity and geometry. This diversity is difficult to classify, but it is possible to demonstrate with specific examples which are conditionally divided into three varieties: A, B and C.

Succession (Gh, Sl, Sm, Fl, Fsm), variety A

Variety A is the most frequently encountered succession, representing the upper unit (up to 55 m thick) of the Pituluşa–Odobesti section. The couplet components are the thickest (conglomerates – up to 16 m, silts – up to 14 m) and gravel components are the coarsest (with solitary fine boulders) among all similar successions of the SEC (Fig. 5a, d). The planar parallel contacts of the couplets dip at an angle of 6° towards the SSE. The couplet components maintain their shape in exposures at different orientations. This shape points to a tabular geometry. The section of the Podgoria 1 is the analogue of the previous one (Fig. 6a), but differs from it by the appearance of underdeveloped paleosols in the uppermost finer-grained member of the succession. In some long exposures that finer-grained facies (Sl, Sm, Fl, Fsm) is seen to replace one another along the strike in one unit.

Succession (Gh, Sl, Sm, Fl, Fsm), variety B

Variety B is best represented in the Tigani Scarp. This scarp outcrops at the upper part of the Ciungi. section (Fig. 6b). This and one other outcrop of the Ciungi have been studied previously by Grasu et al. (2002), who identified six facies. Their description of four of them is close to those seen during the present study, but two of them, both concerning cross-bedded gravels, were not found in the well-represented and numerous Ciungi outcrops. It can be assumed that it was these facies that provided the basis for classifying the entire association of facies as channel deposits.

According to results of present study the Tigani Scarp (as of 2014) demonstrates 7–8 couplets (excluding those less than 0.2 m in height) making up an irregular succession up to 16 m in thickness. The general occurrence along the strike is lenticular; couplet components of variable thickness, with planar and uneven contacts, wherein coarser-grained components are followed continuously at a distance of more than 150 m and finer-grained lenticular components are 10^{-1}–10^{1} m long. The planar surfaces dip (2–3°) towards the SE. In some cases, there are large undulations (up to 3 m high) of their contacts and sharp changes of the sand-mud beds thickness, swelling or thinning (Fig. 6c). This is uncommon with channel erosion, but results from quite uneven deposition of the gravel beds and the enveloping occurrence of the sand-mud beds over gravel beds.

In places, both members demonstrate pinching out along the strike at a metre-scale distance. The fragments of the couplet alternation have been observed in the middle and lower slopes

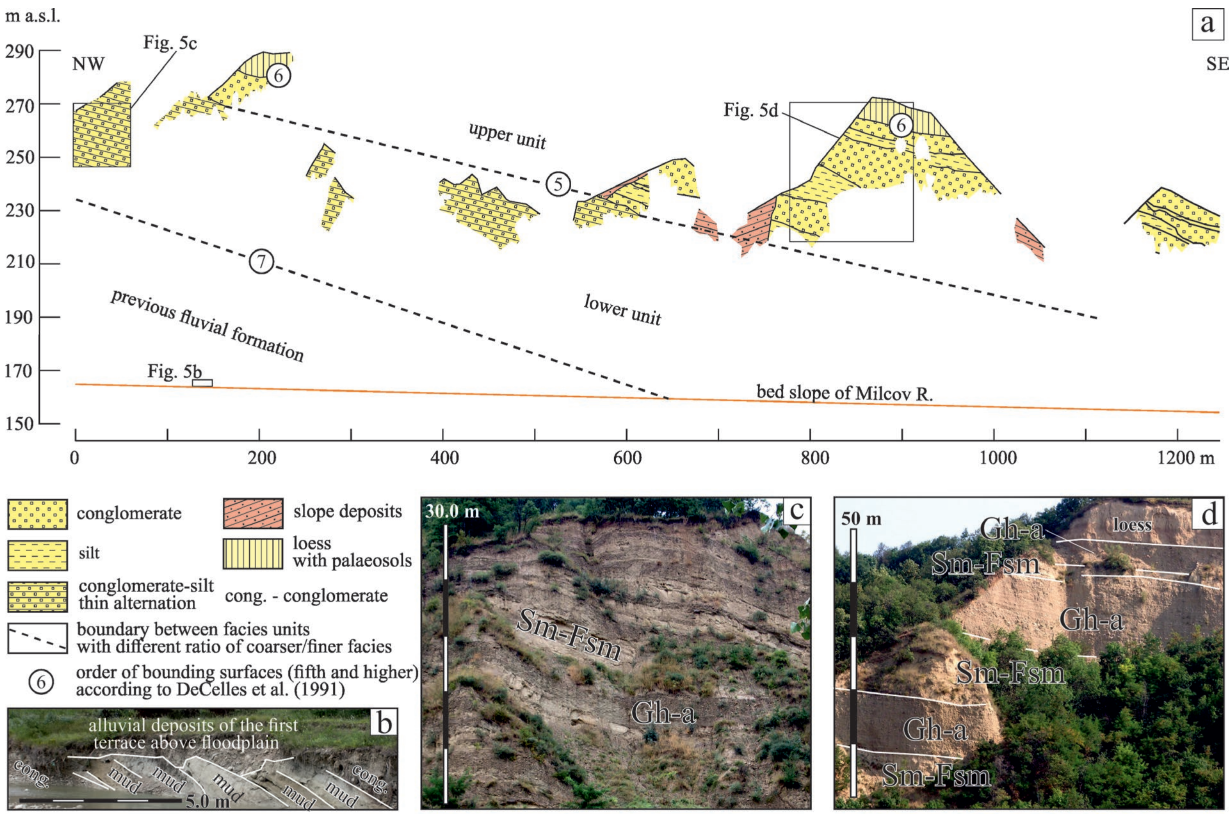

Fig. 5. The vertical change of the multiple couplet succession with the division on lower and upper units (Pituluşa–Odobesti, left flank of the Milcov valley): **a** — general outline compiled according to panoramic photograph with scale correction; **b, c, d** — details of units in corresponded photos. The explanations are in the text. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991).

of the Ciungi Hill in small outcrops. Also, in the middle and upper parts of the section of this hill, two units of badly exposed clays were found (facies F of Grasu et al. 2002) up to 6 m thick. Thus, it can be assumed that at least two thirds of the Ciungi Hill section, except for its uppermost part (where only coarse gravel is recorded in road cuttings), have the same uniform couplet-like vertical style interbedded with clay units.

