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# VOLCANICLASTIC ALLUVIAL APRONS IN THE TERTIARY OF SOFIA DISTRICT (BULGARIA)

(21 Figs.)

## *Aluwialne utwory wulkanoklastyczne w trzeciorzędzie okolic Sofii (Bułgaria)*

(21 fig.)

Wojciech Nemec, Andrzej Muszyński: Volcaniclastic alluvial aprons in the Tertiary of Sofia district (Bulgaria). Ann. Soc. Geol. Poloniae, 52-1/4: 239—303, 1982, **Kraków.**

Abstract: The study area, comprising southwestern fragment of the Lozen Mts. and adjacent parts of the Vitoša and Plana Mts., includes deeply eroded rhyolitoid volcano of Mt. Vissoka Elha and the deposits of epiclastically reworked material derived from it. The volcano is radiometrically dated as Early Oligocene, and the age of its erosion is inferred as probably comprising Late Oligocene and Early/Middle Miocene. The volcaniclastics were deposited as two distinct alluvial aprons, or fans, characterized by different geometry and contrasting facies associations. The northwestern apron was steeper and built predominantly by debris-flows, while the eastern one was constructed mostly by sheet-flood/stream-flood and minor stream-flow processes. The contrasts between the aprons and their depositional pattern, all are explained in terms of fault-controlled syntectonic sedimentation. The role of the volcano weathering and hydrothermal alteration is also discussed. From the palaeogeographical reconstruction and characteristic bi-partition of the Tertiary molasse succession, a model of two-stage development and filling of the Tertiary successor basins is suggested for the region.

Key words: volcaniclastics, redeposition, debris-flow, sheetflood, stream-flow, alluvial fans, syntectonic sedimentation, molasse basin, Tertiary, Bulgaria.

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Treść: Na obszarze badań zidentyfikowano pozostałość po głęboko zerodowanym wulkanie riolitoidowym (Wysoka Ełha) oraz utwory wulkanoklastyczne powstałe z redepozycji materiału budującego pierwotnie jego stożek. Wiek wulkanu określono radiometrycznie jako wczesny oligocen, jego zaś erozyjne zniszczenie nastąpiło najprawdopodobniej w późnym oligocenie i wczesnym/środkowym miocenie. Wulkanoklastyki tworzą dwa rozległe stożki aluwialne, które różnią się geometrią oraz zespołem facji sedymentacyjnych. Obserwowany kontrast między stożkami oraz ich układ przestrzenny wyjaśniono w kategoriach syntektonicznej sedymentacji kontrolowanej przez czynne uskoki. Omówiono także rolę wietrzenia i hydrotermalnego przeobrażenia wulkanu w procesie redepozycji wulkanoklastyków. Poza wulkanizmem, innym nowym faktem w geologii regionu jest charakterystyczna dwudzielność (dwuetapowość sedymentacji) trzeciorzędowej molasy. W oparciu o ten fakt oraz wnioski paleogeograficzne zaproponowano w pracy pewien ogólny model rozwoju i wypełniania trzeciorzędowych basenów molasowych Zachodniego Srednogoria.

## INTRODUCTION

Volcaniclastic deposits have recently received much attention in the literature, primarily because of their importance in recognition of ancient volcanic terrains in deciphering the eruptive history of various volcanoes as it is recorded by deposits of volcanic origin, and in assessment of potential hazards which could result if similar eruptions were to occur in the future.

The present study is relevant to the importance of epiclastic volcanic deposits in stratigraphic record. Eroded and redeposited volcanic material may easily be diluted by ordinary clastic debris and is especially susceptible to weathering and diagenetic alteration, and so it rapidly loses its distinctive character. Therefore, the recognition and analysis of epiclastically reworked volcanic deposits and their discrimination from either *primary* (unreworked) or *secondary* (fresh-state reworked) pyroclastic sediments are often one of the most difficult tasks facing the geologist.

The Tertiary of the Sofia district, Western Srednogorie, has been recognized to contain thick succession of volcaniclastics, which consist of epiclastically reworked material derived from a (plagio-)rhyolitic volcanic source. The source for these deposits is inferred to have been the weathered, now entirely eroded, volcanic cone of Mt. Vissoka Elha volcano at the margin of Lozen Mountains. The purpose of the present paper is to describe the sedimentary characteristics of the volcaniclastics and to recognize their origin and mode of emplacement, then to discuss a model of their depositional setting and to suggest possible geological implications to the Tertiary molasse sedimentation in the region.

## REGIONAL GEOLOGICAL SETTING

The study area lies in the western part of the Sredna Gora Zone (Fig. 1), and includes southwestern fragment of the Lozen Mountains and adjacent parts of the Plana and Vitoša Mountains. The Sredna Gora Zone (or Srednogorie), itself northern strip of the Thracian (or Macedonian-Rhodope) Massif (S. Bončev, 1930), underwent rapid tectonic break-up during the Late Cretaceous time. During the late-Alpine ("miogeosynclinal") stage it became a typical rift feature, with extensive and strong magmatism (E. Bončev, 1975; see also Gočev, 1961). The prolonged tensional regime continued in the region during the post-Laramian ("molasse") stage, resulting in horst-and-graben tectonics and deposition of the Tertiary molasse (Jordanov, 1966, p. 125; E. Bončev, 1971).

In the Western Srednogorie, the Tertiary molasse basins display complex, horst-and-graben internal structure. These internal (or successor) basins are poorly separated one from the other, and are thought to have been formed and filled almost contemporaneously (Dimitrov, 1937; E. Bončev, 1971); the largest are the Sofia Basin, the Rakita Graben and the Čukurovo-Bistritza Graben. The basins are relatively "shallow". Their basin-fill displays thicknesses of the order of a few hundred metres, and consists of continental clastics which in many areas contain economically important brown-coal seams.

The age of the widespread Tertiary molasse in Western Srednogorie has been the subject of much controversy. Dimitrov (1937) suggested Pliocene age for these deposits, while Jaranov (1960) considered them as Oligocene. Jordanov (1966, p. 109), in turn, postulated Miocene age for these sediments, but suggested Pliocene age for the Tertiary in the Sofia Basin. Also E. Bončev (1971) stated that all of the Tertiary basins in Western Srednogorie are Neogene in age. T. Kostadinov and his co-workers (unpublished report V-103 to Bulg. Geol. Survey, 1973) assumed Neogene, $N_1$ (= Miocene?), age for the Tertiary in the area of the present study, but suggested Pliocene age for similar deposits extending north of German (Fig. 1). Miocene age has been palaeobotanically evidenced by Palamarev (1960, p. 578; 1964, p. 70; 1971, p. 163; 1979) for the Tertiary of the Čukurovo Graben, and consequently suggested also for the remaining parts of the region; on the basis of his palaeobotanical data Palamarev suggests Helvetian-Tortonian age for these deposits.

As regards the pre-Tertiary basement, the study area comprises fragments of the following three geological units (Fig. 1):

(a) to the northeast there is a southwestern fragment of the Lozen Mts., composed of the Permian-Mesozoic sedimentary rocks and Upper Cretaceous andesites (*i.e.*, andesite autoclastic breccias, andesites, trachyandesites, and basaltic andesites);

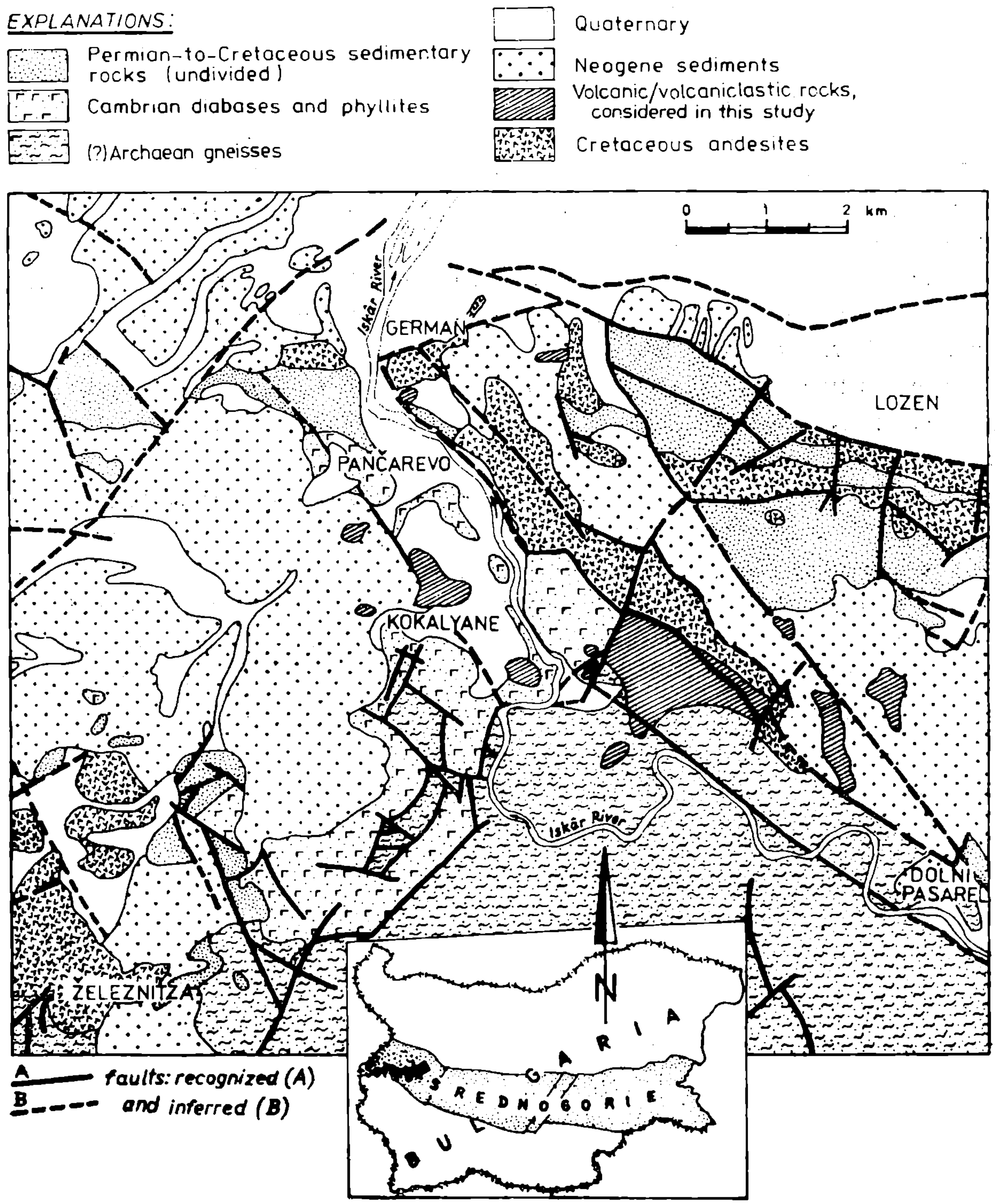


Fig. 1. Geological map of the study area (after unpublished report by T. Kostadinov and coworkers, 1973; slightly simplified and modified, with a few faults completed after Jordanov, 1966)

Fig. 1. Mapa geologiczna obszaru badań (na podstawie niepublikowanego opracowania T. Kostadinova i współpracowników, 1973; nieznacznie uproszczona i zmieniona, z kilkoma uskokami uzupełnionymi za Jordanovem, 1966)

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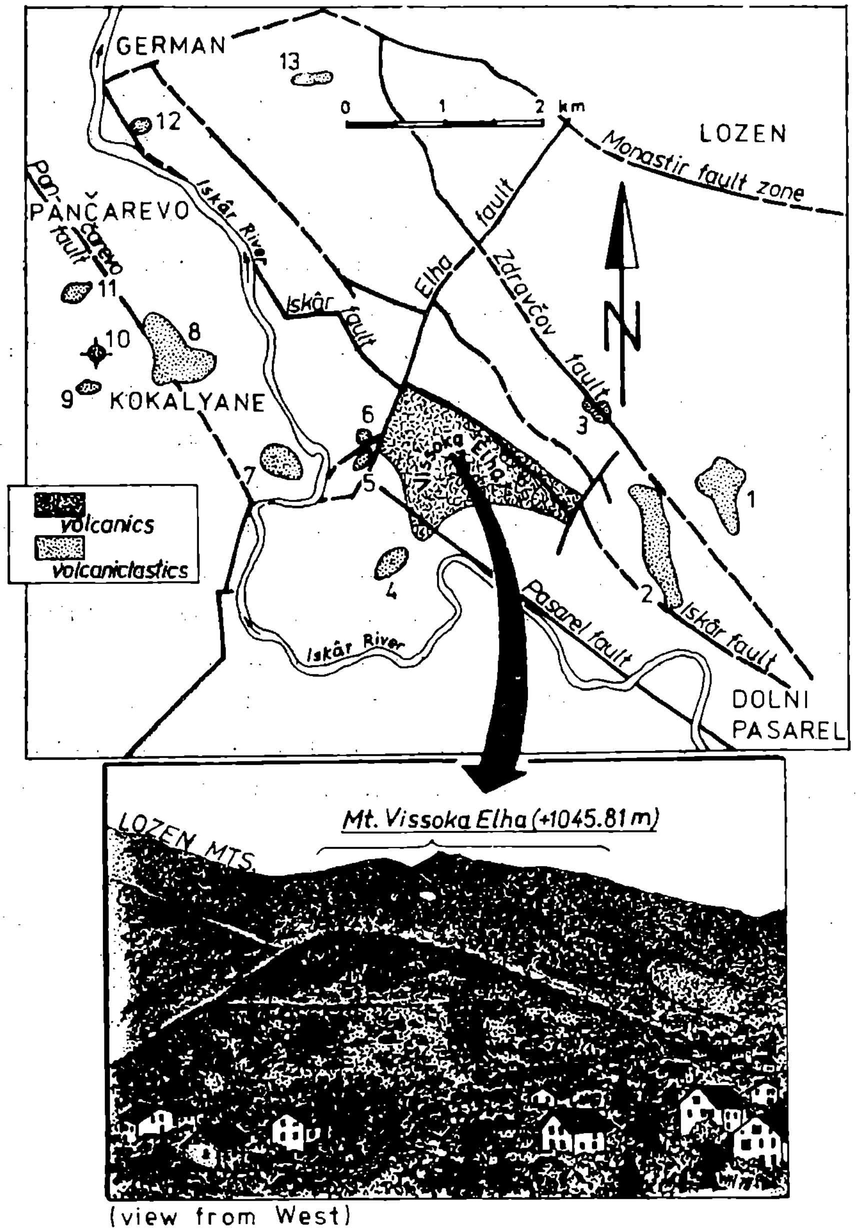
Fig. 2. Index map showing the areal distribution of volcaniclastics occurrences in relation to Mt. Vissoka Elha, and including a picture which shows the presentday morphology of this latter mountain (drawn from a photograph taken August, 1978); the hill in foreground is built of volcaniclastics, and contains outcrop no. 5. Numbers refer to volcaniclastics occurrences: 1 — outcrop belt at Stražarski Dol gorge; 2 — mapped out-




crop at Čubrinov Rid hill; 3 and 4 — mapped outcrops; 5 — roadcut outcrop; 6 — outcrop near an industrial magazine; 7 — mapped outcrop at Vassilitza hill; 8 — outcrop belt in the village of Kokalyane; 9 — roadcut outcrop near Kokalyane; 10 — borehole location near Kokalyane; 11 — outcrop at the valley of Bistritza Creek; 12 — outcrop at the bank of artificial lake at Pančarevo; 13 — outcrop near the road to German monastery. Faults' names are given after Jordanov (1966); the names of Pasarel and Elha faults are informally introduced here as new names

Fig. 2. Mapa ukazująca powierzchniowe rozmieszczenie wystąpień wulkanoklastyków w stosunku do góry Vissoka Ełha oraz ogólny widok ukazujący obecną morfologię tej góry (na podstawie fotografii wykonanej w sierpniu 1978 r.); wzgórze na pierwszym planie zbudowane jest z wulkanoklastyków i zawiera odsłonięcie nr 5. Numery odnoszą się do wystąpień wulkanoklastyków: 1 — pasmo odsłonięć w wąwozie Strażarski Doł; 2 — wykartowane odsłonięcie na wzgórzu Čubrinov Rid; 3 i 4 — wykartowane odsłonięcia; 5 — odsłonięcie we wcince drogowej; 6 — odsłonięcie w pobliżu magazynów przemysłowych; 7 — wykartowane odsłonięcie na wzgórzu Vassilitza; 8 — pasmo odsłonięć we wsi Kokalyane; 9 — odsłonięcie we wcince drogowej w pobliżu wsi Kokalyane; 10 — lokalizacja otworu wiertniczego w Kokalyane; 11 — odsłonięcie w dolinie potoku Bistritza; 12 — odsłonięcie w brzegu jeziora zaporowego w Pančarevie; 13 — odsłonięcie przy szosie do klasztoru w German. Nazwy uskoków według Jordanova (1966); nazwy „uskok Pasareł" i „uskok Ełhy" zostały tutaj wprowadzone nieformalnie, jedynie dla łatwiejszego powoływania się na te uskoki w tekście

(b) to the south and southeast there are the (?) Archaean gneisses which surround the northern margin of the Plana pluton and form part of what is termed the Plana Mts.;
(c) to the west there are the Cambrian diabase/phyllite complex and Upper Cretaceous andesites, surrounding the eastern part of Vitoša pluton and belonging to the Vitoša Mts. range.

At this point it is noteworthy that exactly at the juncture of these three geological units there is present the igneous body of Mt. Vissoka Elha (see Figs. 1 and 2). This is present at the cross-cutting of the Iskâr Fault Zone (SE—NW) and Kyustendil Fault Zone (SSW—NNE). The two fault systems are extensive fracture zones (up to few kilometres wide) which are internally complex and consist of numerous minor branches (for details see Jordanov, 1966; and E. Bončev, 1971). Many of the individual faults have been named by previous authors (Fig. 2; Jordanov, *op. cit.*), but several remain unnamed. Therefore, for convenience of referring in further text, the Pasarel fault and Elha fault are informally introduced here as new names for specific elements of the two fault zones, respectively (see Fig. 2).