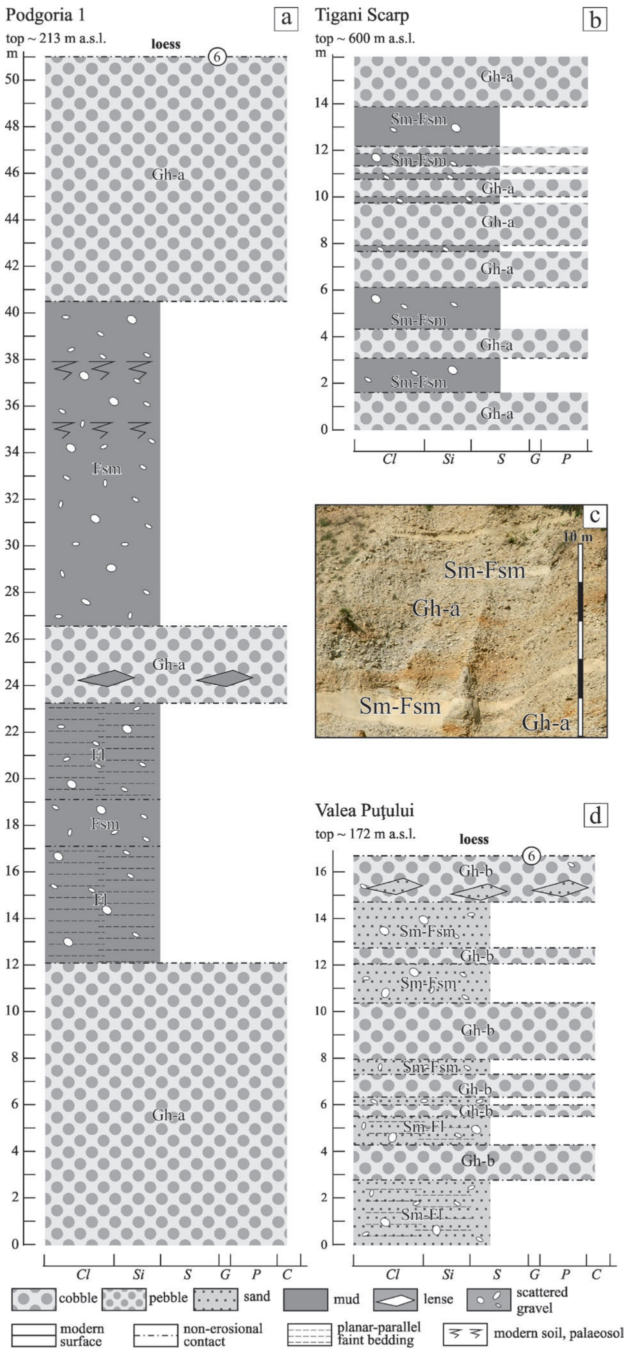

Fig 6. Couplet successions: **a** — Podgoria 1 (log); **b** — Tigani Scarp (log); **c** — geometry of couplet components along the strike (photo); **d** — Valea Puţului (log). The explanations are given in the text. For the figures in the circle see Fig. 5. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991).

The succession of the lower unit at the Pituluşa–Odobesti sites (Fig. 5a, c) is similar to the description above. The separate exposures show these deposits as a lenticular alternation of conglomerates with silts-sands of comparable thickness (0.2–3.0 m) and their couplet arrangement (Fig. 5c).

The succession in the Valea Puţului can also be referred to variety B. It is irregular, the thickness of the coarser-grained components ranging from 0.2 to 2.3 m and finer-grained ones – from 0.2 to 3.2 m and above, i.e. almost identical for both (Fig. 6d). Their contacts are unclear. Some are planar and smooth; others show gradual transitions between components and wedging out along the strike. In the Cârligele–Vîlcele the succession is represented by the same set of facies. However, they pass from one into another both vertically and along the strike, pinching out at a distance of 1–4 m and forming irregular lenticular and dove-tail shape arrangement. This is all seen as one stratum.

Sohn (1997, fig. 4) called such phenomena 'diffuse stratification'. In the Fig. 7 two diffusely stratified units, separated by a massive uniform silt (loess) with two mature paleosols, are illustrated.

Succession (Gh, Sl, Sm, Fl, Fsm), variety C

Variety C is a very regular uniform succession with 10^1–10^2 cm scale thickness of the couplet components with the coarser-grained units being dominant. It was encountered in the Viişoara (Fig. 3i), where it occupies a continuous section over 8 m thick. The notable feature of the Mălureni succession is the presence of a thin (0.8–1.0 m) layer of the Gh-b conglomerate that can be followed at a distance of more than 1 km. The Fsm silt, up to 14 m thick, visually underlies it and Fl–Fsm silt (4–6 m thick), with characteristic diffuse stratification and scattered pebbles in places, caps it.

Interpretation of the succession (Gh, Sl, Sm, Fl, Fsm), varieties A, B, C

The interpretation presented here, based on the couplet succession and, in particular leads to an alternative explanation of the Ciungi units' origins. Blair & McPherson (1994 and references therein) consider successions built up by repetitive couplets as most common for alluvial fans caused by sheetfloods, dispersed from fanhead in a radial sheet-like pattern. They have a high Froud number as well as a high attenuate and deposition rate. According to their data, numerous sheetflood couplets can be deposited during one flash-flood event, as recorded by the accumulation of as many as 15 such couplets (Blair & McPherson 1994). This is about 1.5–4.5 m (current estimation from the average thickness of couplets −0.1–0.3 m).

In the case of the CF, the thickness of individual couplets at some sites is much greater and their components differ strikingly in granulometric composition, compared to examples of other authors. Thus, the succession (Gh, Sl, Sm, Fl, Fsm) was related to much more powerful sedimentation,

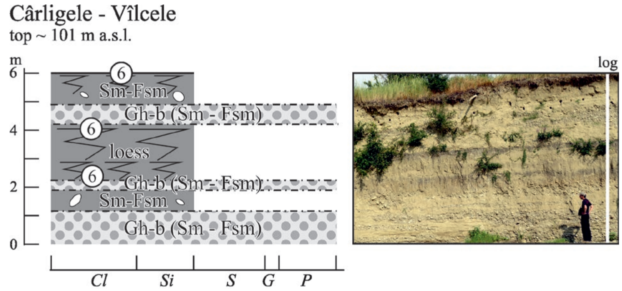

Fig. 7. The couplet succession demonstrates irregular diffuse stratification both coarser and finer components (Cârligele–Vîlcele, log combined with photo). The explanations are in the text, legend for the logs – in the Fig. 6., for the figures in circle see Fig. 5. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991).

super-abundant inflow of clastic material and its sharp separation on bedload (traction carpet), and suspension load during deposition.

Meanwhile, the relative intensity of sedimentation, attesting to the competence and capacity of the fan streams, was quite different. Most thick and clearly separated couplets and their components can be associated with maximal intensity values (variety A), while thinner couplets with the lenticular occurrence, blurred, diffuse stratification represent medium values (variety B). Those with small couplets and/or increased thickness of finer-grained components, reflect minimal values (variety C). It is notable that the A–B varieties of couplet successions are completed by conglomerates, probably resulting from wind winnowing of finer-grained deposits. This is indicated indirectly by the imprints of depositional interludes (paleosols, loess deposition) immediately preceding successions' completion. The fact that such aeolian process is possible is clearly shown by Pullen et al. (2018) in the example of modification of the unconsolidated gravels in the wind-eroded Hami Basin (northwestern China).

If the linkages of the facies associations with the well-known analogue examples from the literature and established views are correct, they generally correspond to sedimentation within an alluvial fan environment. Using Boggs' terminology (2006) with modifications relevant to the present case, most of the CF belong to hyperconcentrated-flow-dominated fans, and a part of them, that is those within the EC, to debris-flow-dominated fans.

Comparison of the CF facies and alluvial facies of ordinary foothill rivers

The facies of ordinary alluvial deposits of foothill rivers resemble most closely the facies under review here, with which they could be confused. Their characteristic is sketchy and cited only for comparison with the CF facies into which they are embedded near the studied sites. Most of them are represented by the Gh facies (clast-supported crudely bedded gravel) with an uneven expressive erosional base, often with lag deposits (Figs. 3d, 5c). They show normal grading and pass upwards into muds with scattered gravel (Fm facies). All other facies belonging to the CF (Table 1) are not known in the alluvial deposits of the foothill rivers, although they may occur as minor elements. Besides, they include organic-rich facies with undecomposed plant remnants in places. Thus, these facies form one succession of minor thickness (from tens of centimetres to 3–4 m) with two components: lower, channel and upper, overbank deposits, sometimes including abandoned channel interlayers. Grasu et al. (2002) demonstrated the transition from the debris flow facies, channel and overbank facies downstream at the Clit sites.

Architectural elements

The CF architecture is considered on the basis of the currently identified facies and their associations as well as the conclusion of their belonging to the alluvial fan sedimentary environment where they have specific features (e.g., DeCelles et al. 1991; Blair & McPherson 1994). The available data do not allow the architecture to represented systematically, but only some of its major elements. Nevertheless, their recognition significantly complements and/or confirms the interpretations made at the facies level and allows to outline a general picture of the fans' construction. The current architectural analysis relies on the ranking principles of the bounding surfaces and lithosomes (genetically related sedimentary bodies) developed specifically for the alluvial fans (DeCelles et al. 1991). In the current study they are combined into three ranks (lower, medium, high), characterised by their geometric shape, areal extent, facies and their associations (Miall 2006).

Lower rank elements

The separate faint laminae and beds form 1st order surfaces while those bounded facies and facies couplets should be referred to 2nd–3rd order. The dip of the couplet surfaces is the main element for reconstruction of the paleo-flows directions, while the measurements of imbrication are rare and random in nature. It is worth adding that flows paleo-orientation within the fan may potentially occupy nearly a semicircle (flow expansion angle, Blair & McPherson 1994). In this study, there was no possibility to make statistically relevant number of measurements, but the individual values obtained show a predominant direction close to normal towards the trend of the Carpathian chain (Fig. 2).

According to 2- to 3-dimensional observations, the bodies of couplets and their components are sheets and lenses dipping at 1–9° from the mountains being 10^1–10^3 m thick. They include an assemblage of the Gh, Sl, Sm, Fl and Fsm facies associated with hyperconcentrated flow. Their shape is also confirmed by the absence of channel margins, i.e., sheets and lenses were deposited by non-channelised flows typical for alluvial fans.

Concerning the debris and mudflow facies (Gmm, Gcm), the data about which are limited, one can assume that they were initially ordinary or multiple lobes and sheets.

Medium rank elements

DeCelles et al. (1991) and Blair & McPherson (1994) differentiate bodies of medium rank (restricted by 4th–6th order bounding surfaces): separate fan lobes, trenches or incised channels, backfills of trenches and individual alluvial fans, including their surface.

The lower-upper units of the Pituluşa–Odobesti section demonstrate an almost full proximal section of the fan lobe (Fig. 5). This is what DeCelles et al. (1991) called 'dispersion-stage deposits' while those that infilled the trenches were called 'trench-backfill deposits'. Importantly, those internal surfaces of the higher rank than facies and couplets are found only in the Ciungi section where rare mud beds separating couplet successions and their bounding surfaces can be considered as the 4th order. This indicates predominant continuous superimposition of couplets at the dispersion stage, leading to accretion of fan lobe vertically and radially, in places divided into substages.

The identified fan trenches are rare. One of them is presented in the Podgoria 2 section where the trench is clearly visible in the longitudinal section, being incised up to 30 m into the deposits of the more ancient fan lobe and can be followed by as much as 200 m (Fig. 8a). The trench backfill succession (5–6 m thick) and facies set are similar to the couplet pairs earlier discussed (see Results 'Interpretation of the succession (Gh, Sl, Sm, Fl, Fsm), varieties A, B, C'), differing from the facies of the common river terraces (see Results 'Comparison of the CF facies and alluvial facies of ordinary foothill rivers'). In the modern relief it is expressed as a narrow fan terrace. Downstream the backfill plunges below the valley bottom, being steeper than valley slope.

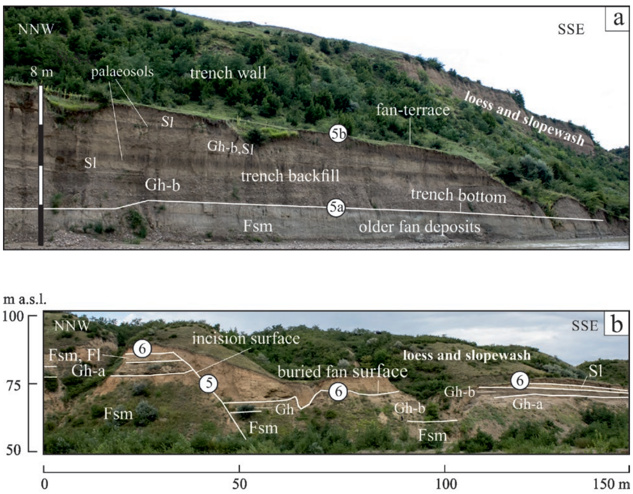

Fig. 8. The bounding surfaces and sedimentary bodies of medium rank: **a** – fan trench and trench backfill (Podgoria 2, left flank of the Rîmnicu Sărat valley); **b** – relationship between two lobes (generations) of fans (Mălureni – Jonăşeşti, left flank of the Siret valley). The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991).

In the Solca 2 locality the backfill is represented by Gmc facies (Fig. 4d). The trench-backfill deposits form segmental 'shoe-string' bodies similar to the usual fluvial channel fills, but in this case, they are represented by the facies of channelised hyperconcentrated flows.

Another section (Mălureni–Ionăşeşti) exhibits the incision and inset of the younger dispersion-stage deposits into the older units (Fig. 8b). This is noticeable by the different altitudinal position of the facies contacts and different facies sets in the older and younger generation fan lobes. Despite the fact that the CF deposits are covered by a loess and slopewash blanket, the scarp between fan lobes is traced as a kink in topography. Such kinks, tens of kilometres long, separate vast (incomparably larger than backfills) surfaces of two–three generations of fans in the SEC area (Fig. 2). The preserved, buried or slightly eroded surface of fans is recorded in sections of many sites (Figs. 5a, d; 6a, d; 7; 8b).

The appearance of trenches and new fan generations is considered as a natural stage of fan expansion, in particular this arises from oversteepening of the upper local fan gradient (DeCelles et al. 1991) and/or extension of the feeder channel onto the fan (Blair & McPherson 1994). It could also be a consequence of a fall in the base level of erosion.

High rank elements

The high rank of fan bodies is a fan formation or a tract of amalgamated fans (DeCelles et al. 1991). In the EC area the CF deposits are considered as a 'terrestrial formation' without a proper name (Vaschenko et al. 2003, 2007). In the SEC area they are recognised in the regional formal stratigraphy as Cândeşti Formation (Andreescu et al. 2013). The base of formations (classified as the 7th order boundary in terms of DeCelles et al. 1991), the bodies of the CF and their facies (facies associations) have specific characteristics within each CF area and district.

Upper Dniester district: In this district (Fig. 2a) the CF base can only be seen directly in the Bolekhivska H. It is planar, sharp, and non-erosive (Fig. 9a). The bedrock clays display a modified layer of reddish-brown colour directly below the conglomerate, whereas below these clays are grey in colour. This potentially relates to the weathering under terrestrial conditions. Other evidence is associated with boreholes (Fig. 9b, c), which show that the CF deposits (successions of debris and mud flows) overlie a regional erosional surface with angular unconformity. This surface cuts the deformed and non-deformed more ancient Miocene molasses as well as different units of the platform sedimentary cover under the common plane inclined towards the platform (according to the views: e.g., Demediuk 1976; Artyushkov et al. 1996; and materials: e.g., Vaschenko et al. 2007; Andreyeva-Grigorovich et al. 2008). It also extends beyond the distribution area of the CF, indicating that the erosion took place before the CF was deposited. According to DeCelles & Giles (1996) the noted unconformities are commonly found in the wedge-top depozone of the foreland basins.

The scattered CF erosional remnants are related to piedmont interfluves and cap the tops of several hills near the Carpathian escarpment. Some of them are close to the outlets of the mountain valleys, others are not. Together the remnants occur in a strip up to 40 km in width between the mountains and the Dniester valley (Fig. 2a), occupying the altitudinal interval of 500–300 m a.s.l. (Fig. 9b, c). The upper contact of the CF in most cases coincides with the modern piedmont surface, that also dips gently eastwards. The CF are overlain by gravity-flow and slopewash deposits in places as well as loesses near the Dniester valley. The CF thickness varies from 3–4 m to 44 m (Fig. 9c).