In the study area (Fig. 1) the Tertiary cover is commonly fault-bounded, and occurs within some tectonic grabens and halfgrabens of different size and orientation. The largest is the graben extending between the vicinities of Železnitza and Pančarevo (SSW—NNE); this comprises almost total width (about 6 km) of the Kyustendil Fault Zone, and may be considered as part of the Čukurovo—Bistritza Graben of E. Bončev (1971). The herein considered fragment of this graben will be informally referred to as "Železnitza graben" in farther text; the above-named Elha fault is one of its marginal dislocations.

Transversely to the axis of Železnitza graben there goes the Iskâr Fault Zone, on which the former graben is superposed. The result is the presence of SE—NW trending, smaller graben (about 1.5 km wide) which is bounded by the Pančarevo fault and Elha fault, respectively (Fig. 2); this is often referred to as the "Pančarevo graben" (Jordanov, 1966).

The great majority of faults in the study area are older dislocations, which pre-existed the Tertiary. Later, however, they must have undergone some tectonic rejuvenation, and now essentially post-date the Tertiary molasse. Several of these faults have been active during the (syntectonic) Tertiary sedimentation, and represent in fact syndepositional faulting; some respective evidence will be given in this paper.

## PREVIOUS WORK AND INTERPRETATION

The prominent outcrop of igneous rocks at Mt. Vissoka Elha (Fig. 2) was among the first occurrences of the volcanics herein considered which were known to the earliest workers in the region, and which had been geologically studied as early as the second decade of this century. In 1912 Stoyanoff (cited by Dimitrov, 1937) considered them as "trachyandesites", but later Dimitrov (1937) has argued that they belong rather to calc-alkalic series and referred to them as "biotitic dacites". Jordanov (1966) renamed these rocks as "rhyodacites", but T. Kostadinov and his co-workers (1973, unpubl. report V-103), and Alexiev and Jurova (1976) again referred to them as "trachyandesites".

The igneous body of Mt. Vissoka Elha was interpreted as an intrusive stock, and its age was suggested to be younger than that of the adjacent (Upper Cretaceous) andesites (Dimitrov, *op. cit.)*; however, T. Kostadinov and his co-workers *(op. cit.)* state that its age is Late Cretaceous, rather than Tertiary.

The above-reviewed inferences, as made with respect to the "fresh--rock" outcrop at Mt. Vissoka Elha, were almost automatically extended by the previous authors to the nearby, isolated occurrences of fragmental volcanics (Figs. 1 and 2); this interpretational extrapolation was made by the previous authors despite the fact that these latter occurrences appeared to contain no solid rock and to be composed exclusively of volcanic rock debris. These isolated occurrences of the fragmental volcanics were interpreted by Dimitrov (1937) as some smaller volcanic stocks, which, he thought, were closely related to the Vissoka Elha intrusion (its presumed apophyses) and happened to be strongly weathered and disintegrated in their surfacial outcrops. This unsubstantiated interpretation, as introduced by Dimitrov in 1937 (paper republished in 1961), had continued almost to the present, while most of the later discussion was concerned with the composition and age of these presumed "intrusions" (see review above).

Later, however, on the basis of new outcrops (nos. 1 and 8, and part of no. 11 in Fig. 2) and a borehole profile (no. 10 in Fig. 2) Alexiev and Jurova (1976) recognized that the fragmental volcanics under consideration are virtually of epiclastic nature. They refer to these rocks as "volcanomictic rocks", and the major facts mentioned in support include the presence of foreign (nonvolcanic rock fragments and the occurrence of polymict gravels at the base of these volcaniclastics (see borehole profile in Fig. 3).

Moreover, Alexiev and Jurova *(op. cit.)* have recognized that the individual volanic fragments, though of variable appearance and preservation, do often display internal properties (texture, structure) which may be suggestive of subaerial lava consolidation. From this fact they deduce that

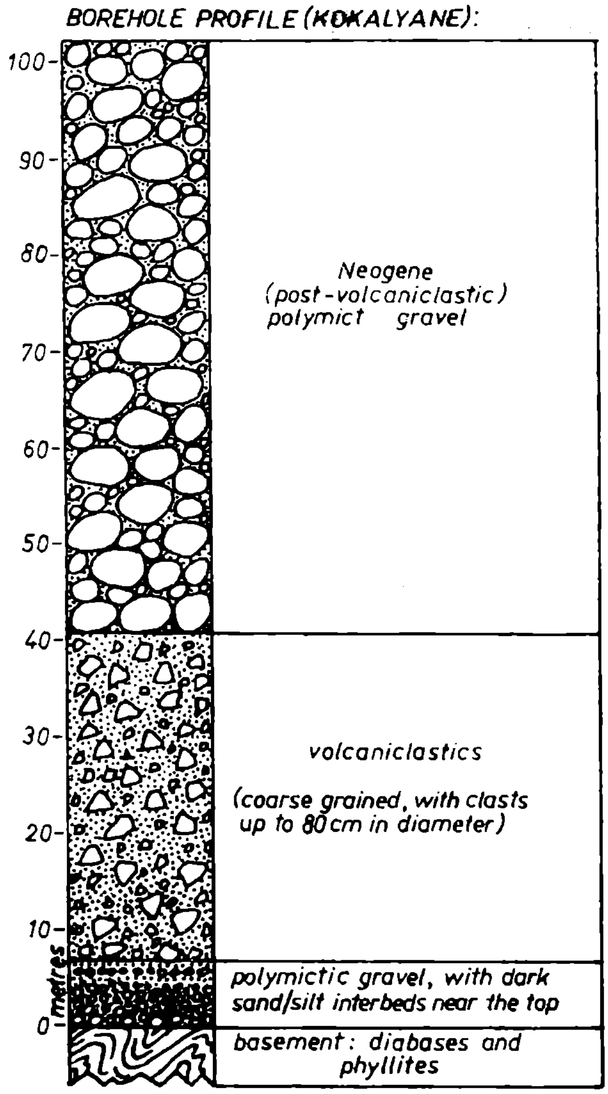



Fig. 3. Profile of the borehole near Kokalyane (from information in Alexiev and Jurova, 1976). For borehole location see locality no. 10 in Figure 2

Fig. 3. Profil otworu wiertniczego w pobliżu Kokalyane (na podstawie informacji w pracy: Alexiev i Jurova, 1976). Lokalizacja otworu przedstawiona jest na figurze 2 (no. 10)

there must have probably been some volcanic activity in the region. Alexiev and Jurova assume that the volcaniclastics are probably Miocene in age, and consequently suggest pre-Miocene (presumably Oligocene) age for the volcanic activity. However, no reliable arguments are presented by these authors in support of their inferences, and neither are also suggested the centers of inferred volcanic activity and the mechanism of the volcaniclastics emplacement. These and related problems, therefore, have still remained an enigma in the regional literature.

## TERMINOLOGY AND METHOD OF STUDY

The term "volcaniclastics" and the adjective "epiclastic" are used here according to Pettijohn's (1975) definitions.

The term "facies" is used in strictly descriptive sense of a rock (see *e.g.* Walker, 1975; Reading, 1978, p. 4), though interpretation of a process by which the rock in thought to have been formed is also given. For easier reading the facies are introduced in text under the headings which reflect their interpretation aspects, but facies descriptions and interpretations are clearly separated in text.

Following Bluck (1967) several authors have emphasized the importance of measuring bed thickness in conglomerates or gravels, especially with reference to the plotting of a bed thickness *versus* maximum particle size diagram. Maximum particle size (MPS) of a bed is obtained from the average of the 10 largest clasts measured within a lateral interval of about 2 m. Bed thickness is taken from the same place as MPS-parameter. Relationship between bed thickness and MPS has been determined using Pearson's correlation coefficient and the method of least-squares linear regression.

Volcanic rock compositions have been determined in thin sections and through chemical analyses. With a few exceptions, lithology of nonvolcanic clasts has been identified macroscopically. Clay mineralogy has been determined using the methods of differential thermal analysis and X-ray diffractometry.

## GENERAL STRATIGRAPHY

General lithostratigraphy of the post-Mesozoic rock units in the study area, as inferred from the field data and local palaeobotanical evidence, is given in Figure 4. Problems of formal lithostratigraphy are out of the scope of this paper, and, therefore, no formal names are introduced here for the distinguished rock units.

The volcaniclastics under consideration from the oldest portion of the Tertiary molasse in the region. The volcaniclastic pile is erosionally bounded, and so its original thickness is difficult to estimate. From palaeogeographical reconstruction, to be presented later, the original thicknesses are inferred to have been presumably about 200 m and more than 350 m immediately east and west of Mt. Vissoka Elha, respectively; even originally, however, the volcaniclastic pile was probably much thinner further out from the last-named locality. At present its thicknesses are considerably reduced, and probably nowhere exceed 100—150 m.

In their extent the volcaniclastics appear restricted to the nearest few kilometres around Mt. Vissoka Elha, and rest unconformably upon stratigraphically different parts of the basement (Fig. 1). Data on the basal contact are poor, but suggest that the volcaniclastic succession in usually

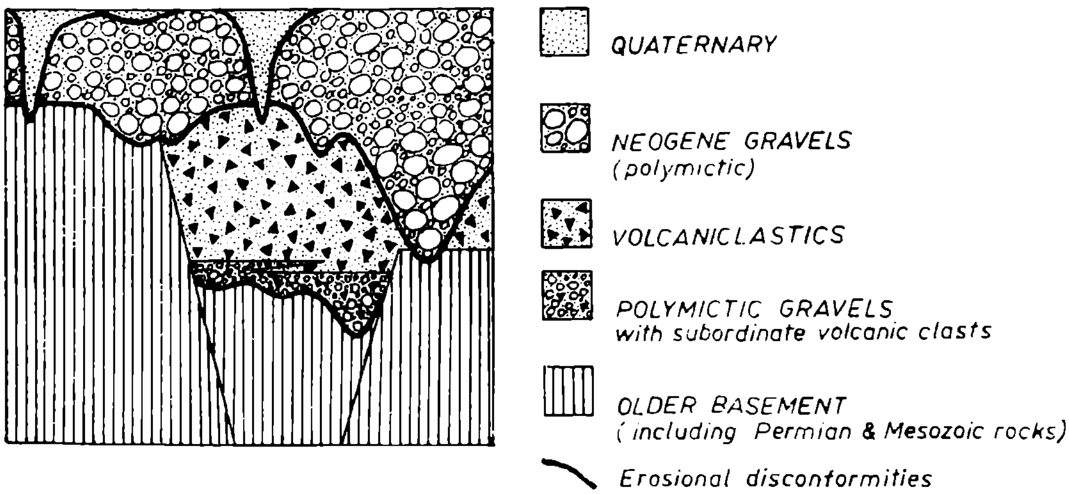



Fig. 4. Diagrammatic scheme showing the inferred post-Mesozoic lithostratigraphy of the study area (no scale)

Fig. 4. Diagram przedstawiający interpretowane następstwo utworów post-mezozoicznych na obszarze badań (bez skali)

underlain by a relatively thin (few tens of metres) sequence of polymict gravels (Fig. 4).

These basal polymict gravels are considered here as an integral portion of the major volcaniclastic succession, and this is primarily because they contain subordinate volcanic clasts of the same provenance as the overlying volcaniclastics. Moreover, the contact between the basal polymict deposits and the overlying volcaniclastics is likely to be transitional, rather than sharp, since isolated beds of similar polymict composition are locally found within the volcaniclastic succession itself; and even though the contact may be locally abrupt, it is by no means discordant. Thus, these polymict deposits are thought to be, in part, also a time equivalent of the volcaniclastics themselves.

The volcaniclastics have yielded no recognizable plant remains, but locally contain crushed charcoal and some drifted logs of charred wood; these latter have been identified by Dr E. Palamarev (personal commun., 1979) as probably representing parts of stems and branches of conifers. The basal polymict deposits, locally containing veneers of crushed charcoal too, have yielded only one recognizable plant fragment. This has been identified by Dr E. Palamarev as *Actinida faveolata* C. *et* E. M. Ried, which unfortunately gives no precise dating of the sediment since is known to have an Oligocene to Early Pliocene stratigraphic extent in this part of Europe.

The volcaniclastics are erosionally overlain by an extensive, up to 300—400 m thick cover of coarse polymict deposits (Fig. 4) which vary from cobble/boulder gravels to coarse, pebbly sands; these have been regarded as Neogene gravels by some of the previous authors. Good examples of their basal erosional contact with the volcaniclastics are at the outcrops at Bistritz Creek and Stražarski Dol gorge. The erosional unconformity is of considerable relief (locally up to 100—150 m) and probably reflects some important tectonic events in the region, but is thought to comprise geologically insignificant break in time. No significant angular disconformity has also been noted as associated with this particular unconformity.

Outside the extent of the volcaniclastics, the polymict Neogene cover rests directly on the pre-Tertiary basement and, presumably, locally also on the aforesaid "basal polymict gravels"; in this latter case the boundary between the two may not always be obvious. Thus, there is an overlapping relationship between the Neogene cover and the underlying volcaniclastics.

Finer grained sediments of the polymict Neogene cover, which are locally present and contain occasional mud layers, have yielded some recognizable plant remains. These have been identified by Dr E. Palamarev (personal commun., 1979) as *Myrica boveyana* Chandl., *Magnolia sineata* Kirchh., and *Passiflora kirchheimeri* subsp. *bulgarica* Palam. Through comparison with the palaeobotanically documented Tertiary profile of the nearby Čukurovo Coal Basin (Palamarev, 1964, 1971). Dr E. Palamarev has suggested Middle-Late Miocene (probably Helvetian-Tortonian) age for the abovesaid polymict deposits.

Since the volcanic body which did yield the volcaniclastics is radiometrically dated as Early Oligocene (see next section), consequently the age of the volcaniclastics is inferred here to be between the Late Oligocene and Early Miocene; strictly saying, their age is probably not older than the Late Oligocene and not younger than the early Middle Miocene.

Quaternary deposits in the study area are restricted mostly to river valleys, and comprise also local colluvial sheets and recent landslide deposits. They are, however, more extensively developed north of Lozen and German (Fig. 1), where they cover some broad areas composed of the Neogene gravels.

## PETROGRAPHY, COMPOSITION, AND PROVENANCE

### VOLCANIC AND VOLCANICLASTIC ROCKS

Volcanics of Mt. Vissoka Elha. These are petrographically and chemically monotonous. They generally contain 10—15% phenocrysts, which include mostly plagioclase, Mg-biotite, and Ca-amphibole; neither quartz nor potassic feldspar are developed as phenocrysts. The plagioclase phenocrysts display an oligoclase-andesine composition and are often zoned, showing an oligoclase outer portion and an andesine core. Amphibole and biotite are equally abundant, and their total amount is up to 3%. The amphibole phenocrysts are common in these volcanics, but are far less frequent in the respective rock fragments constituting the volcaniclastics (see below).

Rock groundmass consists of K-feldspar/quartz microlites displaying a complex mosaic of intergrowth. Primary accessory minerals are magnetite, apatite, zircon, and moasenite. Among the secondary minerals are cristobalite, K-feldspar, sulphides (pyrite, galene, sphalerite, and chalcopyrite), clay minerals (mostly mixed-layered montmorillonite-illite), calcite, zeolite heulandite, hematite, and barite.

These volcanics always show porphyric texture, including both primary crystalline textures and secondary (recrystallization) textures. More detailed petrographical description of these rocks is given by Muszyński (1980).

The volcanic rocks under consideration have been petrochemically classified by Muszyński *(op. cit.)* as ranging between plagiorhyolite and rhyodacite, with the former composition corresponding to the volcanic fragments present in the volcaniclastics and with the latter corresponding to the volcanics of Mt. Vissoka Elha. Absolute age determinations (K-Ar method) of Mt. Vissoka Elha volcanics are: 32, 34, 35, 36, and 38 ($\pm$ 2) m.y. BP, indicating Early Oligocene age of the magma emplacement.

V o l c a n i c l a s t i c s. — These are generally semiconsolidated, what gives them often a "concrete-like" appearance. The clasts weather in relief, giving many units a characteristic knobby surface texture. The partial induration of the volcaniclastics is due to clay diagenesis and possibly also to devitrification of clay-size ash; however, it is to be stressed that no traces of vitriclastic texture or shards have been noted in the matrix of these deposits.

The volcaniclastics consists of detrital material which is of plagiorhyolitic composition, transitional to the (rhyodacitic) composition of the volcanics of Mt. Vissoka Elha. Admixtures of nonvolcanic rock fragments are fairly common, but are low and generally not greater than 5—10%. Except for the polymict basal portion of the volcaniclastic pile (see below), relatively few beds have been noted in which foreign fragments are more abundant. It is noteworthy that the suite of the nonvolcanic clasts is identical with that of the "basal polymict gravels" (see below in text).

The individual volcanic clasts vary in their appearance from those which are virtually unweathered and almost fresh, to such clasts which are densely altered to clay on their outside or almost totally argillized; the clayey substance is predominantly mixed-layered montmorillonite-illite. The altered and unaltered clasts are intimately mixed, and are present in almost every bed. However, even the nonargillized clasts show variable degree of weathering; clasts with well-preserved biotite phenocrysts and glass fragments occur beside clasts in which phyllosilicates are totally replaced by clay and hydromicas, and in which volcanic glass is completely devitrified.

Many clasts are vesicular and(or) display well-visible internal fluidal structures, sometimes plastically deformed. These are features suggesting subaerial lava flows or related extrusion phenomena (see also Alexiev and Jurova, 1976). Many clasts display spherulitic internal texture, indicative of near surfacial to subaerial consolidation of the primary volcanic rock. Several others are internally aphanitic, felsitic, or pilotaxitic.

It is noteworthy that only relatively few of the clasts do contain any amphibole phenocrysts. Even when present, the amphibole is usually strongly altered to clay (up to complete pseudomorphoses). This scarcity of amphibole remains in a clear contrast to the amphibole-bearing rhyodacites which actually crop out at Mt. Vissoka Elha. From this fact, and from the already mentioned petrochemical variation there comes a suggestion that the valcaniclastics have been primarily derived through erosion of an upper (plagiorhyolitic) part of the volcanic plug, which was devoid of the amphiboles and which was later entirely eroded. This statement is also supported by several of the afore-mentioned, structural/textural properties of the volcanic clasts.

The clasts are embedded in coarse, usually muddy matrix which consists of lithic/crystal volcanic material, some foreign admixtures (mostly detrital quartz), and abundant clay. The clay is predominantly mixed--layered montmorillonite-illite, with minor amounts of montmorillonite and kaolinite. Among the common components of the matrix are also detritial cristobalite and authigenic carbonate (calcite). Locally present are zeolite heulandite and hydromicas.