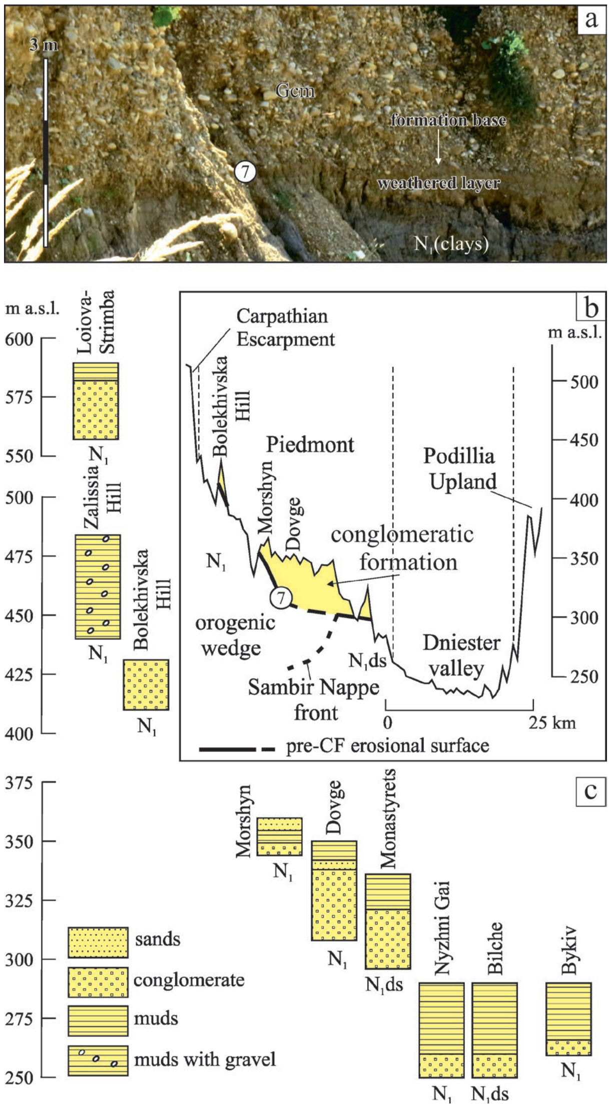

Fig. 9. General features of the CF occurrence in the Upper Dniester district: **a** — base of the formation (Bolekhivska H.); **b** — schematic general cross-section through the Upper Dniester district; **c** — thickest exposures and borehole sections which drilled the CF at full thickness and their rough lithologic division. N1(SN) – Miocene rocks of the Sambir Nappe; N1ds – the Sarmatian Dashava Formation. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991). For the location of sites, see Fig. 2a.

Upper Siret district: In this district (Fig. 2b) numerous small fragments of the CF share the same occurrence, but their minor thickness (up to 2–3 m) and absence of good exposures for facies identification provides very little information on their architecture and composition.

Suceava–Moldova district: Here the base of the CF deposits (couplet succession of the hyperconcentrated flow) lies on the deformed Miocene substrate of the orogenic wedge near the mountains, and further away from them – on the younger Miocene deposits of the foreland basin (Fig. 10). The contact of the first type is recorded in the Clit (Grasu et al. 2002) and Solca (Fig. 3d). The common occurrence of the CF as erosional remnants (Ciungi–Clit Piedmont outliers, Băcăuanu et al. 1980) does not differ there from those considered above, occupying the piedmont strip about 18 km wide (Fig. 2b). In the northern part of this district, the CF cover separate hills (up to 27 m thick) within top surface 560–420 m a.s.l.

However, in the southern part there are several large hills and one of them, the Ciungi Hill (Fig. 2c), has a unique dome-like geometry. It occupies about 15–16 km^2 composed of the CF deposits, 150–190 m thick (estimation from the results of detailed observations). In the Tigani Scarp the CF beds are gently (no more than 2–3°) inclined towards the ESE. Based on the continuous succession of the Ciungi it can be assumed that this hill-dome is a separate alluvial fan.

SEC area, Cândeşti Formation: In this area the beds of the Cândeşti Formation South from the Trotuş R. are inclined (6–11°) towards the depocentre of the Focşani Depression overlying conformably the tilted Late Cenozoic strata (Necea at al. 2005; Leever et al. 2006; Fig. 11). It is unclear how much of the observed tilt in the CF results from tectonic tilting, as argued by Necea et al. (2013), considering that the observed dips are common for alluvial fans (Boggs 2006). The elements of the conformability are found at the base of the Pituluşa–Odobesti section (Fig. 5a, b) and in the Cernăteşti. The appearance of rare thin conglomerate beds among the basin fluvial deposits (sands, muds, gley paleosols) of the more ancient formation is recorded in both cases. It can be assumed that the transition from one to another formation was gradual.

The CF are represented by couplet succession of the hyperconcentrated flow. In the SEC area the debris flow facies are not found. There are some trends in the facies occurrence in the relation to the Carpathian Mountains. The varieties of the couplet succession (A, B, C) are located sequentially in the direction away from mountains in order of decreasing intensity of sedimentary processes (see Results 'Interpretation of the succession (Gh, Sl, Sm, Fl, Fsm), varieties A, B, C'). In many cases this is accompanied by reduction of gravel clast size and increase in thickness of finer components.

In the SEC its main outcropping segment is followed in the tract 10–20 km width, whereas in its buried part the width of this tract could frequently attain 60–70 km (Fig. 2c). In the version of Andreescu et al. (2013) the strip of the Cândeşti Formation was extended far to the west within the foothills of the Southern Carpathians. The maximal observed thickness of the Cândeşti Formation (more 100 m) is recorded

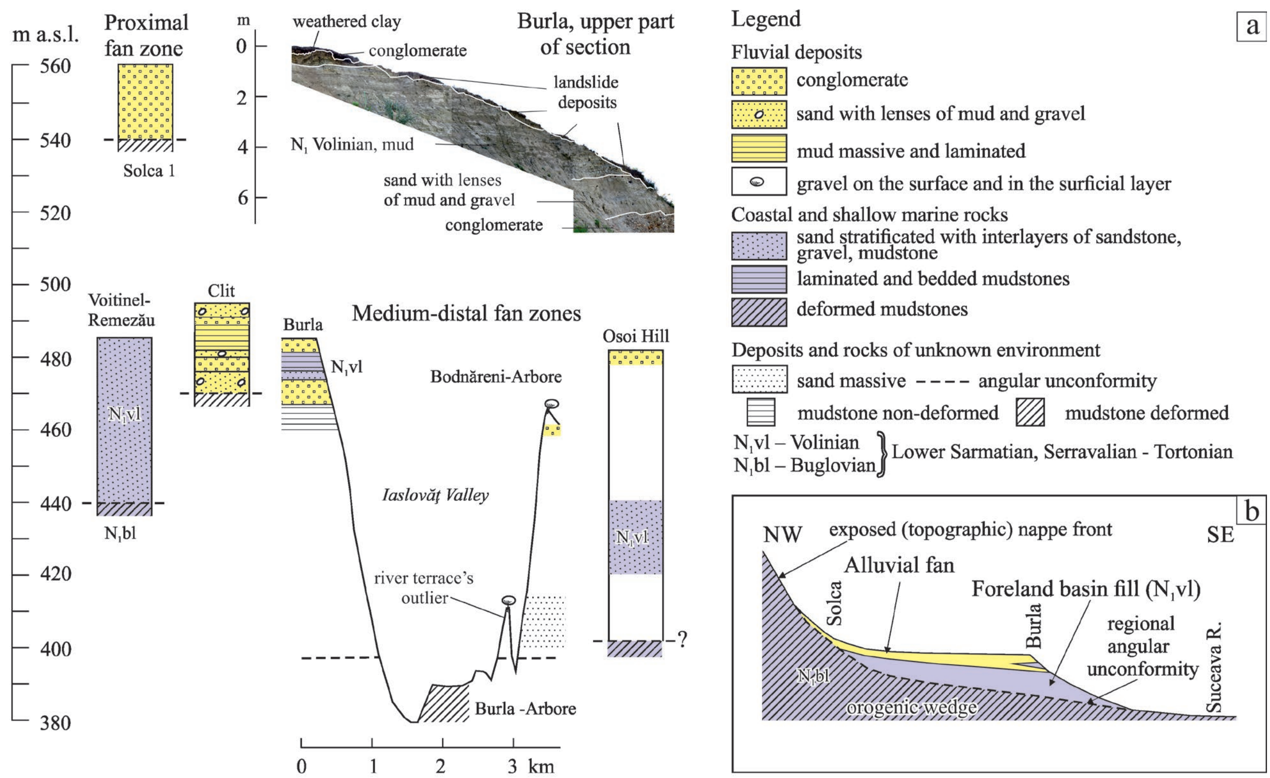

Fig. 10. The CF of the Suceava–Moldova district: **a** — cross-section and single sections based on data from previously published sources: Osoi Hill, Burla, Burla–Arbore (Ionesi 1971); Clit (Grasu et al. 2002, fig. 23); Voitinel–Remezău (Miclăuş et al. 2011) and author's data (Solca, Burla, Burla–Arbore, Bodnăreni–Arbore 1, 2) with interpretation of fan zones; **b** — primary occurrence of alluvial fan on hypothetical composite cross-section as version of the previous cross-section (Grasu et al. 2002, fig. 33). The age designations refer to the sections with microfaunistic data and stratigraphic interpretation. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991). The location of all the sites is given in Fig. 2b.

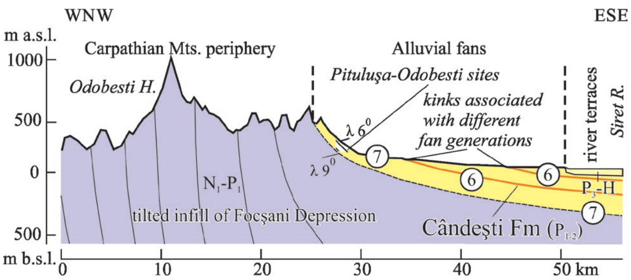

Fig. 11. The inferred schematic 2-D architecture of the Cândeşti Formation along the profile 'Odobesti H. – Siret R.'. λ – angle of dip of conglomerates at the base and roof of the Cândeşti Formation. The figures in circle are orders of the bounding surfaces according to DeCelles et al. (1991). The loess blanket over formation (out of scale) is not shown. For the location and symbols see Fig. 2, details are given in the text.

in the Pituluşa–Odobesti, while the thickness of the buried part in the Focşani Depression is estimated at 400–1000 m (Liteanu 1967; Matenco et al. 2007) and can even exceed 2500 m (Andreescu et al. 2013). These values are so high owing to the very probable inclusion of more ancient fluvial deposits into the Cândeşti Formation (see Discussion 'South-Eastern Carpathians, Cândeşti Formation').

In any case, a part of the Cândeşti Formation is a sedimentary infill body, and the other part is a sheet accumulated by amalgamated and prograded fans. It is much better preserved in comparison to its EC analogues, occupying the lowest part of the mountains and penetrating far into the foreland (Fig. 2c). It is expressed in the modern surface as smooth, gradually flattening slope, sharply contrasting the rugged relief of the westward mountains' periphery (Fig. 11). There this slope is essentially a modern piedmont. A major part of the CF is buried by loess which was an additional factor of the surface levelling. Necea et al. (2013) recorded river incision after the Cândeşti Formation emplacement.

The comparison of conglomeratic formations of the Eastern and South-Eastern Carpathians is summarised in Table 2. They are united by their setting, part of common facies and conglomerate clast size. Their differences include: presence (EC) and absence (SEC) of debris-flow facies, relationship to the underlying substratum, thickness of bodies (on the whole greater in the SEC) and preservation level (much greater in the SEC). To this it is worth adding the different CF provenance, according to the reports in the literature (see Summary of previous research 'Petrography of gravel clasts and provenance'): outer orogen regions (EC) and inner orogen regions (SEC). These differences are also discussed below.

Table 2: Comparison of conglomeratic formations of the Eastern and South-Eastern Carpathians.

Areas	Setting	Facies	Interpretation by flow type	Predominant clast size of conglomerates	Relationship to underlying stratum; the CF bodies shape	Thickness of the CF bodies m	Expression in modern relief
Eastern Carpathians	foreland, piedmont	Gmm, Gcm, Gh, Sl, Sm, Fl, Fsm	debris-mud flows, hypeconcentrated flow	−3 to −8 φ,	superimposition; covering, dome	10^{-1}–10^{2}	erosional remnants, trenches
South-Eastern Carpathians	foreland, piedmont	Gh, Sl, Sm, Fl, Fsm	hypeconcentrated flow	−3 to −8 φ,	gradual transition; infill, covering	10^{1}–10^{2}	regional slope, kinks, trenches

Discussion

Stratigraphical position and age of the conglomeratic formations

The ‘upper Cenozoic’ interval of the CF is the limit of the possible stratigraphical accuracy at present. The key reason for this is the almost complete absence of *in-situ* fossils. This results from the objective feature of the sedimentary environment of alluvial fans and, primarily, to the transience and high intensity of the processes of their formation. To date, numerical dating methods have not been applied to the CF. Therefore, the only available approach that can help in determining their stratigraphical position is to distinguish bounding fossiliferous marker beds or beds that can be dated by other methods. The available data for stratigraphic estimates differ in the districts studied.

Upper Dniester district

In this area the CF have been considered so far as the uppermost member of the terrace’s stairway in the Dniester River basin. In this version, they were attributed to the time period (Raskatov 1966; Vaschenko et al. 2003, 2007; Gerasimov et al. 2004; Jacyszyn 2010) which is referred to as the Lower Pleistocene Subseries in modern stratigraphical classification (Cohen et al. 2013). These ideas are speculative as a consequence of the absence of any factual basis. In particular, Jacyszyn (2010), referring to the paleomagnetic dating, found the Brunhes–Matuyama magnetic reversal in the paleosol overlying the alluvial deposits. Accordingly, the conclusion reached was that the Loieva Level dates to the late Lower Pleistocene Subseries. Alternatively, it can be suggested that this level could be older than this paleomagnetic event, but no older than the Dashawa Formation of the Sarmatian, the youngest of the formations underlying the CF (Fig. 12).

Andreyeva-Grigorovich et al. (2008) assigned the Radych Beds (conglomerates of the Radych H.) in their scheme of the Sarmatian as a terminating element above the Dashava Formation (Fig. 12). As noted in Results ‘High rank elements’, large-scale erosion took place after the last thrusting event (Sambir Nappe) and alluvial fans capped this erosional surface.

Suceava–Moldova district

In the Suceava–Moldova district the CF rest on the deformed mudstones, presumably of the Buglovian (Lower Miocene) strata, while eastwards, according to Ionesi et al. (1971), in the Burla the gravels overlie and are overlain by the Volinian or Volhinian in some schemes (Fig. 10). Ionesi (1971) distinguished the association of the Lower Volinian foraminifera with *Elphidium rugosum* in clays that occurred between two beds of conglomerates in the Burla (Fig. 10a). In the Fig. 10b. the geological position of the CF as a version of the previous cross-section can be seen (Grasu et al. 2002, fig. 33).