*Basal polymict gravels.* As already mentioned, the basal portion of the volcaniclastic pile consists of polymict sandy gravels containing subordinate volcanic clasts in their upper portion. These semiconsolidated deposits are composed of pebbles which lithologically represent quartz and various gneisses (70—80%), some older sedimentary rocks (15—20%), micaceous schists (10%), and also a volcanic rock (1—5%). Sand-size fraction consists mostly of quartz and gneissic detritus.

The gneissic assemblage represents the (?) Archaean gneisses of the nearby Plana Mts. (Fig. 1). The same regards the provenance of the remaining metamorphic rock fragments and subordinate gabbroic pebbles; a considerable gabbroic body is known as present among the abovesaid gneisses at the Pasarel fault, west of Dolni Pasarel (T. Kostadinov *et al.*, unpubl. report, 1973). The pebbles of sedimentary rock, comprising reddish-brown quartzitic sandstones and minor conglomerates, were derived from the Lower Triassic (Bunter) rock series; this was originally much more extensive in the region, and probably covered the aforesaid gneissic complex (Fig. 1). The subordinate volcanic clasts represent an almost aphanitic rock with small biotite flakes, which closely resembles the plagiorhyolite fragments present in the overlying volcaniclastics; most these clasts are weathered, and some are variously altered to clay.

### POLYMICT GRAVELLY COVER

The polymict "Neogene gravels" (Middle-Upper Miocene), which unconformably overlie the volcaniclastic succession, are part of a widespread sedimentary cover. Their composition, grain size and colour are therefore highly variable.

East of Mt. Vissoka Elha, in the vicinity of Stražarski Dol gorge, these polymict gravels and associated sandy deposits are composed predominantly of metamorphic rock fragments (50—90%) derived from the nearby Lozen and Plana Mts.; these are clasts of various gneisses, quartz, quartzites, and some unidentified quartz-feldspar metamorphic rocks. Several beds abound in reddish-brown clasts (10—45%) derived from the Lower Triassic sandstones, while many others contain pebbles (1—5%) derived from the Cambrian diabase-phyllite complex (Fig. 1). Sand-size fraction consists of detrital quartz and gneissic material. Subordinate volcanic clasts are present only at the contact with the underlying volcaniclastics, and have evidently been derived through erosion of these latter.

West of Mt. Vissoka Elha, in the vicinity of Bistritza Creek, the polymict cobble/boulder gravels consist predominantly of rock fragments derived from the Cretaceous andesites, syenites, and monzonites (80—100%). Subordinate are fragments of cordierite hornfelses, andalusite-cordierite quartzites, and cordierite-biotite paragneisses (0—20%), which all have been probably derived from the eastern part of Vitoša Mts. (comp. Dimitrov, 1934). Locally present are also fragments of gabbro, microgabbro and migmatites (0—3%). Volcanic rock fragments are subordinate and present only at the erosional contact with the underlying volcaniclastics.

## SEDIMENTOLOGY OF THE VOLCANICLASTICS

### FACIES A: DEBRIS-FLOW DEPOSITS

Description. These are unsorted, unstratified deposits which occur in extensive sheet-like beds. Rock colour is from yellowish white to light yellowish (or greenish) grey. These diamictons vary in texture from cobble-size paraconglomerates to (subordinate) pebbly mudstones; these latter are usually greenish in colour, and may imperceptibly grade into granule mudstone within one bed. Grain size distributions vary from extremely bimodal to moderately bimodal and polymodal, and this va-

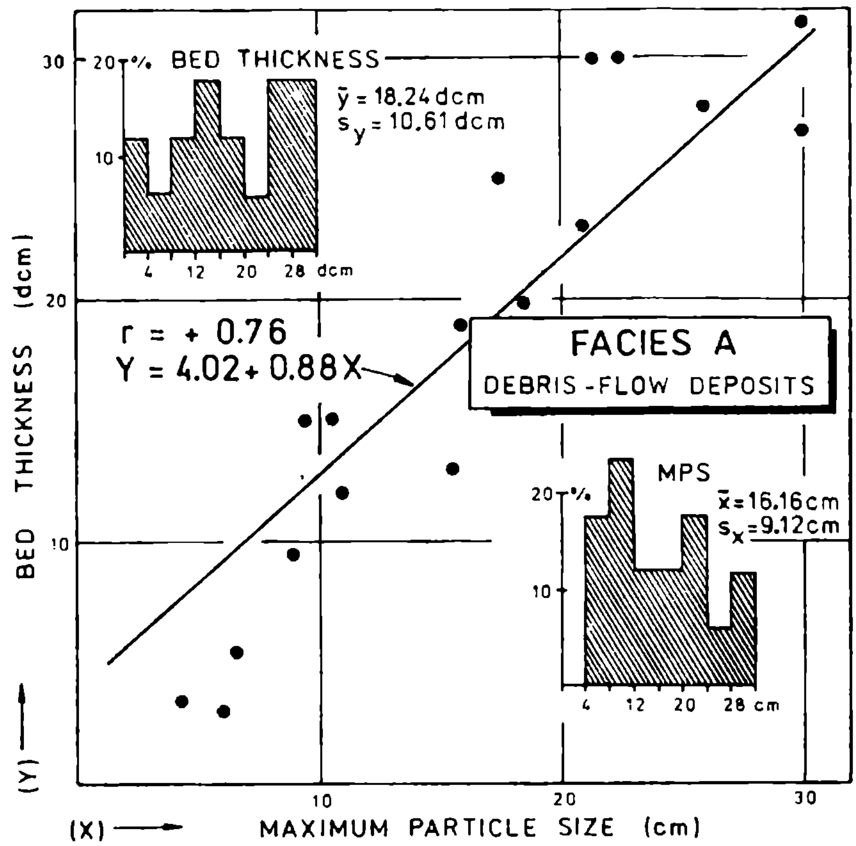



Fig. 5. Data on the bed thickness and maximum particle size (*MPS*) of facies A deposits. The slant line is the linear regression line fitted to the data by least-squares method ($r$ = correlation coefficient); the histograms show the frequency distributions of bed thickness and maximum particle size ($\bar{y}$, $s_y$ and $\bar{x}$, $s_x$ are the mean and standard deviation values of these two parameters, respectively)

Fig. 5. Dane dotyczące miąższości ławic i największej średnicy klastów (parametr *MPS*) w utworach facji A. Ukośna linia prosta na wykresie jest linią regresji, dobraną do przedstawionych danych metodą najmniejszych kwadratów ($r$ = współczynnik korelacji); histogramy ukazują rozkłady częstości miąższości ławic i największej średnicy klastów ($\bar{y}$, $s_y$ oraz $\bar{x}$, $s_x$ to odpowiednio wartości średniej arytmetycznej i standardowego odchylenia obu wspomnianych parametrów ławic)

riation is not uncommonly found among superposedly stacked beds of the same type.

Clasts are mostly angular to subangular, and relatively few show any marked signs of abrasion. Maximum particle size (MPS) of the beds ranges between 4 and 30 cm, with a mean of 16 cm and a standard deviation of 9 cm (see histogram in Fig. 5). The clasts are embedded in an unsorted coarse-grained matrix, and from an unsupported clast framework. The sediment is rich in matrix (commonly more than 60—70%) and the clasts are always matrix-supported. Mud content ranges from 10 to 50%, and is as high as 70—80% in pebbly mudstones.

These deposits abound in angular clasts of volcanic rock which range in appearance from those which are almost fresh and virtually unweathered, to such clasts which are densely altered to clay on their outside or

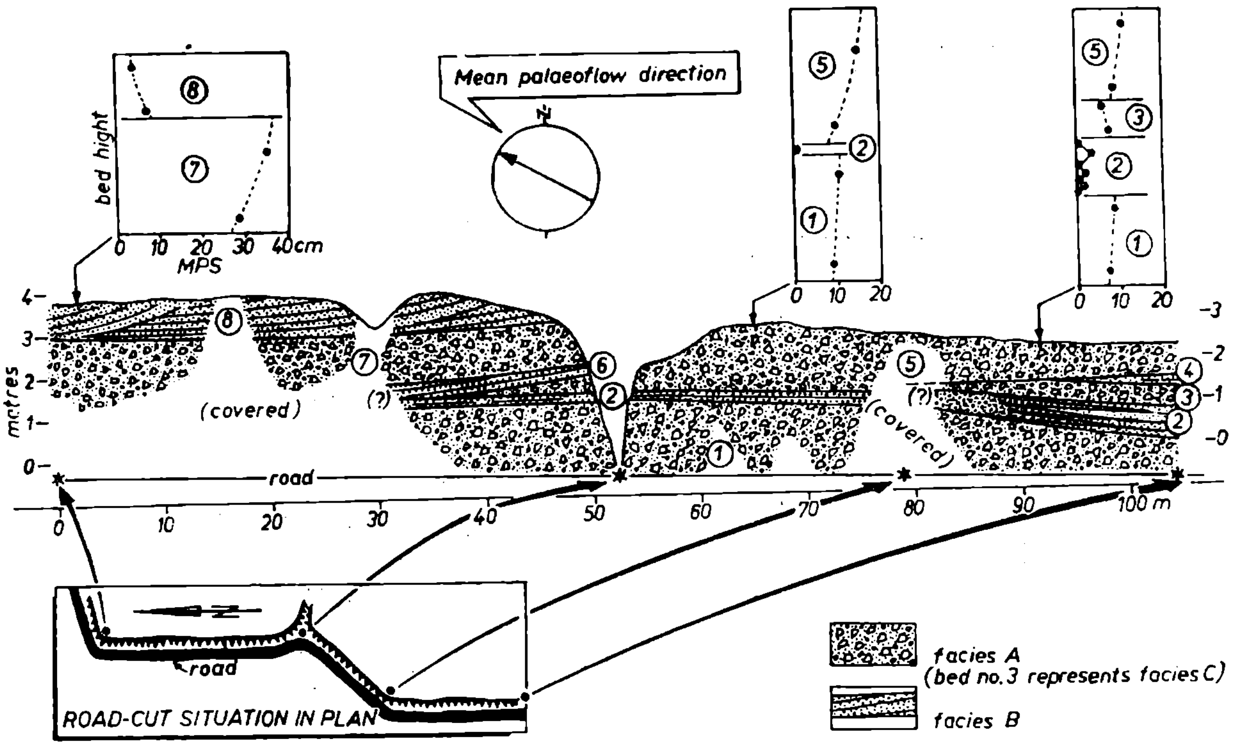



Fig. 6. Roadcut section showing the spatial relationships between facies A and B. (For outcrop location see locality no. 5 in Fig. 2). Above the cross-section are given the mean palaeocurrent direction for the outcrop and the vertical plots of maximum particle size *(MPS)* values; beds are numbered according to the inferred chronology of depositional events in outcrop. Note the vertical exaggeration

Fig. 6. Przekrój geologiczny wcinki drogowej, ukazujący relacje przestrzenne między utworami facji A i facji B. (Lokalizacja odsłonięcia zob. punkt nr 5 na fig. 2). Ponad przekrojem przedstawiono na rysunku średni kierunek paleotransportu oraz wykresy pionowej zmienności parametru *MPS;* ławice ponumerowano stosownie do interpretowanej chronologii wydarzeń depozycyjnych w odsłonięciu. Zastosowano pionowe przewyższenie skali

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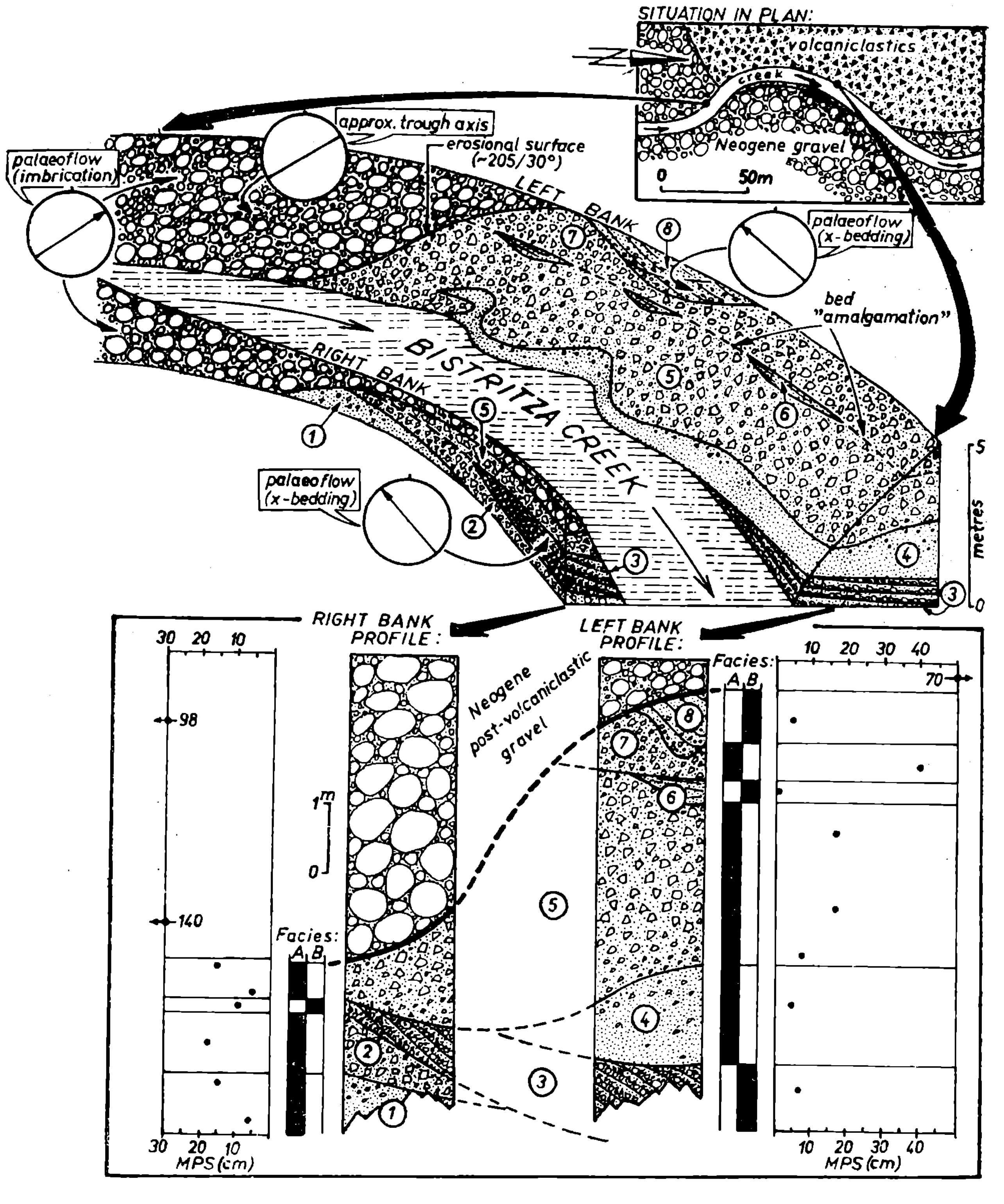
Fig. 7. Diagrammatic section and summary profile showing the spatial relationships between the deposits of facies A and B and the "Neogene gravelly cover". (For outcrop location see locality no. 11 in Fig. 2). Beds are numbered according to the inferred chronology of depositional events in outcrop; beds no. 1, 2, 4, 5, and 7 represent facies A (beds no. 1 and 4 are "mudflow" units), while beds no. 3, 6, and 8 represent facies B (bed no. 3 consists of polymict gravels with minor volcanic clasts). In the upper diagram are shown the mean palaeoflow directions (note the divergence in palaeoflow between the volcaniclastics and the overlying "Neogene gravels"). Lower section represents profiles of the right and left creek banks, and includes the vertical plots of maximum particle size *(MPS)* values

Fig. 7. Blokdiagram i sumaryczny profil, ukazujące relacje przestrzenne między utworami facji A i B oraz „żwirami neogeńskimi". (Lokalizację odsłonięcia przedstawia nr 11




na fig. 2). Ławice ponumerowano stosownie do interpretowanej chronologii wydarzeń depozycyjnych w odsłonięciu; ławice nr 1, 2, 4, 5 i 7 reprezentują fację A (ławica nr 1 i ławica nr 4 to jednostki depozycyjne utworów typu „mudflow"), natomiast ławice nr 3, 6 i 8 reprezentują fację B (ławica nr 3 zbudowana jest ze żwirów polimiktycznych, zawierających podrzędne klasty skały wulkanicznej). Na górnym diagramie przedstawiono średnie kierunki paleotransportu (widoczna jest rozbieżność paleoprądów między wulkanoklastykami a wyżejległymi „żwirami neogeńskimi"). Przekrój u dołu przedstawia profile pionowe prawego i lewego brzegu potoku oraz zawiera wykresy pionowej zmienności parametru *MPS* w tych profilach almost totally replaced by a yellowish-green clayey substance. Large, scattered argillite clasts of this latter appearance are also locally present, and these fragile clasts were evidently emplaced together with the entire mass of sediment and did not weather *in situ*.

The nature of facies A deposits can be seen in Figures 6 and 7. These sections, and all others examined, show the beds of facies A to be extensive, blanket-like bodies which are either parallel or subparallel with each other and also with the adjacent beds of other facies. This observation is supported by those few transverse and longitudinal sections which intersect. The total width of most beds is unobtainable because of either erosion or lack of exposure, but their thickness can be measured accurately.

There is a considerable between-bed variation in both thickness and grain size of facies A. The thickness of the beds ranges between 0.3 and 3.2 m, and has a mean of 1.8 m and a standard deviation of 1 m (see histogram in Fig. 5). In general, the thicker beds abound in coarse material, while the thinner ones are relatively finer grained. Although being generally lower than between-bed variation, the within-bed variation is also significant in many instances. That is probably because many of the facies A beds have been apparently emplaced upon irregular (eroded) surfaces, while the upper surfaces of these beds do not mirror the bottom irregularities and are often fairly flat despite the character of the surface beneath ("mantle bedding"). In several instances the beds of facies A fill broad erosional scours which are floored with the deposits of facies B (see examples in Figs. 6 and 7). There is always an evidence that the scour is itself primarily due to the deposition of facies B, though some additional erosion possibly exerted through the deposition of facies A cannot be confidently excluded in most cases.

With the few exceptions already noted, there is however no compelling evidence that most of the facies A beds entrained any considerable amounts of the underlying sediment during their emplacement. This suggestion is supported by observation in those few convenient outcrops where marked textural and(or) colour differences occur between the beds of facies A and the underlying sediment. For example, the deposits of facies A in some outcrops are found on top of thin, humified portions of underlying beds, which presumably would be easy to erode. Moreover, at the basal contacts of two individual beds of facies A in outcrop no. 5 (for location see Fig. 2) examples of thin (mm) veneers of finely crushed charcoal have been locally found as present immediately beneath the sediment which apparently lacks any charcoal fragments; so the charcoal was evidently grinded during the emplacement of overlying sediment, but has not been incorporated by the overriding flow.

Thus, despite the remarkable thickness some beds of facies A attained during deposition, they evidently did not erode deeply the underlying sediment. This statement is also supported to some degree by statistical inferences about the bed thickness/particle size relationships in facies A.

Following the reasoning of Bluck (1967), the thickness of a bed may be taken as an approximation to the amount of sediment discharged by the depositing agent at the point of measurement, while the maximum particle size (MPS) gives an indication of the agent competence at that point (as argued by Bluck, the true competence of the flow would be obtained some very small distance immediately upcurrent of this point of measurement). Diagrams in Figure 5 show the data on the bed thickness and maximum particle size of facies A, as derived from various superposed beds in a number of outcrops. From these data there appears to be a high positive correlation between the abovesaid two bed-parameters of facies A; the corresponding linear arithmetic function, which may serve as an approximation of this latter relationship, and the frequency distribution of the two parameters are all given in Figure 5.

The dominant bed types of facies A are graphically shown in Figure 8. Despite the general lack of grading in matrix, these beds commonly show a vertical grading of large clasts ("coarse-tail" grading), as it is also illustrated by the data in Figures 6 and 7. The grading is most often of

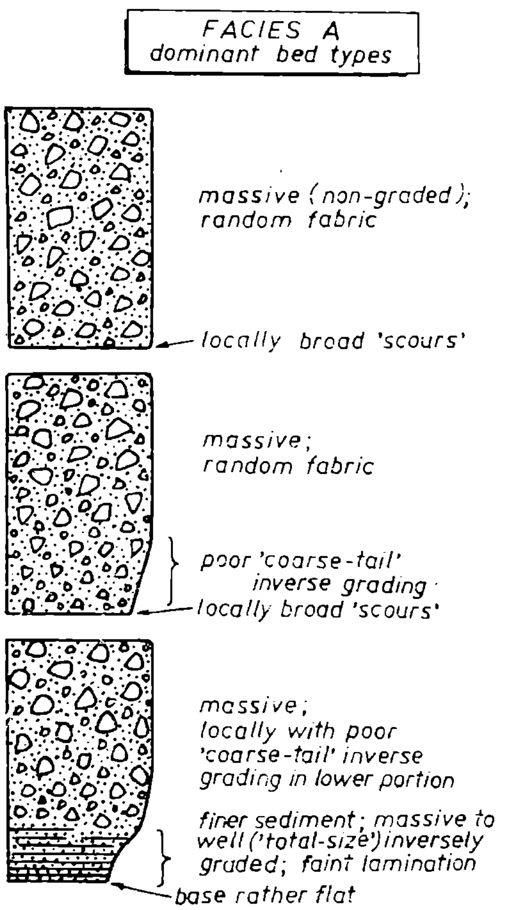



Fig. 8. Predominant bed types of facies A

Fig. 8. Dominujące typy ławic w utworach facji A

inverse type, though examples of normal grading and inverse-to-normal grading are also locally present. In a few instances a lateral change from one grading type to another has been noted within one bed. Most often, however, the graded portion of a bed passes laterally into a virtually ungraded (massive) portion. Several beds appear massive throughout the substantial outcrop, even in a vertical succession of a number of beds at the same locality.

The grading is usually weak, and is therefore most conspicuous where the entire thickness of the bed is exposed. It is often best shown by the lower few decimetres of a bed, to which the grading is not uncommonly restricted; an outcrop of the upper and even middle portion of a bed may show no such grading to the eye.

In a few instances the unsorted beds show presence of thin (15—30 cm) and laterally declining (extent of a few metres) basal portions, which consist of fairly well sorted, sand/granule sediment containing small scattered pebbles and which almost imperceptibly pass upwards into the major, coarse-grained portions of the beds (Fig. 8). The contact between the two portions of a bed is transitional, and the "basal layer" is thought therefore to be an integral part of the bed; this opinion is based on a detailed examination of some representative examples in the outcrops no. 5 and 11 (for location see Fig. 2). The abovesaid "basal layer" does commonly show a faintly developed, horizontal discontinuous stratification and a more or less distinct inverse grading (total-size). Flat pebble-size clasts, when present within it, usually lie with their maximum projection planes parallel to the bed lower boundary.

When superposed directly upon one other, the beds of facies A are often variously amalgamated, and so their contacts are sometimes poorly defined. However, upon close examination of vertical clast size variation, the bedding planes can be always discerned in outcrops; also the presence of thin lenticular units of facies B, which locally separate the amalgamated beds, makes the bed boundaries readily apparent (see example in Fig. 7).

Interpretation. The sedimentary features of facies A, together with the lack of characteristic fluvial structures, all strongly suggest that these poorly sorted diamictons, in which angular pebbles and cobbles are set in an unsorted finer-grained matrix, are the result of debris flows. Mud-dominated beds of facies A correspond the "mudflows" in terminology of many authors. The overall aspect of facies A points into alluvial-fan depositional settings.

For several reasons it is unlikely that the deposits of facies A were laid down by "warm-state" mass flows, *i.e.*, by hot debris-flows (or lahars) directly related to volcanic eruptions. There is evidence suggesting that the mass-flows were most probably caused by the saturation and avalanching of a loose regolith of weathered volcanic material during periods of heavy precipitation, and presumably also by avalanching of volcanic rock which was chemically decomposed by hydrothermal activity (see also further in text). As it is also discussed later, some tectonically triggered landslides are particularly likely here.

The effects of the mass-emplacement of a debris-flow, which is often referred to as "freezing", are well compatible with the data derived from facies A (Fig. 5). High positive correlation between the bed thickness and maximum particle side in facies A implies that there has been a close relationship between the capacity and competence in individual flows, and that the individual beds were emplaced as single events due to "freezing" of the entire flowing mass (*cf.* Bluck, 1967). According to Bluck (*op. cit.*) the high rank of correlation among the size (thickness data does also suggest that there has not been any preferential erosion at the bases of the beds during their emplacement. If these deposits have been derived from the erosion of underlying beds then the correlation between bed thickness and particle size would be low, if any, since the thickness of the eroded bed would bear little or no relation to flow discharge (*i.e.*, such beds would appear "too thin" in relation to their particle size).

A source rock of regolithic type, as evidenced for the volcaniclastics considered (see later in text), is well compatible with the herein postulated mode of mass-flow deposition of facies A. Most modern debris-flows are known to begin as landslides on hillslopes, moving both *in situ* and transported regolith (Vinogradov, 1969; Johnson and Rahn, 1970; Rodine, 1974). The evidence presented by Rodine points into landsliding an the dominant initiating agent, but it has been argued by Vinogradov that slope-wash processes can also supply material to debris-flows. As regards facies A, it seems rather unlikely that each bed of these deposits is the result of one landslide. Most probably the landslides contributed sediment to the channels in alluvial-fan catchments, and, once sufficient debris had accumulated, the addition of water from rainfall initiated a debris-flow (*cf.* Wasson, 1977, p. 790). The mudflow units of facies A did probably originate in similar way, due to the slope-wash supply of fine material which was removed from regolith by rain splash.

The abovesaid mode of processes, in involving a threshold of stability of the material within a channel (Wasson, 1977; see also Bull, 1977), seems to be likely here since stream-flow deposits (which are to be discussed later) are locally found as interbedded with the deposits of facies A. Moreover, most of the "frozen" stream-channels within the volcaniclastic succession appear to be fairly steep and attain an angle of 15°. On the other hand, however, large landslides probably occurred as well, each producing a debris-flow which reached the surface of alluvial apron. The debris-flows, therefore, could well have been the result of either one landslide or a mixture of material from a number of landslides.

The considerations above seem to serve as an adequate explanation of the textural, and partly also compositional heterogenity observed among the beds of facies A in many autcrops. The mixing of material from several landslides, together with the variability inherent in the source material have probably all contributed as a source of variability for the texture and clast properties of the debris-flows considered. Some degree of sediment sorting, or selection, during mass transport is also to be taken into account (*cf.* Bull, 1964, 1977; Bluck, 1964; Hooke, 1967, Wasson 1977), though is difficult to demonstrate here. The individual beds of facies A are difficult to correlate because of the poor and widely scattered outcrops, and so it is impossible to recognize any downflow change in the particle size of a bed; this difficulty in making bed correlations here includes also the well-known fact that successive debris-flows usually occupy various portions of fan surface (Beaty, 1963, 1968; Bull, 1964, 1977; Hooke, 1967; Johnson and Rahn, 1970).

Within a debris-flow the large clasts are rafted along rather gently, slowly shifting position but not being violently tossed about (Middleton and Hampton, 1976, p. 209). This is compatible with the evidence from facies A, in which angular clasts predominate and are scattered-throughout the bed without essential clast-to-clast contacts. The presence of large fragile clasts in facies A points into a fairly passive mode of emplacement (*cf.* Davies and Walker, 1974, Fig. 5; Long, 1977, p. 2506 and Fig. 16), probably within a "rigid plug" formed at some level within the flow, as has been demonstrated in experimental debris-flows by Hampton (1972, Figs . 4 and 5).

Many beds deposited by debris-flow appear therefore to be massive (ungraded), and examples are also present among the beds of facies A. Such beds possibly correspond entirely to "plug flow", and may be ascribed to some most viscous debris-flows (Bull, 1977, p. 237). On the other hand, the presence of poor (coarse-tail) grading in facies A implies that the larger clasts were often able to move with respect to one other within a flow. This type of grading does also imply that the larger clasts were more susceptible to the grading processes, while the finer matrix of a debris-flow remained remarkably homogenous throughout its thickness.

Coarse-tail normal grading, as occasionally found in facies A, can be explained by sinking of large clasts in a debris-flow of relatively low viscosity (Bull, 1977, p. 237); due to the low viscosity of matrix the body forces of larger clasts probably start to overcome the actual strength/ buoyancy support in a flow, and so gradually sink down accordingly to their weight.

Coarse-tail inverse grading, which is common in facies A, is more disputable as to its origin, since may be probably due to a number of factors. *Firstly,* an important agent causing grading of this type in a debris-flow may be the flotation of the buoyed clasts. Since clast with a large surface area could be supported by a relatively weak matrix (Hampton, 1975), it seems not unrealistic that the large clasts might undergo some degree of flotation when being transported in a high-viscosity matrix.

*Secondly*, the dispersive pressure arising from the clast interactions might have considerably aided to the matrix strength and buoyancy in supporting the clasts above the bottom, and thus possibly also in the overcoming of their body forces. As stated by Hampton (1975), most real debris-flows are probably "combination debris-flow/grain-flows" in the sense of Middleton and Hampton (1973); thus, it seems likely that the deposits of facies A were probably laid down by debris-flows transitional to density-modified grain flows (cf. Lowe, 1976; see also behaviour of "slurry flow" in Carter, 1975). Moreover, water mixed into a debris-flow at the base (as it usually takes place during the debris-flow movement along an active segment of stream channel) might have reduced the effective density of the flow, and thus increased the possibility for clast collisions in the lower portion of a flow. This seems to be keeping in with the data from facies A, where the coarse-tail inverse grading is always better developed in, or even entirely restricted to the lower portion of a bed.

*Thirdly*, it has been postulated by Hampton (1975, p. 843) that "the zone occupied by the rigid plug and the thickness of the plug are time--varying properties, and it is conceivable that every level within a particular flow was sheared for some time during the life of flow". At any given time a plug of some thickness exists in the flow, but changes in flow velocity, channel slope and configuration may probably produce enough migration of the plug lower boundary to at least temporarily obtain shear at high levels. Consequently, inverse grading caused by shearing may be probably developed almost throughout the flow thickness for some time, and may so be preserved by the rapid "freezing" of a debris-flow. Certainly, a combination of the abovesaid three possibilities is also likely here. According to Walker (1975), there is probably also a relationship between the inverse grading development and steep depositional slopes.

The afore-mentioned shear forces, as exerted at the base of a moving debris-flow, make also explicable the characteristics of the "basal layer" which is locally present in the beds of facies A. The basal zone of shear, which develops beneath the rigid plug, is known to be characteristic of debris-flows (Johnson, 1970; Hampton, 1975; Middleton and Hampton, 1976). The vague horizontal lamination and parallel alignment of flat clasts within this zone are due to the laminar fashion of sediment flow beneath the plug. The relatively fine grain size and the total-size inverse grading in the basal layer are, however, variously explained by different authors.

Most commonly these latter features are interpreted as the result of the dispersive forces which act normal to the flow lower boundary, and which cause the large particles to shift upwards, toward the zone with the least rate of shear (Bagnold, 1954, 1955). Middleton (1970, p. 267), instead, proposes that inverse grading develops because "the smaller particles tend to fall into the spaces between the large particles and thus displace the large particles toward the surface". Fisher and Mattinson (1968), and Mattinson and Fisher (1970) explain inverse grading in terms of the lift forces due to rotation caused by differences in velocity at the top and bottom of a clast; this lift is a function of particle diameter and velocity gradient, but Southard's (1971) experiments suggest that its magnitude is very small.

Hampton (1975, p. 842) has postulated that grain size is uniform through the basal zone of shear if debris-flow competence is independent of velocity, but in vertically variable (graded) if competence is variable at low velocities. The basal layers in facies A beds are most often vaguely graded, and so this second mode of flow behaviour is preferred here. The lateral discontinuity of the basal layers, as observed in facies A, is not surprising since the position of the plug lower boundary is laterally variable in a flow and depends upon several factors; among them are the slope angle and configuration as well as anything else which controls the internal shear stress, or "driving" stress, of a debris-flow (for details see Hampton, 1975). There is no reason, therefore, to expect the basal layer to be laterally persistent or developed in each bed.

Also the basal relationships of facies A beds are by no means unusual. It has been emphasized by many authors (*e.g.*, Fisher, 1971; Wasson, 1977) that debris-flow deposits commonly overlie easily eroded sediments with little or no features of erosion. Smoothing of the topography, often referred to as "mantle bedding", is also among the characteristics of debris-flow emplacement (*e.g.*, Crandell, 1971; Wasson, 1977, 1979).

In a few instances the deposits of facies A have been observed to be inseted, either as a single concave-up bed or as two superposed beds of similar geometry, within some broad, few-metres deep channels which are floored with a veneer of fluvial gravel and capped (smoothed) by a thick, extensive unit of facies A. These inset beds bear resemblance to debris--flow erosional remnants, and so their concave-up geometry might be possibly suspected to be the result of erosion ("downcutting") exerted at the bases of successive debris-flows. However, it has been emphasized by many authors that some conventional inferences which can be made with regard to the aggradation and downcutting of alluvial channel-fills are not pertinent to debris-flows (among the numerous others, some illustrative examples are given by Crandell, 1971, p. 7—8). It has been also suggested earlier in text that the debris-flows did not generally exert strong erosion at their bases.

Accordingly, the abovesaid bed relationships are thought to have been due to the well-known propensity of many debris-flows to travel down along the channels and to leave only a thin deposit on the channel sides and floor to mark their passage (*e.g.*, Crandell, 1971, p. 7; Wasson, 1977, p. 791—792). The concave-up, veneer-like, inset beds of facies A are probably the result of such phenomenon.

### FACIES B: STREAM-FLOW DEPOSITS

Description. This facies consists primarily of well bedded, clast--supported pebble/cobble size-grade gravels with minor interbeds of coarse sand and(or) pebbly sand. Sandy beds are also locally present, and these are usually muddy and contain various amounts of granule/fine--pebble size fraction. Rock colour is from light grey to yellowish grey. Mud content in beds varies from a few to several per cent, and exceptionally attains 20—30% in certain sandy units. In contrast to the remaining facies considered, the deposits of facies B generally display good separation of size fractions.

In the study area the deposits of facies B are volumetrically subordinate to those of either facies A or C. They are present as isolated beds, or as relatively thin bed sequences, which separate and(or) randomly replace the units of remaining two facies. The deposits of facies B are present in the outcrops both east and west of Mt. Vissoka Elha (Fig. 2), but are relatively more frequent in this former direction.

Thickest sequences of facies B are present at the base of the entire volcaniclastic succession, where they form its "polymict basal portion" (see "General stratigraphy" earlier in text). In this portion the beds of polymict gravel and pebbly sand, containing subordinate volcanic clasts, are vertically stacked into a sequence which is 7 m thick in the borehole profile of Kokalyane (Fig. 3; for location see Fig. 2) and presumably attains a few tens of metres in the section of Stražarski Dol gorge (Fig. 9; for location see Fig. 2); neither the lower nor upper boundaries of this sequence are exposed at the last-named locality, and so its thickness can only be suspected to attain about 50—70 m there. In both of these "basal" occurrences the deposits of facies B attain exceptional thicknesses and appear to be fairly well consolidated to semi-consolidated. Higher up in the succession, however, they are always semi-consolidated irrespective of their compositional variability.