If the noted scant data, excluding unfounded opinions and considerable discrepancies (up to 3 Myr and more) in regional stratigraphic schemes can be formally summarised (Fig. 12), the probable time span of the conglomeratic formation in the Eastern Carpathians will be the Sarmatian in the modern regional stratigraphy corresponding to the Miocene Serravallian–Tortonian stages of the international stratigraphical chart (Cohen et al. 2013). The summary of the available facts and views suggests a possible sequence of events:

- the emergence and existence of a basin in the foreland (one of the bays of the Eastern Paratethys Sea) from the end of the Langhian to the end of the Serravallian age;
- the progradation and emplacement of thrust nappes, accompanied by partial involvement of basin sediments in thrusts and their partial overlap;
- the hiatus accompanied by significant erosion and the formation of a weathering crust and planation surface;
- the emplacement of the alluvial fans;
- the origin of the present-day transfer river network.

Thus, there was a certain lag interval between the completion of the nappe thrusting (11–12 Ma according to Artyushkov et al 1996; Schmid et al. 2008; Roger et al. 2023) and fan emplacement.

South-Eastern Carpathians, Cândeşti Formation

A short statement on the age of the Cândeşti Formation is cited in Section ‘Summary of previous research – South-Eastern Carpathians’ and shown in Fig. 13. It is also defined in the volume of the Romanian stage (3.0–2.6 Ma, i.e. upper Piacenzian) by Andreescu et al. (2013). One control point in the stratigraphy for this extended definition within the current study area is the Pralea site. The initial interpretation of this site, litho-facial features of the CF mentioned above, together with the results of the latest study by Van Baak et al. (2015), have led to the alternative more restricted view on the Cândeşti Formation time span.

At the Pralea locality, Athanasiu and Preda (1928) distinguished a thick alternation of pebble, sand, and clays with intercalations of lignite, which they referred to the Levantinian stage, without mentioning 'Cîndeşti Stratum' (the original name of the Cândeşti Formation) which was already known at that time. Their stratigraphical conclusion was based on the faunal remains recovered from the lignite. They included: '*Mammuthus meridionalis* Nesti 1825' (*Mammuthus* Brookes 1828) and *'Rhinoceros* cf. *etruscus' (Stephanorhinus (Dicerorhinus) etruscus* Falconer 1868). Based on modern views, the duration of both species (Lister & van Essen 2003; Radulescu et al. 2003) spans the Early Pleistocene, including the Late Romanian stage.

However, when the region around the Pralea was mapped by Murgeanu et al. (1968a), it was included in the Cândeşti Formation with the lower Lower Pleistocene strata. The Levantinian and its equivalents are not shown in the map legend. This is doubly strange since in the schemes of several contiguous maps (e.g., Murgeanu et al. 1967, 1968b) the Levantinian and lower Lower Pleistocene deposits are mapped separately, the latter being represented by the Cândeşti Formation.

New data could change this discordance. The lithological description of the Pralea section by Athanasiu & Preda (1928) is limited, but close to that made for the upper part of the Romanian section in the Slanicul de Buzau valley, recently studied by Van Baak et al. (2015) using combined paleomagnetic and paleontological methods. This section can be roughly interpreted as alluvial deposits with channel (sands, pebbly sands) facies passing upwards into predominantly an overbank facies (muds). According to the present author' observations, the Romanian pebbles are fine to medium in diameter, usually scattered in sands or forming thin layers of pebbly conglomerates. They therefore appear quite distinct from the coarse, thick Cândeşti conglomerates.

According to Van Baak et al. (2015) downstream of the Slănicul de Buzău valley, the top of the Romanian (1.778 Ma) is represented in the Câldăruşa–Vlâdeni; this locality being identified by the present author. It is separated from the base of the Cândeşti Formation (recorded by the present author in the Cernătesti) by sandy deposits. Considering their dip, the thickness of the unit is approximately a few hundred metres. The extrapolation of the sedimentation rate for the Upper Romanian (0.66 m/kyr, Van Baak et al. 2015) broadly implies a few hundred thousand years for their deposition. Thus, the Cândeşti Formation base falls in the late Early–early Middle Pleistocene interval (Fig. 13). Therefore, its inner boundary shown in Fig. 2c, is shifted much further east within the Trotuş - Putna interfluve in comparison to the geological map of Murgeanu et al. (1968a).

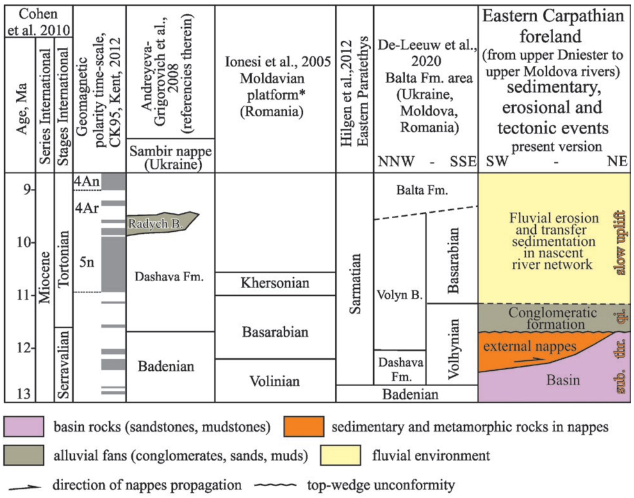

Fig. 12. Correlation scheme of the conglomeratic formation of the Eastern Carpathian foreland based upon the tectonic, sedimentary and erosional processes, with the regional and general stratigraphical schemes. The regional schemes are provided by nannofossil data. *The Moldavian Platform is marginal western part of the East European Platform in Romania. Abbreviations: b. – beds; Fm. – formation; sub. – subsidence; thr. – thrusting, folding, propagation of nappes; qi. – tectonic quiescence. The location of the cross-section is given in Fig. 2c.

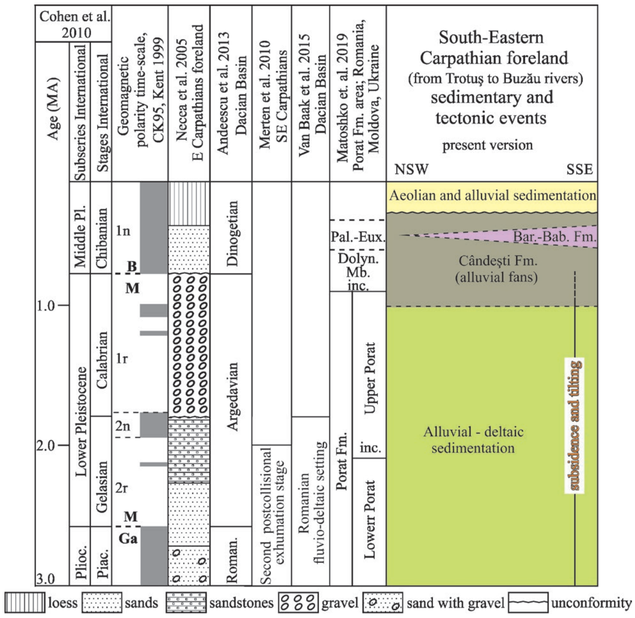

Fig. 13. Correlation scheme of the conglomeratic Cândeşti Formation of the South-Eastern Carpathian foreland at the background of the tectonic, sedimentary and erosional evolution with regional and general stratigraphical schemes, including a lithological column. The regional schemes are provided by geomagnetic evidence (Van Baak et al. 2015), apatite fission track (AFT) and (U–Th/He) thermochronology (Merten et al. 2010); mammal and mollusc paleontology (Matoshko et al. 2019). Abbreviations: Roman. – Romanian, Plioc. – Pliocene, Mid. Pl. – Middle Pleistocene, Piac. – Piacenzian, Pal.-Eux. – Paleo-Euxinian high stage, Bar.–Bab. Formation – Barboşi–Babele Formation in the lower Siret reaches (Saulea et al. 1967).