The deposits of facies B are generally more abundant in nonvolcanic rock fragments than are the volcaniclastics of either facies A or C. They vary in their composition from polymict gravels, through clastics with preponderance of volcanic debris, to those sediments which are identical in composition with the major volcaniclastic of facies A and C. In ge-

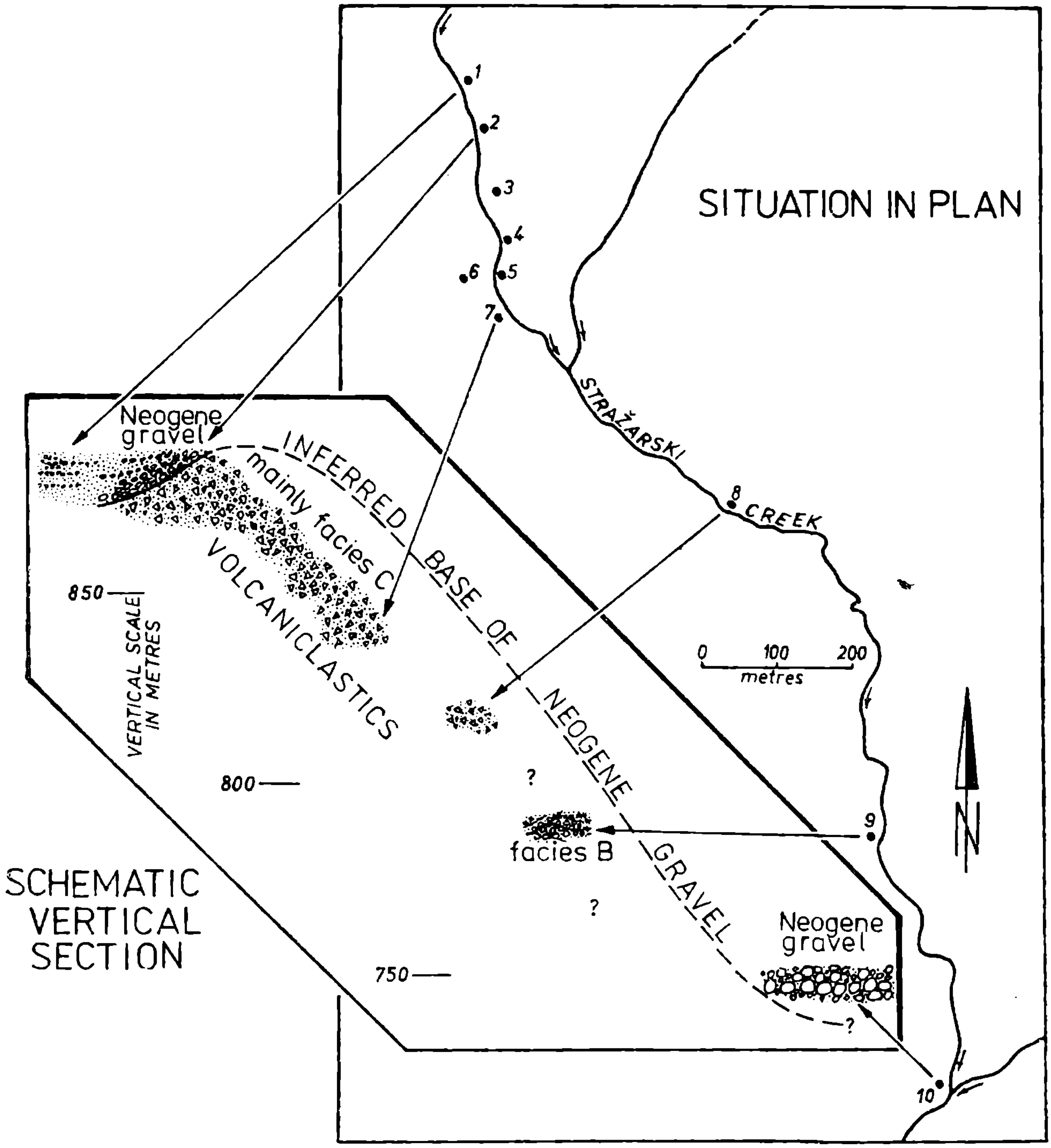



Fig. 9. Schematic vertical section through the outcrop belt at Strażarski Dol gorge, showing the inferred spatial relationships between the volcaniclastic pile and the overlying "Neogene gravels". (For location see locality no. 1 in Fig. 2). Note the high relief of the erosional base of "Neogene gravels"; the cobble/boulder gravels in locality no. 10 are thought to be a lateral equivalent of the pebble/cobble gravels and sands in localities no. 1 and 2 (for details of these latter localities see Fig. 20)

Fig. 9. Schematyczny pionowy przekrój przez pasmo odsłonięć w wąwozie Strażarski Doł, przedstawiający interpretowane relacje przestrzenne między serią wulkanoklastyków a wyżejległymi „żwirami neogeńskimi". (Lokalizacja zob. wystąpienie nr 1 na fig. 2). Zwraca uwagę silnie zróżnicowany relief erozyjnej powierzchni spągowej „neogeńskich żwirów"; skrajnie gruboziarniste żwiry w odsłonięciu nr 10 stanowią najprawdopodobniej odpowiednik boczny gruboziarnistych żwirów i piasków w odsłonięciach nr 1 i 2 (szczegóły tych dwóch ostatnich odsłonięć zob. fig. 20)

neral, the abovesaid compositional trend can be traced vertically, through the successive occurrences of facies B in the profile of the entire volcaniclastic succession considered (when viewed from its base upwards); however, gravelly beds with preponderance of nonvolcanic clasts are also locally present higher up in this succession, at higher stratigraphic levels.

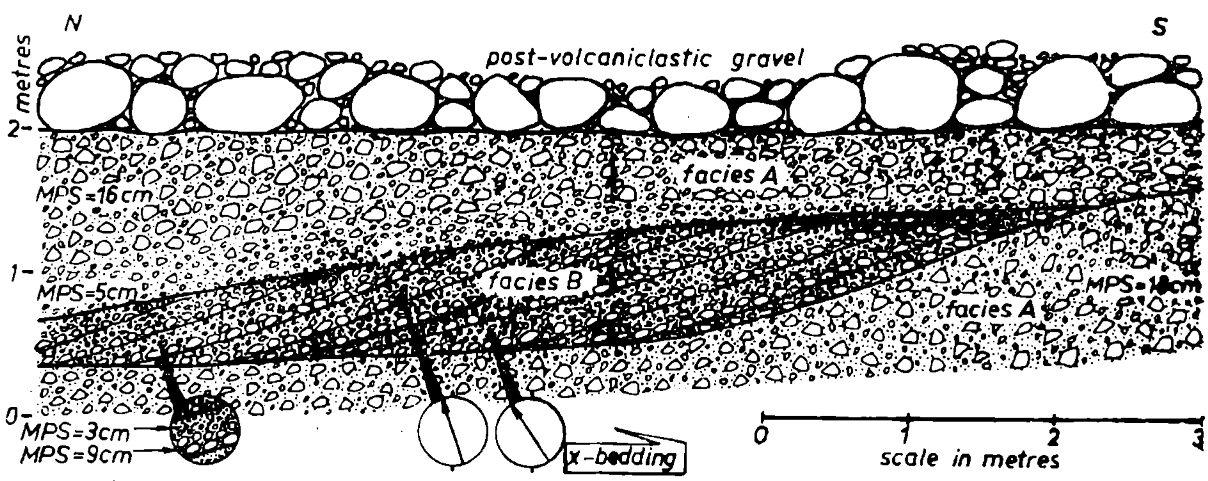



Fig. 10. Example of an isolated unit of facies B present between the beds of facies A. (For location see Bed no. 3 in the right-bank section in Fig. 7). The cross-stratified unit of facies B is of polymict composition, and its individual laminae often display well-developed normal grading (see detail at the bottom left); note the maximum particle size (*MPS*) values enclosed in the section, and the mean palaeocurrent directions presented at the bottom

Fig. 10. Przykład izolowanej ławicy facji B wśród utworów facji A. (Dla lokalizacji zob. ławica nr 3 na przekroju prawego brzegu strumienia na fig. 7). Ta przekątnie warstwowana ławica zbudowana jest ze żwirów polimiktycznych, a jej indywidualne laminy wykazują często dobrze wykształconą, normalną gradację w kierunku prostopadłym do rozciągłości lamin (zob. szczegół u dołu z lewej strony); na rysunku zamieszczono wartości liczbowe parametru *MPS* oraz średnie kierunki paleoprądów

The deposits of facies B occur either as lenticular gravelly beds (often less than 1 m thick) inseted into scour channels which are cut into extensive units of the other facies (Figs. 10 and 7), or as sandy beds (either single or compound) which are laterally more persistent and sometimes almost parallel to one other (Fig. 6). Individual units lens out, both across and downcurrent, over a distance ranging from a few metres to more than 70 m. Thin (10—60 cm) lenses of muddy sand, pinching out and displaying lateral extent of less than 1—2 m, are also locally present, and these separate the superposed, extensive units of facies A (for example see bed no. 6 in Fig. 7).

Every bed of facies B is erosionally based, and the scoured erosion surfaces are abundant in this facies. They are sometimes almost flat, laterally extensive and lack steep sides (Fig. 6), but more commonly are concave-up and show cross-cutting relationships (Fig. 11).

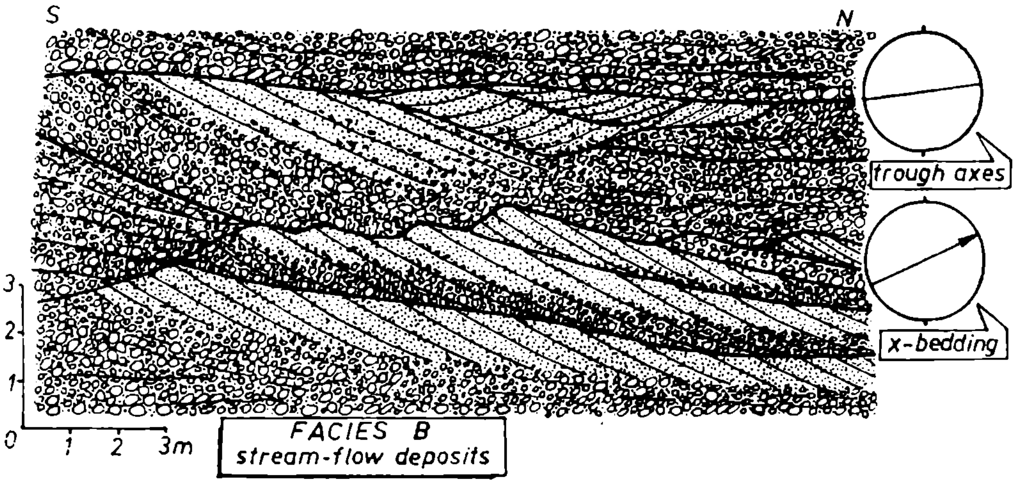



Fig. 11. Sequence of facies B deposits, as developed near the base of the entire volcaniclastic succession. (For outcrop location see locality no. 9 in Fig. 9). These deposits represent the "polymict basal portion" of the volcaniclastic succession (see details in text). Note the abundance of erosional surfaces and their cross-cutting relationships, and the rapid lateral changes in lithology in outcrop. The diagrams show mean palaeocurrent directions as determined in the outcrop

Fig. 11. Sekwencja utworów facji B, występujących w pobliżu spągu całej rozpatrywanej serii wulkanoklastyków. (Lokalizacja zob. odsłonięcie nr 9 na fig. 9). Utwory te reprezentują „polimiktyczną partię spągową" serii wulkanoklastyków (szczegóły omówiono w tekście pracy). Widoczne są liczne powierzchnie erozyjne oraz nagłe zmiany boczne w litologii utworów. Diogramy przedstawiają średnie kierunki transportu ustalone w odsłonięciu

Irrespective of their variable geometry, the majority of the facies B beds display a more or less distinct internal stratification. Coarser grained deposits often show sub-horizontal to horizontal, discontinuous stratification composed of laminae of mixed sand and gravel. The laminae are usually from 5 to 20 cm thick, and display normal grading in the direction perpendicular to their lower boundaries. Imbrication is often well developed, showing an upcurrent dip of pebble intermediate axes, *i.e.*, representing the *a*(t)*b*(i) fabric in Walker's (1975, p. 137) notation. The horizontal discontinuous stratification may sometimes grade upwards and(or) laterally, either downcurrent or across, into lenticular stratification (*sensu* Picard and High, 1973, p. 150); such change is always accompanied by a concurrent increase in the separation of size fractions and grain-size contrasts between the adjacent (lenticular) laminae. Low-angle foreset cross-stratification and large-scale trough cross-stratification are also fairly common in the gravelly deposits. Relative abundance of these sedimentary structures is variable, and locally one type may dominate.

In several instances the gravelly sediment occurs as thick (often high--angle) foresets adjacent to channel banks, and gravel layers are lens--shaped near channel centers; these lenses are unstratified or vaguely horizontally stratified, and are often flanked by sand. Sand occurs in lows flanking gravel lenses which it commonly succeeds vertically in well-

-defined, broad, shallow channels, and as thick masses up to several decimetres thick, in which channel boundaries are not recognizable.

Most sand deposits are characterized either by foreset and trough cross-stratification, or more often by crude horizontal stratification. Simple foreset units have planar to concave-up bounding surfaces which converge either upcurrent or (locally) downcurrent (Figs. 10 and 11), or assume a sigmoidal shape with low-angle lower and upper bounding surfaces (Fig. 10). Erosional surfaces are commonly associated with the cross-stratification, which is locally so eroded that only a thin (remnant) layer is left (Fig. 11); an originally larger set thickness is indicated by thick individual cross-strata.

Many of the extensive sand units are vaguely horizontally stratified. These usually consist of medium-grained, muddy sand with common lenticular interlayers of coarse sand with granules and locally with small pebbles. Individual layers are 5 to 15 cm thick, and persist laterally for a few metres before pinching out. Most pinchouts apparently are the result of non-deposition, but some are the result of subsequent erosion. The attitude of lenticular layers is commonly horizontal, but is inclined at a low angle near a channel margin (for examples see Fig. 6).

In outcrops no. 1 (locality 9 in Fig. 9) and no. 5 (Fig. 2), fine charcoal fragments are locally present as thin (mm) veneers between certain horizontal layers of the abovesaid type. In this latter outcrop (no. 5), which presumably represents an uppermost segment of the volcaniclastic succession considered, certain muddy-sand units of facies B display greyish brown or light yellowish brown colour in their topmost portions; these darker horizons are variously humified (duff), and may locally show weak palaeosol features.

Both bed thickness and particle size of facies B are highly variable from bed to bed. Bed thickness ranges between 30 and 400 cm, with a mean of 140 cm and a standard deviation of about 90 cm. Maximum particle size (MPS) of the beds varies between 0.5 and 18 cm, with a mean of about 8 cm and a standard deviation of 5 cm.

The thickness of facies B beds is plotted against maximum particle size in Figure 12. This plot is not however to be compared with that in Figure 5, since the thickness of facies B beds has been taken as the entire thickness of a unit of sediment bounded by some prominent erosion surfaces in outcrop; such a unit may comprise either a single set of strata or a coset, and thus may comprise in fact many pulses of sediment transportation and erosion. In facies B the equivalent of "bed thickness" as used in the previously discussed facies A is probably the thickness of individual strata (laminae). It is also not possible to make meaningful measurements of the maximum particle size and thickness of individual units, since most of the deposits have suffered contemporaneous erosion and there are few recognizable uneroded units of single depositional

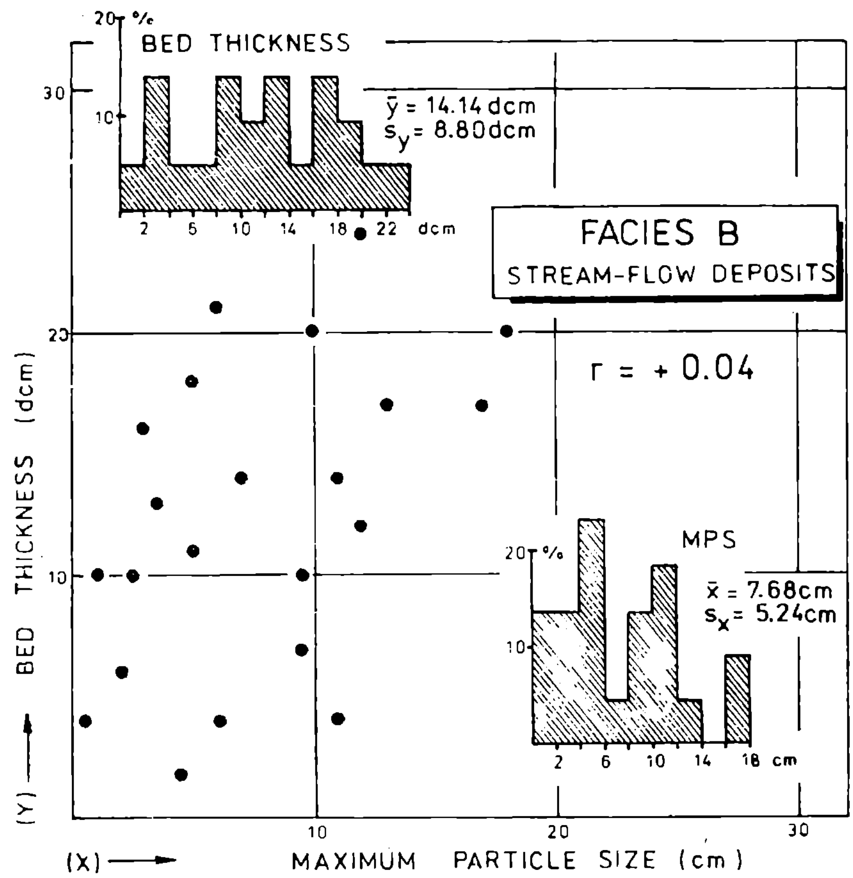



Fig. 12. Data on the bed thickness and maximum particle size (*MPS*) of facies B deposits. The histograms show the frequency distributions of bed thickness and maximum particle size ($\bar{y}$, $s_y$ and $\bar{x}$, $s_x$ are the mean and standard deviation values of these two parametres, respectively); $r$ = correlation coefficient value

Fig. 12. Dane dotyczące miąższości ławic i największej średnicy klastów (parametr *MPS*) w utworach facji B. Histogramy przedstawiają rozkłady częstości miąższości i największej średnicy klastów ($\bar{y}$, $s_y$ oraz $\bar{x}$, $s_x$ to odpowiednio wartości średniej arytmetycznej i standardowego odchylenia obu tych parametrów); $r$ = współczynnik korelacji liniowej

phase. Maximum particle size has been taken from the coarsest portion of a given unit and thus, because of the compound nature of this latter, is not related to flow competence but rather to the maximum flow competence during a longer period of deposition (*cf.* Bluck, 1967, p. 150).

However, in spite of these sources of error, there appears to be an increase in the thickness of facies B units with increasing particle size of the coarsest sediment present during a given depositional phase (Fig. 12). This relationship is however weak, and much less distinct than in the facies A. In statistical terms, the linear correlation between the abovesaid two parameters in facies B is low ($r$ = + 0.04) and statistically non-significant.