Based on infrared stimulated luminescence (IRSL) dating (Necea et al. 2013), the oldest age of the loess at its lower contact with the conglomerates in the Beciu is ~148 kyr (late Middle Pleistocene). This provides the upper age-limit for the Cândeşti Formation, whereas the real age of its top may be older. Necea et al. (2013) pointed to the occurrence of the denudation surface developed on the underlying conglomerates, and upon which the loess rests. This erosional surface is indicated by the fact that the couplets of coarser/finer-grained deposits in the uppermost part of some sections are completed by the lower component, while the upper finer component was probably eroded (see Results 'Succession (Gh, Sl, Sm, Fl, Fsm), variety A'). In any case, the contact of fluvial and subaerial deposits implies break in sedimentation, the duration of which remains unknown at present.

Thus, single evidence, less convincing in case of the East Carpathians and more justified in case of the South-Eastern Carpathians, confirms the differing age in the EC (Serravallian–Tortonian stages) and SEC (late Early–late Middle Pleistocene subseries). The reasoning behind regarding the two similar diachronous conglomeratic formations, spaced 10 Myr apart, supports the view that the Carpathian orogenesis activity was diachronous from north to south (Sanders et al. 1999). This tectonic diachroneity was developed by the formation of an axial movement aligned along the strike evolution of the fluvial system of the East Carpathian foreland (De Leeuw et al. 2020).

Factors of the conglomeratic formations' origin and development

General synthesis

Alluvial fans are often regarded as the end-points of particular erosional–depositional systems (e.g., Bull 1977; Ventra & Clarke 2018 with references herein). They are considered among major depositional bodies, accumulating over periods of 10^5–10^6 yr (Miall 2006). The examples of local estimates of the duration of fan (fan-complex) accumulation by the results of the magnetostratigraphy and optically stimulated luminescence (OSL) dates in Himalaya and Tibet, where conglomerates of similar thickness and coarseness to those of the CF give values of 10^3–10^5 yr in terms of their duration of deposition (Kumar et al. 2003; Srivastava et al. 2009; Gao et al. 2018). This is markedly closer to the assumed duration of the SEC fan development (Fig. 12).

In general, the direct agents initiating the development of the fans are flowing surface water, high contrast relief and erodible rocks in the catchment. Increasing flooding and gradient accompanied by high levels of rocks mobilisation and supply, together accelerate fan-formation processes as well as increasing sediment coarseness and the fan dimensions. The long-lasting (repeating) reproduction of these conditions resulted in the fan development. They include radial spread, aggradation, coalescence, progradation and entrenchment when autogenic hydrological and geomorphic control within the fans themselves becomes significant.

The boundary between the Carpathian orogen and its adjacent foreland was a potentially conducive place where all these conditions came together, but realisation of such potential in the form of the fan erosional–depositional systems depends on a certain interplay of the regional allogenic factors, among which the main driving of them are tectonics, base level and climate.

Tectonic background

The uplifted orogen and stable or subsiding foreland, which increases slope and relief as well as creates accommodation space, are obvious tectonic prerequisites for the mobilisation, transport and deposition of clastic material in the form of fans. It can also be added by the probable high seismicity characteristic of collisional zones. Artyushkov et al. (1996) included conglomerates in their scheme of plate convergence, which they applied to the Eastern and South-Eastern Carpathians as a significant component. This deposition occurred at the end of each convergence phase as a result of post-collisional orogenic uplift. According to Artyushkov et al. (1996) there were six such phases and 'the formation of the Carpathian Mts. during the last 3 Myr was associated with a deposition of large volume of conglomerates of Carpathian origin in the adjacent basins.'

The discussion concerning the links between tectonic movements and coarse-clastic sedimentation, and especially their timing is not over. The coarse clastics are conventionally interpreted to directly relate to the rejuvenated uplift of the orogen. According to Bull (1977) the optimal conditions for accumulation of fan deposits occur where the rate of uplift exceeds the rate of downcutting. The opposite idea associates marine, lacustrine and fine-grained fluvial units of the foreland basin with active subsidence, while coarse-grained (conglomeratic) units accumulate during relatively quiescent phases (e.g., Blair & Bilodeau 1988). The area studied here offers some arguments confirming the second of these conclusions.

In the EC the conglomeratic formation lies with an angular unconformity on the previous erosional surface within the collisional wedge, whilst beyond it dips gently, the deposit becoming thinner in the direction away from the mountains (Fig. 9). This is evidence of the tectonic quiescence within the foreland basin during sedimentation and unconfined fan progradation. Grasu et al. (2002) suggest that the coarse sediment of the Suceava–Moldova district originated under conditions of tectonic quiescence, following the deformation stage within the wedge-top depozone.

The Cândeşti Formation of the SEC is the top-most member of the quasi-continuous (lacking obvious regional breaks), deformed (primarily tilted) and uplifted, exposed Sarmatian–Lower Pleistocene basin megasequence that infilled the Focşani Depression during its subsidence (Necea et al. 2013; Fig. 11). According to a low temperature thermochronological study by Merten at al. (2010), this deformation and exhumation occurred following the Late Miocene termination of thrusting and included two exhumation stages at 6–3 and

3–2 Ma before present (Fig. 13). During the second stage, the combined zone of exhumation-subsidence shifted eastwards accompanied by syntectonic sedimentation and facies coarsening, including deposition of the Cândeşti conglomerates.

However, the timing of exhumation does not directly correlate with the sedimentary history (see Discussion 'South-Eastern Carpathians, Cândeşti Formation'; Fig. 12). According to Vasiliev et al. (2004) the same sequence, along Rîmnicu Sărat R. consists of mostly uniform mud-sand alternation with rare admixture of coarser sediment from the Upper Sarmatian to the Upper Romanian (10–2 Ma). Indeed, Van Baak et al. (2015) indicate some coarsening of the fluvial deposits during the Upper Romanian and continuation of this trend beyond the limit of the Romanian. This could reflect the period of active post-collisional uplift-erosion of the westerly adjacent nappe pile (Necea et al. 2005; Matenco et al. 2007). This was replaced by tectonic quiescence in the orogen, tilting and sharp deceleration of subsidence in the fans' emplacement area (Focşani Depression). The latter could have caused a slope break and change in alluvial sedimentation style (Fig. 13). Part of the conglomeratic formation was trapped in the depression, whilst the other prograded laterally following basin infilling.

Active faults bounding mountain fronts are also one of the key tectonic factors in the models of fan initiation and development (e.g., DeCelles et al. 1991; Blair & McPherson 1994; Miall 2006). They provide relief contrast required for sediment accumulation immediately behind the fault zone. Such faults are unknown in the part of the Carpathians under review here. However, the current topographic analysis of the eastern macro-slope of the EC has been shown to be steepening towards the foot and often the clear escarpment 100–150 m high within the range of 500–700 m a.s.l. (Figs. 2a, 9b). The lower break of the escarpment is evident in the interior boundary of the piedmont and CF. This landform intermittently stretches along the Carpathians and can be a reflection of a regional fault separated stable foreland and uplifted orogen.

Base level of erosion

Only one case of sea entering the CF area during its accumulation (Burla, Fig. 10, see Results 'Massive matrix-supported conglomerate (Gmm)') is recorded. At the remaining sites no traces of marine activity have been found, confirming terrestrial emplacement of the fans. At the same time there is some evidence (see Results 'High rank elements') of proximity the CF deposition to the sea (lake). In particular, Barbu et al. (1966) and Grasu et al. (2002) associated the development of the coarse sediment in the Suceava–Moldova district with the retreat of the Volinian (Lower Sarmatian) basin water body along the Carpathians.