There is a variety of structures of directional significance in facies B, and the following have been used as palaeoflow indicators: scour channels (both small and large, if measurable in outcrop), clast imbrication, and cross-stratification (trough axes and directions of foreset dip). The palaeoflow directions derived from the outcrops east of Mt. Vissoka Elha (Fig. 2) appear to be just opposite to those determined in the outcrops located to the west; these data are to be discussed further in text. It is noteworthy that in either of these areas the directional dispersion is low and generally does not exceed 60°. This is even much lower in individual outcrops, where directional discrepencies (though variable for individual structures) are commonly less than 20°.

Interpretation. The deposits of facies B are characteristically water-laid, and are interpreted here as the result of stream-flow processes.

The abundance of concave-up, cross-cutting erosional surfaces and the characteristic assemblage of sedimentary structures, together with the isolated nature of the particular units of facies B, all point into an environment of rapidly shifting, presumably ephemeral channels in which bedload deposition often occurred under upper flow regime conditions (*cf.* Simons *et al.*, 1965). These are characteristics of "flashy" (variable discharge), possibly ephemeral braided streams, which have rapidly migrating channels within which there takes place much contemporaneous erosion and deposition (Krigström, 1962; Miall, 1977; Collinson, 1978). The great bulk of the fluvial deposits under consideration belong to Allen's (1965) "substratum", or in-channel, deposits, with the gravels probably representing flash-flood deposition; no "topstratum" (or overbank) sediments have been recognized, though there are units of muddy sand which apparently are the product of lower energy fluvial sedimentation, probably during the wanning stages of certain floods.

Most of the gravelly deposits of facies B probably comprise multistory longitudinal-bar accumulations, which are characteristic of many braided streams (for recent review see Miall, 1977). This suggestion is supported by observation in those few convenient (three-dimensional) outcrops in which several gravel layers have been recognized to form elongate bodies parallel to transport direction. Deposits of gravelly longitudinal bars are known to be either unstratified ("structureless") or characterized by thin to thick, often crude, horizontal strata which may grade across and downcurrent into low- to high-angle foreset cross-strata (Ore, 1963; Hein and Walker, 1977).

The vaguely horizontally or sub-horizontally stratified and unstratified gravels of facies B were probably formed as aggraded sequences of diffuse gravel sheets, which are characteristic of longitudinal bars (Hein and Walker, 1977). Foreset cross-stratified gravels, as locally associated with these former deposits, were probably formed from bar avalanche faces, while the structureless gravels were probably related to riffle margins of the bars.

Hein and Walker *(op. cit.)* state that of crucial importance to the type of stratification in a longitudinal gravelly bar is the rate at which the bar aggrades versus the rate of its downstream migration. If the water and sediment discharge is high, the diffuse sheet lengthens downstream faster than it aggrades, and no angle-of-repose slip face is developed; and so the downstream depositional margin of a bar is essentially a riffle. In such instance the resulting deposit would be unstratified, or show crude horizontal to slightly inclined stratification. If, however, the rates of sediment discharge are much lower, the diffuse sheet aggrades vertically rather than is swept on downstream; as the bar grows in high, gravel is rolled across the top and tends to avalanche down the foreset. Thus, Hein and Walker believe that lower rates of sediment discharge are more likely to result in deposition of foreset cross-stratified gravels; from this it is inferred here that relatively high rates of sediment discharge did often prevail during the deposition of facies B. However, Collinson (1978, p. 21) suggests that gravel size grade may also be important to the development of either a slip face or a riffle at the downstream margin of a longitudinal bar; flow velocity may be probably added as an important factor too.

The gravelly deposits of facies B are thought to have been laid down by a network of braided distributary channels which were active on the surface of alluvial fan(-s), and in which high sediment discharge generally prevailed. These settings were probably due to the common accessibility of easily eroded material of a wide grain-size spectrum to any one flash flood (comp. grain size of facies A earlier in text). Flash floods probably resulted in surges of sediment-laden water within the ephemeral stream channels; in effect, channels became rapidly filled with sediment, and then shifted a short distance and became abandoned. The longitudinal gravelly bars in channels were probably formed by rapid flow conditions during flood crest, since upper flow regime (antidune phase) transport is known to be common on such bars (Boothroyd, 1972). The imbrication of *a*(t)*b*(i) type, as noted in facies B, is also known to be characteristic of longitudinal gravelly bars (Boothroyd and Ashley, 1975, p. 203).

Most of the gravels and sands of facies B were deposited contemporaneously in individual units. Sand occurs as thin sheets atop gravel layers, as shallow channel-fills within the gravel units, and as thicker units lateral to or downstream from the gravel masses. Such lensing of gravel and sand is particularly characteristic to braided-stream deposits (Allen, 1965; Bull, 1977). Vertical recurrence of either gravel or sand does not necessarily imply a change in sediment availability, stream gradient, or discharge; rather it records only a lateral migration of contemporaneous depositional centers essentially within one flow regime.

The same regards the lenticular stratification which is present within many of the mixed, sand-gravel units of facies B. This type of stratification forms as a result of shifting depositional centers, and is known to be prevalent in braided streams (Picard and High, 1973, p. 152; Bull, 1977, Fig. 6). Unlike beds, however, lenticular stratification does not record the steady and progressive change in conditions within one flow regime; rather it records laterally fluctuating regimes, as from low-relief bar to shallow channel and back again, resulting in thin lenses of sediment deposited atop one other (Picard and High, *op. cit.*). The units with lenticular stratification in facies B did probably result during stages of low bar-and--channel topography, with chaotic and rapid shifts of the shallow anastomosing channels; such topography develops during falling stages on the top of emergent bars, or within broad, shallow, short-lived water flows in channels (Collinson, 1978, p. 21, Fig. 3.7; Bull, 1977, p. 232, Figs. 6 a, b).

Flow through certain channels was sluggish in comparison to many others. This is because several channels appear to have been not always scoured and filled during the same depositional phase, as indicated by the frequent occurrence of a muddy sand floor suggesting that sluggish to stagnant conditions often occurred following scour of a channel. Such broad channels are not uncommonly floored with only a thin (10—30 cm), though laterally persistent, inset layer of muddy sand with granules which is directly overlain by thick debris-flow unit of facies A filling the channel and often extending out of it (see beds 6—7 in Fig. 6, and beds 3—5 in Fig. 7).

Small, isolated lenses of muddy sand with granules, which locally separate debris-flow units (see beds 6 in Fig. 7), are interpreted in terms of reworking and winnowing of the upper parts of earlier deposited debris-flows through the action of shallow, anastomosing water flows. Some of these discontinuous layers of unstratified to vaguely laminated sediment are thought to be the result of sheetwash (*cf.* Wasson, 1979, p. 40; Moss *et al.*, 1979), while several others are probably due to "rillflow" in Bull's (1977, p. 229) terminology. The deposition of these units might have been sometimes a raindrop-stimulated phenomenon (Moss *et al.*, 1979).

The nature of facies B and its close association with extensive debris--flow deposits of facies A, all point into alluvial-fan depositional settings (*cf.* Bull, 1977). Thus the overall interpretation of facies B requires deposition within braided (or low-sinuosity) distributaries, while avulsion probably allowed the emplacement of debris-flows directly onto the water-laid, in-channel sediments (*cf.* Wasson, 1979, p. 34). The vertical succession of stream-flow and debris-flow deposits indicates that channel traction processes were suddenly replaced by mass-flow sheets with lateral dimensions very much greater than any of the fluvial channel--fills. The debris-flow units are not significantly reworked, suggesting that they were deposited over, or in inactive channels. This statement is in keeping with the mode of behaviour of braided distributaries on the surface of many alluvial fans, where stream channels rapidly change position by avulsion; this results in frequent abandonment of areas of traction load sediments, which are then covered by subsequent mass-flows rather than eroded or swamped by overbank deposits. The humified portions of certain mud-rich channel-fills suggest that there must have been sometimes a substantial delay between the channel scouring and subsequent amplacement of overlying debris-flow.

### FACIES C: SHEET-FLOOD DEPOSITS

Description. These are poorly sorted deposits which occur in beds often 70—90 cm thick, with bed-thickness range of 10—150 cm and a mean of 87 cm; maximum particle size (MPS) of the beds varies between 1.5—8 cm, with a mean of 4.65 cm (see histograms in Fig. 13). A significant positive correlation has been recognized as present between the bed thickness and maximum particle size. This relationship and the corresponding arithmetic function, together with the frequency distributions of the abovesaid two parameters are all given in Figure 13.

Rock colour is from greyish-white and light yellowish-grey, to light grey and greenish-grey. Bedding is usually well defined, being made

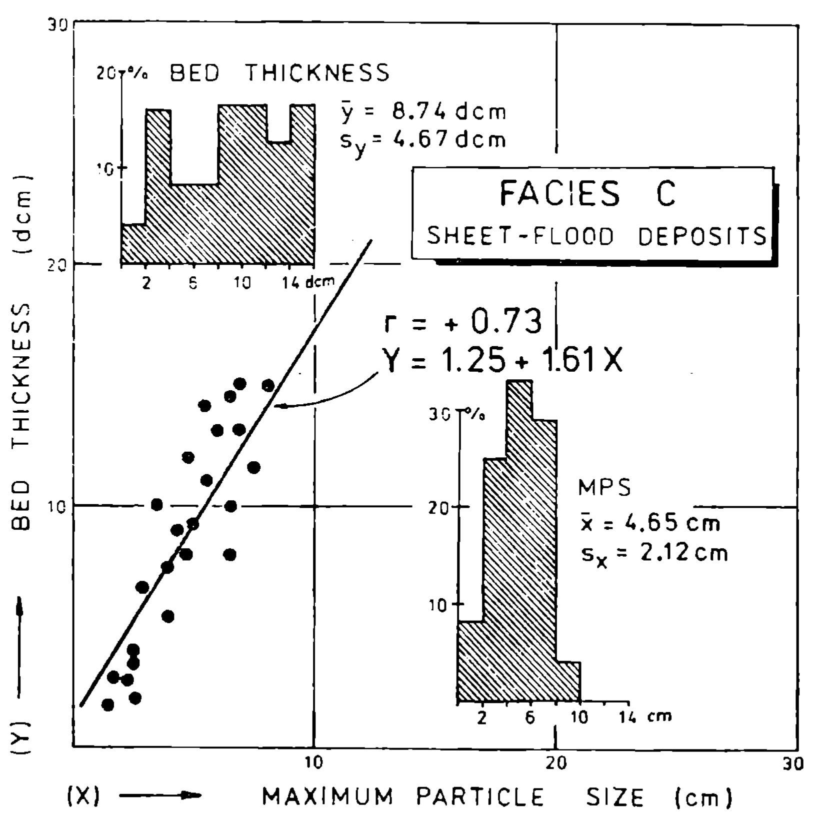



Fig. 13. Data on the bed thickness and maximum particle size (*MPS*) of facies C Explanations as in Figure 5

Fig. 13. Dane dotyczące miąższości ławic i największej średnicy klastów (parametr *MPS*) w utworach facji C. Objaśnienia jak na fig. 5

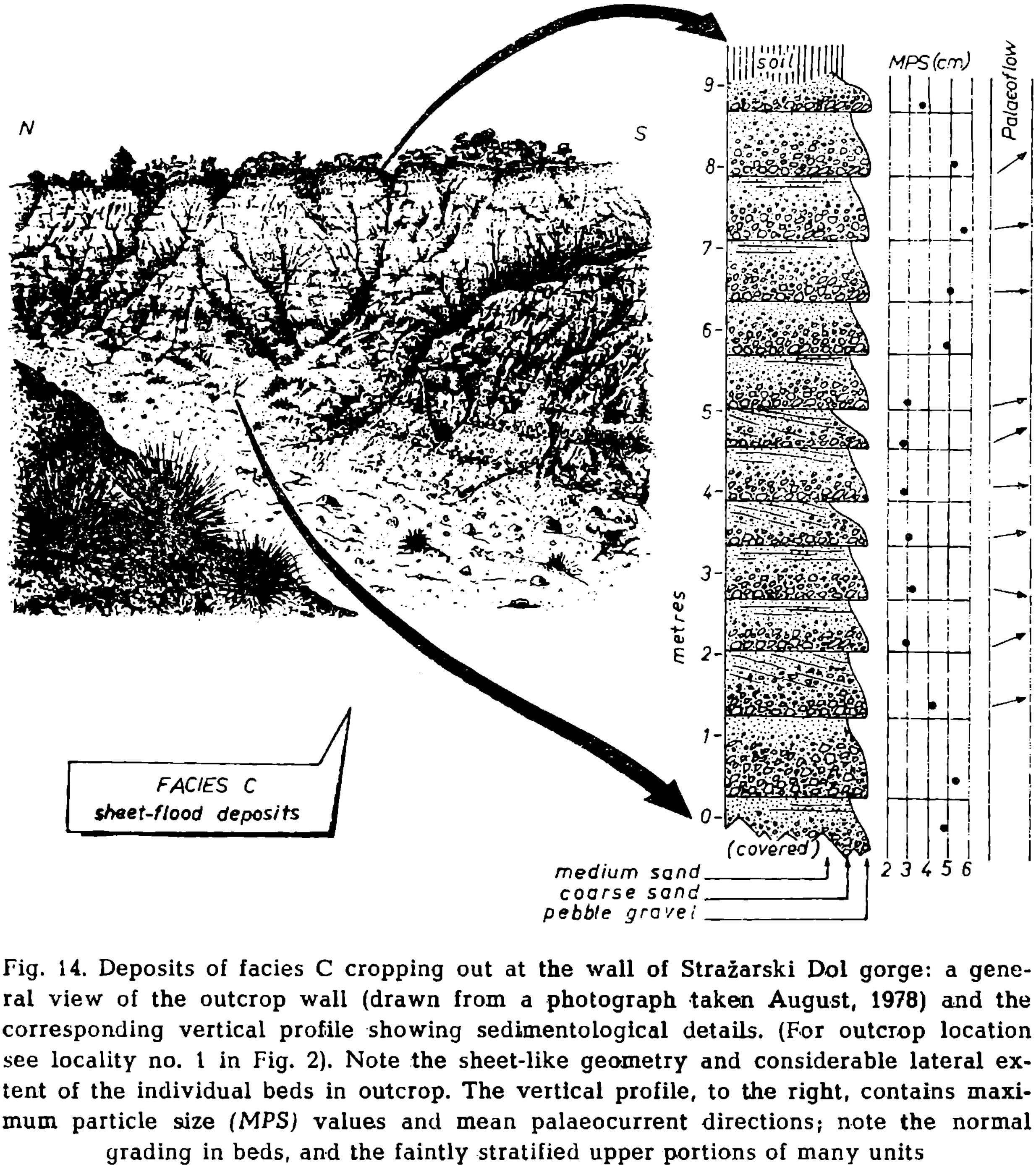



Fig. 14. Deposits of facies C cropping out at the wall of Strażarski Dol gorge: a general view of the outcrop wall (drawn from a photograph taken August, 1978) and the corresponding vertical profile showing sedimentological details. (For outcrop location see locality no. 1 in Fig. 2). Note the sheet-like geometry and considerable lateral extent of the individual beds in outcrop. The vertical profile, to the right, contains maximum particle size (*MPS*) values and mean palaeocurrent directions; note the normal grading in beds, and the faintly stratified upper portions of many units

Fig. 14. Utwory facji C odsłaniające się w ścianie wąwozu Strażarski Doł: widok ogólny odsłonięcia (rysunek z fotografii wykonanej w sierpniu 1978 r.) oraz profil pionowy ukazujący szczegóły sedymentologiczne. (Lokalizacja zob. wystąpienie nr 1 na fig. 2). Widoczna płaskorównoległa geometria ławic oraz ich znaczna rozciągłość boczna. Profil pionowy (na prawo) zawiera też wartości parametru *MPS* i wskazania średniego kierunku paleotransportu; zwraca uwagę obecność normalnej gradacji w ławicach oraz występowanie mało wyraźnej laminacji w ich partiach stropowych

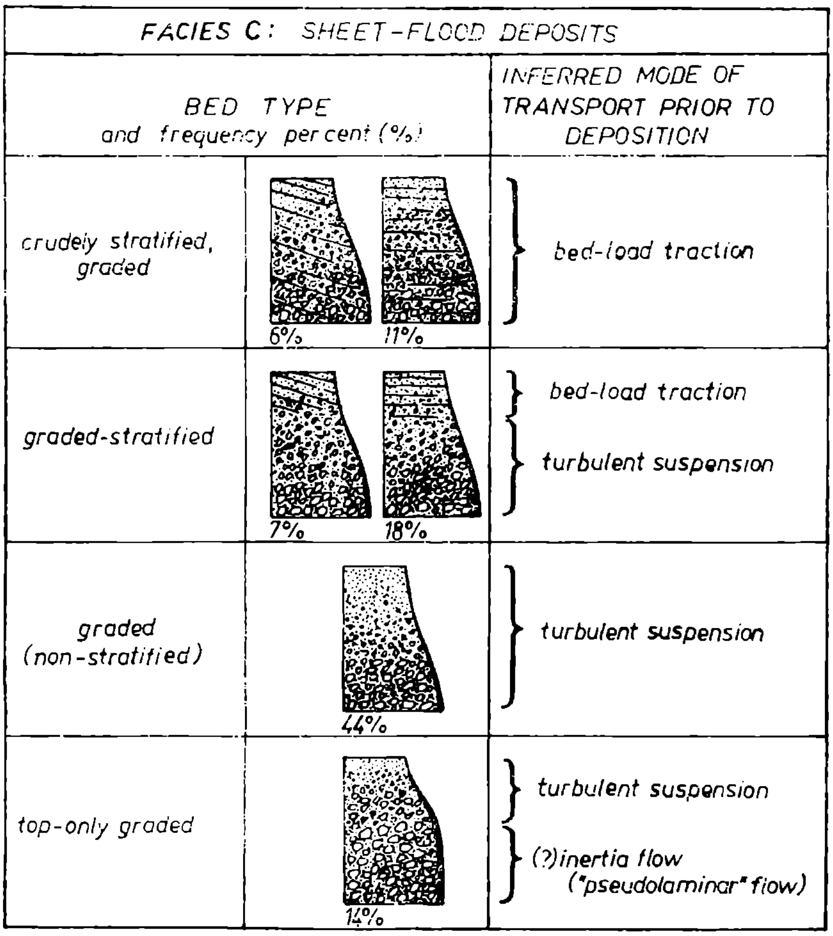



Fig. 15. Predominant bed types of facies C, and their interpreted mode of origin (for interpretational details see text). Note the relative frequencies (%) of individual bed types

Fig. 15. Dominujące typy ławic w utworach facji C oraz ich interpretowana geneza (szczegółowa interpretacja w tekście pracy). Podano względną częstość występowania (%) poszczególnych typów ławic

clear by grain-size variation between beds. Beds are laterally continuous across outcrops few tens of metres wide (Fig. 14) and any considerable channelling is rarely observed, with relatively few beds being lenticular in transverse sections and extending up to 5—7 m only. Traced laterally, only some entirely cross-stratified beds show any marked irregularities in their thickness.