As shown in the correlation table (Fig. 13), the time represented by the Cândeşti Formation partially falls on the depositional time of the alluvial Upper Porat Formation, which, including the Dolynske Member, is its neighbour to the East (Fig. 2c). The Porat Formation is considered to form part of the axial fluvial system of the East Carpathian foreland (De Leeuw et al. 2020). Both the Cândeşti and Upper Porat formations, being the final infills of the Eastern Dacian Basin, marked the establishment of terrestrial conditions in that area. The Dolynske Member provides the first tangible evidence for the breakthrough of the river Danube through the Galati Passage (Matoshko et al. 2019). This was caused by the fall of the main base of erosion (Black Sea basin level). It is highly probable that this forced regression simultaneously caused an entrenchment within the fans, together with the appearance of new fan generations in the SEC (see Results 'Medium rank elements').

There was also a short marine transgression in this area during the Paleo-Euxinian high-stand of the Black Sea (middle Middle Pleistocene, Barboşi–Babele Formation: Andreescu et al. 2013). During this phase the base level was raised, impeding fan progradation (Fig. 13).

All this supports the assumption that the alluvial fans, at least in two cases, were formed close to coastline or within the coastal plains, i.e., close to the main base-level of erosion.

Climate–storms–floods complicity

It is obvious that alluvial fans are fluvial products and, in any case, depend on precipitation and thereby on climate. Nevertheless, this relationship, as well as the role of climate in fan development, is not so clear. Some researchers suggest that the climate dominates control of erosional or depositional regimes related to fan sequences (e.g., Blair & McPherson 1994; Harvey et al. 2005). This factor could be considered through global and regional circumstances.

Molnar (2004) indicated the acceleration of erosion in mountains worldwide during the last global cooling and abruptly since the last 4–3 Ma. This could result from the glacial regime of the mountain streams associated with the spread of mountain glaciation to lower latitudes during the Quaternary. However, according to modern views, it only occurred during the Late Pleistocene (Rinterknecht et al. 2012; Popescu et al. 2017) in the Eastern and Southern Carpathians. The South-Eastern Carpathians were not glaciated (Necea et al. 2013, and references herein). At the same time during the cold intervals of the Pleistocene, an alpine zone of the mountains became more extensive. Physical weathering and gravitational processes intensified mobilisation and delivery of sediment. This could provide an additional explanation for fans' origin in the SEC.

On the other hand, the conglomeratic formations, and in particular the alluvial fans, are not only phenomena of the late Cenozoic. Are they all associated with global cooling? Both groups of the CF studied in this paper belong to the different spans of the common, long-term cooling trend that characterises the Cenozoic.

Considering the climatic conditions of the Volinian in the Romanian Carpathians, Sanders et al. (1999) admitted that the amount of precipitation was the same as today. Grasu et al. (2002) wrote about the warm and temperate climate, including periods of heavy rainfall. The reconstructions of precipitation change based on paleoecological interpretation of different groups of organisms give extremely contradictory assessments. According to Kvaček et al. (2006), Syabryaj et al. (2007) and Bruch et al. (2011) the climate of the late Cenozoic in the Carpathians and its surroundings was sufficiently humid and mountains were forested. Van Dam (2006) come to the conclusion that a large wet (800–1200 mm/yr) zone existed from Spain to Ukraine between 12 and 9 Ma. At the same time, according to Böhme et al. (2008) it was dry period for Central and Eastern Europe with typically less than 50 % precipitation compared to recent.

Comparing global and regional approaches to climate influence it should be remembered that both age groups of the CF exhibit similar facies assemblages and this prompts the writer to look for common reasons in their origin. Mountains in all climatic zones are moisture concentrators, conditioning orographic precipitation, an unconditional source for surface water flow. Researchers who examine the coarse-grained fluvial deposits or fans very often use the prefix ‘storm’ to describe the characteristics of facies or corresponding sedimentary processes (e.g., Costa 1984; DeCelles et al. 1991; Blair & McPherson 1994; Pierson 2005; Ventra & Clarke 2018). This associates conglomeratic formations with heavy storm showers, energetic snowmelt, or rain-on-snow events, all of which give rise to riverine floods.

As a result of what combination (coincidence) of different circumstances do these events arise? We don’t know for sure. According to O’Connor et al. (2002), there were ‘flood epochs – times when climate and topography collude to produce higher than-typical frequencies of large floods’. Such cataclysmic floods occur on timescales of 10^2–10^5 yr.

Acknowledgements: I am particularly grateful to Michal Šujan for the patient and thorough editing of the three submissions and Philip Gibbard whose contribution to improving scientific English style is irreplaceable. Thanks to Anton Matoshko and Arjan de Leeuw for providing field materials and friendly criticism, as well as to Crina Miclăuş, who assisted in finding important publications of the last century. The valuable corrections are done by František Teťák and other anonymous reviewers. I am grateful to my wife Luidmila Kosinskaia and Vladimir Kulichenko for their support in expedition work and my daughter Olexandra Matoshko for primary English proof reading.

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Appendix

Sites studied by the author, detailed location

Site name	Number in Fig. 2	North latitude	East longitude	Type of site
Radych H. 1	1	49.575255°	22.844200°	road trench
Torganovychi	3	49.482450°	23.083610°	road trench
Dovgoluka–Vulychne	6	49.195170°	23.080566°	road trench
Bolekhivska H.	7	49.085442°	23.846115°	pit
Bolokhiv 1	9	49.097841°	24.197477°	channel scar
Bolokhiv 2	10	49.083766°	24.194801°	channel scar
Zavadka	11	49.101567°	24.250768°	trench
Verkhnii Maidan 1	13	48.610831°	24.619572°	channel scar
Verkhnii Maidan 2	14	48.599000°	24.656000°	gully thalweg
Vyzhnytsia	17	48.236000°	25.193500°	road trench
Migovo	18	48.135622°	25.403045°	surface placer
Burla	22	47.778840°	25.907611°	slope scar
Solca 1	23	47.697808°	25.811304°	channel scar
Solca 2	23	47.697705°	25.811636°	channel scar
Arbore–Bodnareni	24	47.748768°	25.920659°	trench
Tigani Scarp	25	47.542037°	25.996575°	slope scar
Ciungi 1	26	47.540446°	25.992338°	slope scar
Ciungi 2	27	47.536627°	25.993064°	slope scar
Viişoara	28	46.230746°	26.879308°	pit
Mălureni	30	45.916404° 45.913571°	27.274009° 27.275371°	slope scar
Burca	31	45.911150°	26.967763°	channel scar
Ionăşeşti	32	45.913571° 45.911789°	27.275371° 27.276113°	pit
Sârbi	33	45.868659°	27.058907°	channel scar
Pituluşa	34	45.771867° 45.768455°	27.015153° 27.024626°	slope scar channel scar
Odobesti	35	45.770688° 45.766213°	27.028580° 27.044888°	slope scar channel scar
Cârligele–Vîlcele	37	45.683477°	27.108340°	channel scar
Dragosloveni–Terchești	38	45.566884°	27.067708°	channel scar
Dumbraveni	39	45.552534°	27.087649°	channel scar
Obrejiţsa	40	45.498075°	27.079189°	pit
Liesti–Obrejiţsa	41	45.492258°	27.068014°	pit
Podgoria 1	42	45.433654° 45.425779°	27.014844° 27.019793°	slope scar
Podgoria 2	43	45.412016°	27.023824°	channel scar
Cernătești	45	45.258028°	26.756072°	channel scar
Valea Nucului–Valea Puţului	46	45.255446° 45.247907°	26.748418° 26.747801°	slope scar channel scar
Valea Puţului	47	45.243222° 45.240985°	26.764058° 26.760859°	channel scar