These sheet-like beds are sharply based, occasionally sole marked (grooves), and relatively few exhibit marked erosional features at their lower boundaries. The individual beds always fine upwards, displaying a distribution (total-size) normal grading. The grading is developed irrespective of the bed grain size.

The beds are graded throughout, or display a graded upper portion associated with an essentially non-graded or only slightly graded portion below (top-only graded beds; Fig. 15). Gravelly lower portions of the beds, which are predominantly clastsupported (matrix content generally less than 40—50%) and contain abundant pebble- to cobble-size fraction, are usually overlain by a coarse sandy portion with granules and fine pebbles. These upper, sandy portions of the beds may be crudely horizontally stratified or vaguely cross-stratified (graded-stratified beds; Fig. 15), or are more often massive and irregularly bedded; in several instances the crude stratification, either horizontal or inclined, was observed troughout the bed thickness (crudely stratified beds; Fig. 15).

Within the graded-stratified beds, their lower (unstratified) division and upper (vaguely stratified) division are usually transitional, suggesting that the beds were formed by single depositional events. The faint cross-stratification, when present, is of low-angle (about 10—20°) planar type, while trough cross-stratification is rarely observed.

Certain beds are separated by layers of greenish-grey to brownish-grey mudstone, or silty sand, which vary in thickness from a few centimetres up to about 60 cm. When present near the top of the entire volcaniclastic succession considered, these clay-rich units may sometimes exhibit presence of dark, humified zones at their tops.

The pebble/cobble size fraction shows variable strength of orientation. Large number of clasts lie with their maximum projection planes parallel to bed boundaries, though numerous beds show examples of an upcurrent dip of the clast intermediate axes between 5° and 30° (from the bedding plane). In massive, unstratified beds there is locally developed a clast fabric consisting of imbricate clast longer axes parallel to flow, as indicated by adjacent cross-bedding measurements, and the intermediate axes transverse to flow; this corresponds to the *a*(p) *a*(i) fabric in Walker's (1975, p. 137) notation.

In most beds, however, the imbrication is only weakly and(or) locally developed. The large imbricate clasts are usually not in contact (isolate imbrication; Laming, 1966, p. 946). Although clast-supported textures prevail in facies C, these sediments contain a large proportion of fine-grained material (fine sand, silt and clay) intimately mixed with the imbricated gravel. Their mud content varies from 5% to about 30% (averagely 15%), and in certain beds exceptionally attains as much as 50%. In several instances a change from clast-supported to (subordinate) matrix-supported texture has been noted within one bed.

In the section of Stražarski Dol gorge (locality no. 1 in Fig. 2; see also Fig. 9), the beds of facies C are observed as stacked into a succession which is more than 100 m thick and in which the deposits of facies A and B are subordinate (see example in Fig. 16); the grain-size range and distributions of facies C, as well as its composition, all are comparable with those of the two associated facies in considerable outcrops. Within the Stražarski Dol section, the beds are organized into packets (2—5 m thick) within which the successive beds are thinning-and-fining upwards or thickening-and-coarsening upwards (sometimes with an alternation of the

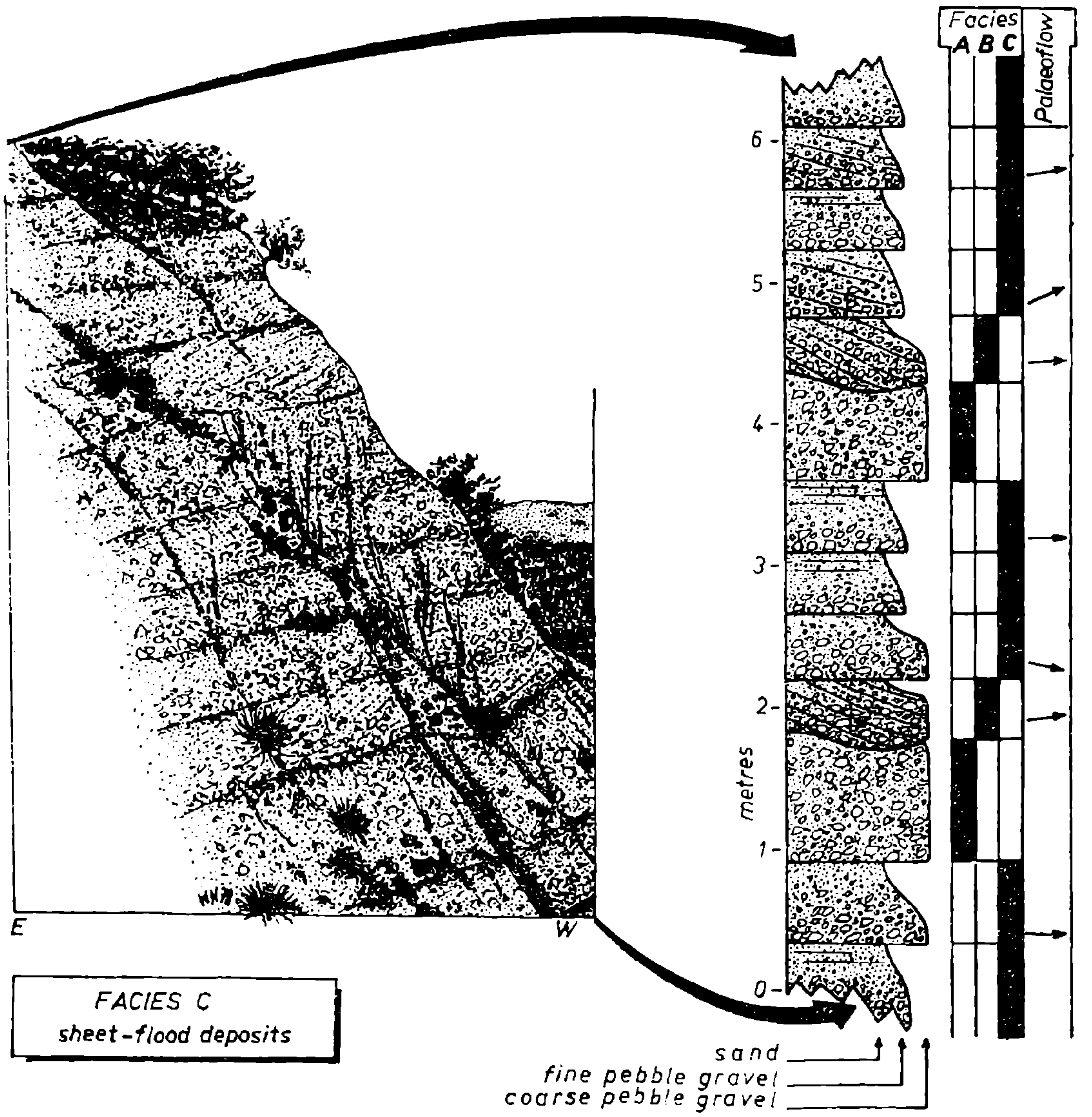



Fig. 16. Deposits of facies C, interbedded with subordinate units of facies A and B, cropping out at the wall of Strażarski Dol gorge: a general view of the outcrop (drawn from a photograph taken August, 1978) and the corresponding vertical section showing sedimentological details. (For outcrop location see locality no. 5 in Fig. 9)

Fig. 16. Utwory facji C oraz podrzędne ławice utworów facji A i B, odsłaniające się w ścianie wąwozu Strażarski Doł: ogólny widok odsłonięcia (rysunek z fotografii wykonanej w sierpniu 1978 r.) i profil pionowy ukazujący szczegóły sedymentologiczne (Lokalizacja zob. odsłonięcie nr 5 na fig. 9)

two trends), or are more often chaotically arranged (see examples in Figs. 14 and 16). Except for the isolated beds of facies B, the succession lacks any adjacent vertically stacked channel complexes. Palaeocurrent data continue to indicate a source area to the west (Figs. 14 and 16), and are also compatible with the axes of the isolated, minor channels (Fig. 16).

Noteworthy is the distribution of facies C in the study area. These deposits are present principally east of Mt. Vissoka Elha (Fig. 2), while are only occasionally found to the west this trend appears to be just opposite to that observed in the areal distribution of facies A.

In the section of Stražarski Dol gorge, in the basal portions of certain coarse-grained beds of facies C some drifted logs of charred wood (tree trunks up to 15 cm in diameter) have been found by the present authors. Most of these logs are charred only on the outside, while being internally petrified, and only few are almost completely converted to charcoal. They all were evidently transported, and no upright tree stumps have been found.

Interpretation. The deposits of facies C exhibit features of which some are suggestive of turbulent water-flows, while several others are indicative of mass-flow transport. Accordingly, the deposits of facies C appear intermediate between the debris-flow deposits of facies A and the stream-flow deposits of facies B.

Total-size normal grading, when present throughout the bed thickness, indicates that the grain-support and flow mechanism were probably dominantly turbulent suspension. Isolate imbrication is the result of low clast/matrix ratio (Collinson, 1978, p. 44), and is produced when clasts are too large for the prevailing current to move; the current undermines the finer sediment from the up-stream end of the scattered large clasts, resulting in an upstream imbrication at a lower angle than with the normal, or contact, imbrication (Leopold *et al.*, 1964; Laming, 1966). The *a*(p)*a*(i) type of fabric is known to be particularly common in resedimented conglomerates (Davies and Walker, 1974; Walker, 1975, 1977), where it is thought to result through clast deposition from turbulent dispersions.

On the other hand, the deposits of facies C apparently are not directly associated with channels and their deposition has not been also accompanied by any considerable channelling processes typical to fluvial transport. As seen in the field, and as to some extent predicted by the high bed-thickness/clast-size correlation (Fig. 13), there does not appear to be any considerable erosion at the bases of the superposed beds of facies C (comp. also interpretation of facies A earlier in text). The high positive correlation, in implying a close relationship between the flow capacity and competence, indicates that large clasts were transported only when the flow was powerful enough; they were available to flows with small loads, but these flows were not competent enough to carry the larger clasts. The decline in size is therefore thought to be due to a decline primarily in flow competence, as it is also to some extent indicated by the data presented in Figures 14 and 16 (see also discussion in next section).

Moreover, the deposits of facies C lack separation of fine and coarse material typical of much alluvium and show also one other unusual feature, namely the local presence of matrix-supported textures within beds dominated by clast-supported textures. While the mud content estimates do not provide suspended sediment concentrations during flow in the stream channels (as suggested above), it seems reasonable to argue that the deposits of facies C were laid down by flows which experienced high concentration of fines. Hence it is suggested that, in general, the matrix to the clasts was sufficiently watery to allow turbulence and clast imbrication, but that the concentration of mud was high enough to reduce the settling velocity of both sand and fine gravel (*cf.* Simons *et al.*, 1965; Wasson, 1977).

Since stream-flow and debris-flow deposits are usually viewed as endpoints on a continuum (*e.g.*, Bull, 1963; Wasson, 1977), it is thought here that the deposits of facies C represent an intermediate part of this latter spectrum of transport mechanisms („intermediate deposits" in Bull's classification). Hence the deposits of facies C are inferred to have been related to relatively low-concentration sediment flows, with the behaviour probably intermediate between the Newtonian viscous flows and Bingham plastic-viscous flows in Dott's (1963) terminology. The distinction of intermediate-type deposits seems to be physically adequate (see also next section), but unfortunately it does not provide sedimentary criteria for discrimination, and is therefore often omitted in the literature in order to simplify field classifications.

In light of the field data the reasoning above seems to serve as an adequate physical explanation, and a s h e e t - f l o o d origin is thought to provide a depositional mechanism capable of explaining the characteristics of facies C (for modern sheet-flood events and deposits see Rich, 1935; Davies, 1938; Bull, 1964, 1972, 1977; Rahn, 1967; Thorbecke, 1973; for interpreted ancient deposits see Cummins, 1958; Laming, 1966; Bluck, 1967; Wasson, 1977, 1979; Heward, 1978).

Sheet floods are short-lived events representing flood flows of relatively low (though possibly variable) viscosity, which expand at the downstream ends of discontinuous channels on the surface of alluvial fan, usually below the fan intersection point (Bull, 1972; Wasson, 1977, 1979). Sheet-flood advances as a shallow, continuous sheet of sediment-laden water, though the flow may be sometimes inhomogenous with anastomosing currents corresponding to bed irregularities. The flow of such dilute slurry is violent and generally develops upper flow-regime conditions (Collinson, 1978, p. 18), but the deposition takes place quickly, due to flow velocity rapid decrease in effect of the shallowing caused by lateral spreading and the draining of water into fan-surface sediments.

The flows considered were preferentially of unchannelled to only slightly channelled type, and their densities were presumably somewhat lower than 1.5 g/cm$^3$ (comp. Hampton, 1972, p. 791; Rocheleau and Lajoie, 1974, p. 834). Bed characteristics indicate that the sediments of facies C have been transported both in turbulent suspension and by bedload traction, with the former being sometimes density-modified to a sort of inertia flow (or "pseudolaminar" flow). This mode of sediment transport, as inferred from facies C beds, is reviewed in Figure 15 and is now to be considered in some more details.

The graded beds displaying faint stratification (horizontal or inclined) throughout their thickness (Fig. 15) indicate bedload transport with sufficient fluidity for the settling of pebbles and formation of traction fall--out structures. It is likely, therefore, that these *crudely stratified beds* have been traction emplaced (*cf.* Rocheleau and Lajoie, 1974); they were probably deposited by sediment-laden sheet water flows, which transported the sediment almost entirely as the traction load and the load held in the tractive carpet (*cf.* Bluck, 1967).

Consequently, the *graded-stratified beds* (Fig. 15) could be interpreted in terms of sheet-floods in which deposition from turbulent suspension prevailed in the lower portion of a flow, with the tractive transport restricted to its upper portion only (*cf.* Davies and Walker, 1974, p. 1210). The *graded unstratified beds* (Fig. 15), in turn, were probably deposited entirely through fallout from turbulent suspension (*cf.* Middleton, 1970; Middleton and Hampton, 1976; Cas, 1979, p. 79).

On the other hand, it seems not unlikely that at least some of the graded unstratified beds can be ascribed to a simple debris-flow mechanism. This is because many of these beds contain abundant mud-rich matrix which ought to be competent enough to carry the clasts, particularly since mud concentrations of as low as 1.1% have been reported from certain debris flows by Curry (1966). It has been also recently suggested (Bull, 1977, p. 237; Carter and Norris, 1977, p. 306—307; see also Hampton, 1975) that a low-viscosity debris-flow will probably have graded bedding and a horizontal or imbricated orientation of flat clasts. It seems likely that a decrease in matrix competence due to debris-flow dilution, for example, may sometimes produce normal grading as the large clasts settle; such beds will show an arrested settling pattern, where both evidence of turbulence and resedimentation are preserved (*cf.* Long, 1977).

The *top-only graded beds* (Fig. 15) might have been deposited from density-modified turbulent suspensions, in which the grain-support has probably been dominantly by dispersive pressure near the base and by fluid turbulence at higher levels in the flow (*cf.* Davies and Walker, 1974), at least immediately prior to and during the deposition. At the lower levels in such flows the turbulent suspension is thought to have been probably subordinate to inertial dispersive pressure and the buoyancy effect of the interstitial fluid (fluid strength) as grain support and flow mechanism (Hampton, 1975); matrix strength was probably the most important factor here, at least in the mud-rich portions of a flow.

In summary, the evidence from facies C points into sheet-flooding as the main process of sedimentation, and to aggrading alluvial fan as the resulting accumulation. These sheet sediments possibly formed as intersection point deposits at the downstream ends of discontinuous channels, but may have formed by run-off from precipitation directly on the fan surface (*cf.* Wasson, 1977, p. 793). Variation in grain size in successive beds did probably result from both variation in rain-fall intensity and the location of the sheet flood initiated by them. It is probable here that the common accessibility of easy eroded material of wide size spectrum (loose regolith) did cause the preferential production of sheet floods, rather than ordinary stream flows.

The locally present thin units dominated by mud and silty sand are thought to be the suspended load deposits of discontinuous channels which spread their load in thin sheets across the fan toes; this mode of deposition was observed in the distal parts of modern intersection-point deposits (Wasson, 1974, 1977). Thus, minor or wanning floods might have been responsible for the deposition of these interbeds. Their humufied cappings, as present in certain beds, may thus be indicative of a temporal relaxation in the depositional processes on fan surface.

Clusters of juxtaposed beds of facies C are thought to be the result of deposition by successive flows, either as discrete events or as pulses in an ongoing depositional event. The thickening-and-coarsening upward and thinning-and-fining upward sequences of these deposits may be possibly analogous to outer fan turbitite sequences of similar style (Mutti and Ricci-Lucchi, 1972; Walker and Mutti, 1973; Mutti, 1974, 1977; Ricci-Lucchi, 1975; van Vliet, 1978). Similar trends have been reported by Heward (1978) from ancient successions of sheet-flood deposits studied by him. As postulated by Heward *(op. cit.)*, the organization of sheet-flood deposits into vertical sequences suggests that the sheet floods have possibly been distal representatives of mid-fan lobes undergoing progradation (coarsening-and-thickening upwards) or abandonment (fining-and-thinning upwards). Variable rain-fall intensity and storm events reworking earlier fan-head deposits, together with any other pulses in sediment supply (as tectonic triggering, for example) all might have been important agencies modifying the vertical trends throughout the bed succession, and so the trend is rather expected to be variable. It seems likely here that many of the small-scale bed sequences considered (see examples Figs. 14 and 16) reflect, by their vetrical trends, the gradual or abrupt initiation and termination of sedimentation within a given depositional segment of the alluvial fan (*cf.* Heward, 1978, p. 473).

## A MODEL FOR DEPOSITION OF SHEET-FLOOD (FACIES C) BEDS

The possible genetic relationships between the main bed types of facies C are graphically shown in Figure 17. There are two obvious ways of producing the low-viscosity (limited competence) sheet-flood slurries capable of depositing the type of beds considered: debris-flow dilution, or progressive increase in the suspended load of stream-flow (*cf.* Wasson, 1979, p. 30). From the field evidence both of these latter explanations seem likely here, since both debris-flow and stream-flow deposits are associated with the deposits of facies C.

A landslide which supplied material into a water flow fed by torrential rainfall, for example, could well have produced a water-dilute slurry, or, if there was sufficient debris or not enough water, an ordinary debris-flow; in such settings the occasional stream-flows might have been due to temporal scarcity of detritus in the catchment area. The catchment settings (particularly the ratio of water volume to sediment supply) east of Mt. Vissoka Elha might have been so that the newly generated landslides or avalanches became preferentially dilute due to incorporation of stream and rain waters, and only occasionally some pure debris--flows reached more distal depositional sites.

On the other hand, the water flows (initiated as stream flows or stream floods) might have become strongly laden with sediment and richer in mud as, after a rainfall, some sediment-laden tributaries reached the main stream, or as the result of excessive sediment supply from landslide or debris-flow directly into a flowing stream (*cf.* Beverage and Culbertson, 1964; Wasson, 1979). Unfortunately, little can be said about the precise characteristics of such flows, since the behaviour of bedload in streams of high suspendedload concentration is virtually unknown (Howard, 1963; Leopold *et al.*, 1964; Graf, 1971).

Anyway, on the surface of an alluvial fan the streams spread out when they reach the end of discontinuous channels; this usually happens at what has been termed the „intersection point" (Hooke, 1967; Wasson, 1974, 1977). Since the stream discharge *(Q)* is equal to the product of the mean width *(w)*, depth *(d)* and velocity *(v)* of flow (Bull, 1977, p. 227):

$$Q = w \cdot d \cdot v,$$

the increase in flow width is accompanied by concurrent decreases in depth and velocity, which are the primary causes of sediment deposition. The discharge may also decrease when the flow occurs over permeable surficial deposits, thereby tending to increase sediment concentration and resulting in deposition. Flow concentration may itself increase due to the aforesaid agencies, and also due to sediment intake from the bottom.

Field data indicate that the catchment settings east of Mt. Vissoka Elha, as opposed to the area extending west of this locality, apparently were conducive to prolonged production of sheet-floods, while produced only subordinate stream-flows and even fewer debris-flows. The prevalence of sheet-floods and the temporal occurrence of the remaining two depositional modes are both thought to have been dependent primarily upon the ratio of sediment to water content of any one flow (*cf.* Wasson, 1977).

The deposits of facies C may probably considered as the result of „hyperconcentrated flows" (Beverage and Culbertson, 1964), which are water/sediment mixtures having between 40 and 80% sediment by weight. These flows are intermediate between normal stream-flow, in which there is no more than 40% sediment by weight, and debris-flow (or mudflow), which has more than 80% sediment by weight according to Beverage and Culbertson (see also Richardson, 1968). However, although the abovesaid flow-type continuum can be theoretically viewed as a graph of volume fraction of sediment against volume fraction of water (see Beverage and Culbertson, *op. cit.*), the boundaries between either the debris-flow or water-flow deposits and the deposits of the intermediate type cannot be specified according to this graph with any confidence (comp. also interpretation of facies C earlier in text). Nevertheless, it has been suggested (Hooke, 1967; Bull, 1972) that as the volume fraction of water decreases, a point is reached where sediment can no longer be deposited selectively, and the entrainment of particles becomes irreversible under conditions of mass flow. Inversely, an increase of, say, 5% water to a slurry of low water content will produce a larger decrease in flow competence than will an addition of 5% water to a more dilute slurry (Hampton, 1975, p. 839). These are, in fact, the basic assumptions of the herein postulated model for the origin of facies C beds (Fig. 17).

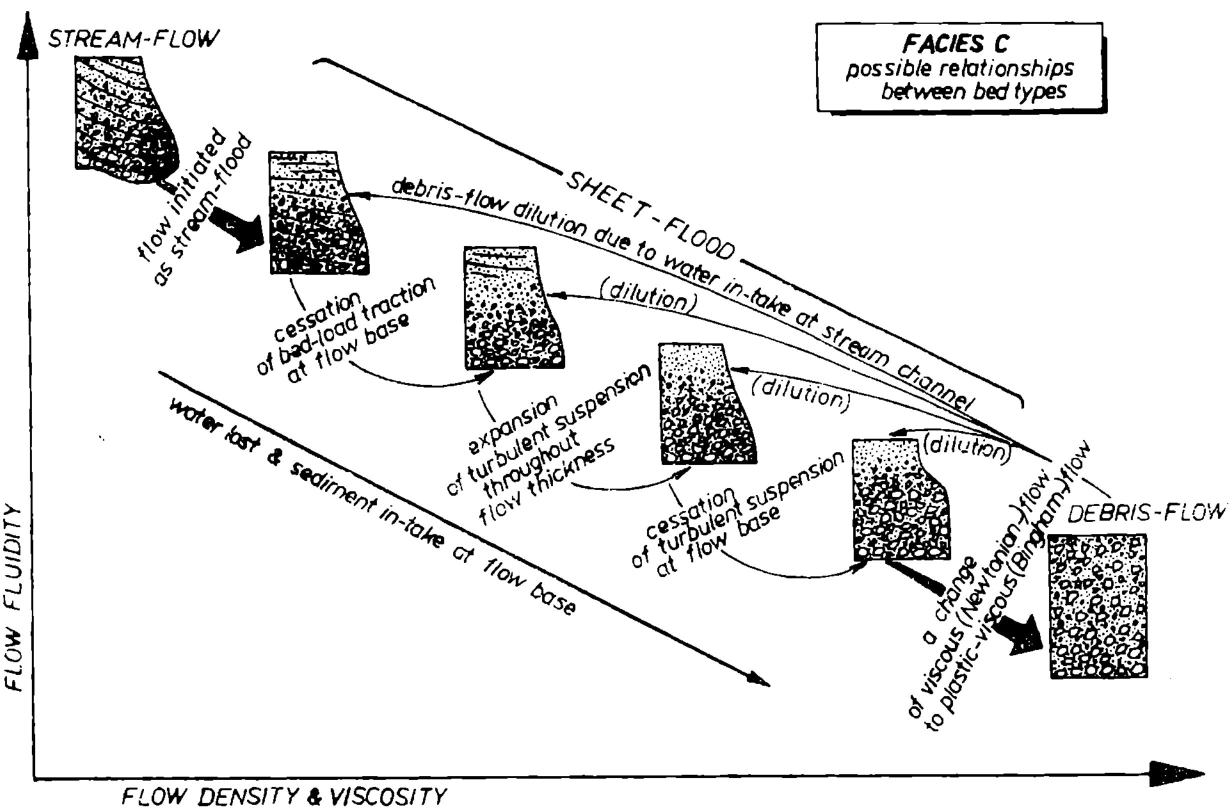



Fig. 17. Hypothetical model of genetic relationships between the beds of facies C (see bed types in Fig. 15, and interpretational details in text)

Fig. 17. Hipotetyczny model genetycznych relacji między ławicami facji C (zob. typy ławic na fig. 15 oraz szczegóły interpretacyjne w tekście pracy)

The basic assumption of this model is that a change in water content (or in water to sediment ratio) results in a respective change in flow density and viscosity, and thus also a change in flow competence; this is because the competence of a fluid matrix is known to be proportional to water content (Hampton, 1975). According to this model, a close relationship between the crudely stratified beds, graded-stratified beds and graded unstratified beds of facies C is postulated here, and this relationship is thought to be transitional in character. This transition (Fig. 17) may be interpreted in terms of the progressive cessation of tractive transport due to density increase at flow base, with turbulence becoming progressively more important at lower levels and rising successively upwards until the entire flow becomes a turbulent suspension. The origin of top-only graded beds (Fig. 17) could thus be due to an almost instantaneous cessation of turbulence concomittant with further increase of matrix competence above the flow base.

In an analysis of static liquefied beds, Wallis (1969, p. 191) recognized simple upward movement of resedimentation surface as a cause of increased matrix competence; this forms near the base of the dispersion and rises up until the entire unit is grain supported. The origin of top-only graded beds, in implying a progressive upward expansion of „intertia-flow" portion (Fig. 17), might have been possibly related to similar phenomenon. However, there might have been an interval when clasts settled through the matrix as turbulence wanned and before the matrix "set up". In such case, an abrupt increase in matrix competence would be likely due to resedimentation of matrix particles within the flow; moreover, this process of "internal redistribution" is probably sufficient to prevent further downslope movement (Lowe, 1976).

When taken together, the details of the model presented in Figure 17 are however largely hypothetical, and the reader is cautioned that several topics discussed under the beds genesis, and relationships are areas of active research and are changing rapidly. As a result, some portion of the above interpretations may be dated and should be applied prudently.

## SUMMARY EVIDENCE OF VOLCANIC ACTIVITY AND EPICLASTIC MATERIAL REWORKING

The volcaniclastics considered are of plagiorhyolitic derivation and all have been evidently derived from the volcanic body of Mt. Vissoka Elha in the centre of the study area (Fig. 2). The palaeogeography of the volcaniclastic seccession is to be discussed in details farther in text.

Mt. Vissoka Elha is inferred here to be an Early-Oligocene volcano. This was active through a relatively short period of time, but is thought to have supplied enough pyroclastic material and lava-sheets to built up an ash-and-cinder cone of considerable volume; the lava sheets were probably of low extent, and were entirely restricted to the summit cone. The volcaniclastics were derived through subsequent erosion of the extinct, strongly weathered volcano. In effect, the volcano became entirely eroded, and only a deeper portion of the volcanic plug is actually cropping out at Mt. Vissoka Elha, with no remnants of the primary pyroclastic cone being preserved.

There is, therefore, no direct evidence of the volcanic activity and primary pyroclastic deposition. These processes, however, are inferred here from some indirect data, which are thought to be a compelling evidence of the volcanic activity. This evidence includes:

(1) the presence of burnt wood fragments in the volcaniclastics, comprising variously charred tree stumps (transported logs) and also numerous interstratal veneers of fragmental charcoal;

(2) the structural and textural characteristics of many volcanic clasts, implying nearsurfacial to subaerial conditions of lava consolidation;

(3) the presence of bentonite layers at the base of the Tertiary succession in the study area (Dr B. Alexiev, personal commun., 1979), which are thought to represent the weathered products of some exceptional ashfalls;

(4) the large volume of the volcaniclastics under consideration, which is unlikely to have been derived solely from the eroded plug; the same regards also the large volume of clay in the volcaniclastics, and the highly variable degree of preservation and variable appearance of the volcanic debris.

Except for the already noted bentonite layers, no primary pyroclastic deposits are found as preserved in the study area. The entire volcaniclastic pile, therefore, is inferred to have been epiclastically derived from the extinct, strongly weathered volcano of Mt. Vissoka Elha. Primary evidence implying the epiclastic derivation includes:

(1) the composition of these volcaniclastics (as discussed in the beginning of this paper); in the absence of a known volcanic process which generates through pyroclastic activity an aggregate of lithic components and crystals and no vitriclastic material (but abundant clay), it must be concluded that the primary volcanic material underwent strong weathering prior to its final deposition;

(2) the inferred mode of the volcaniclastics emplacement, pointing into resedimentation processes; the characteristic association of sedimentary facies implies "cold state" resedimentation of strongly weathered, remobilized valcanic regolith.

Thus, there must have been a considerable time-delay between the volcanic activity and material redeposition. From the inferred chronology of events (Fig. 18) it seems likely that delay was possibly up to 20 million

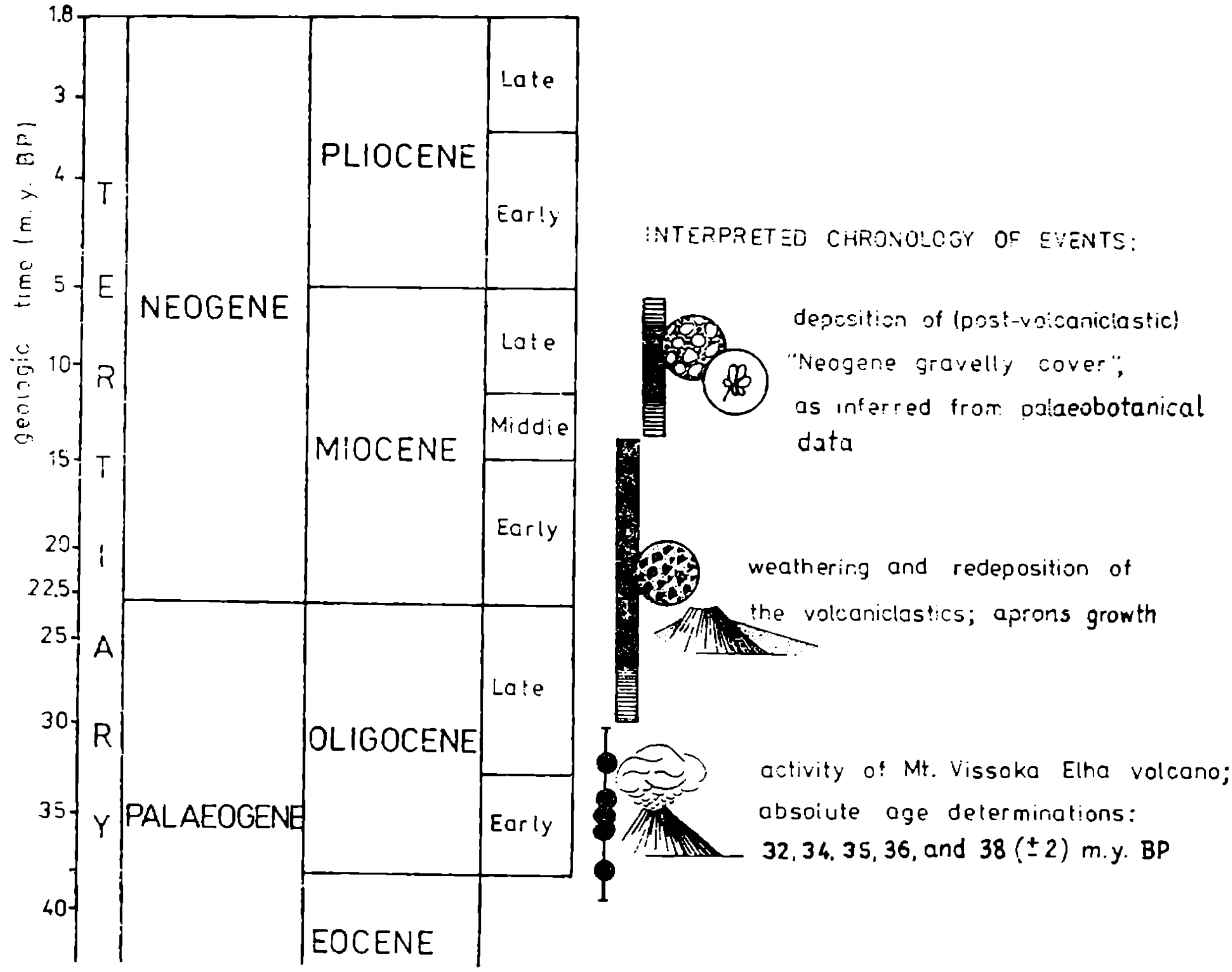



Fig. 18. Interpreted time-scale chronology of the events considered (time-scale according to "Geological Time Table", 3rd edit., compiled by F. W. B. Van Eysinga, 1975, Elsevier — Amsterdam)

Fig. 18. Interpretowana chronologia omawianych w pracy wydarzeń na obszarze badań (skala czasu według: „Geological Time Table", *op. cit.*)

years long. Taking together, the evidence clearly suggests that the redeposition of the volcaniclastics was virtually a typical process of epiclastic reworking.

## SOURCEROCK WEATHERING AND HYDROTHERMAL ALTERATION

In the present case there are two most obvious processes which have contributed to the origin of the volcaniclastics: surfacial weathering, and hydrothermal alteration of Mt. Vissoka Elha volcano. Although there is evidence that both of these agencies did in fact contribute to the origin of the deposits considered, we believe that the weathering process has been the more important of the two and provided a more adequate source for the large volume of clay present in the volcaniclastics.

Weathering processes. The volcaniclastics are thought to have been derived from the blankets of weathered volcanic ash and debris which did originally cover the flanks of Mt. Vissoka Elha volcano; weathering of the volcanic plug and local lava-sheets might have also been an important contributing factor. The regolith cover was then successively stripped away from the volcano by slopewash and other substantial erosion processes.

The abundance of weathered material at the volcano is readily explained by the predominance of chemical weathering. On the basis of his palaeobotanical data, Palamarev (1964, 1971) has recently suggested that humid, subtropical climate prevailed in the region during the Tertiary. Also the extensive brown-coal beds in the nearby Čukurovo Basin are proof of a considerable humidity for long stretches of the Tertiary. Palamarev (1964) suggests a mean annual precipitation of about 800—1000 mm, and a mean annual temperature of about 16—18°C; he postulates mean temperatures of about 20°C and 6°C for the warmest and coldest months, respectively.

Such climatic settings must have inevitably led to strong chemical weathering and high denudation rate. On the account of these factors, extensive clay profiles of weathering must have been developed in an almost continuous manner throughout the area of pyroclastic deposition; and with a continuous chemical denudation the entire pyroclastic cone did probably become virtually weathered throughout.

Recent examples of thick, widespread blankets of weathered rhyolitic ash, virtually altered to a greasy yellowish or greenish clay containing abundant montmorillonite/illite are described by Fiske *et al.* (1963) among the numerous others. From a subtropical igneous terrain Drever (1971) reports modern streams draining blanket of unconsolidated rhyolite ash which have up to 820 ppm dissolved solids. He also reports a chemical denudation rate of 91 tons/km$^2$/year for such ash blankets, showing the influence of highly reactive starting material. These processes are likely to have been of primary importance to the problems discussed herein.

Hydrothermal alteration. On the other hand, in the present case there are enough proof to state that Mt. Vissoka Elha volcano did suffer hydrothermal alteration. Cristobalite, sulphide minerals, barite, hematite, and calcite are all known to be normal and expected products of hydrothermal alteration within a volcano the same may also regard the origin of zeolite heulandite, hydromicas and K-feldspar (Whitebread, 1976). As already described in text, all these minerals are common in the volcanic plug of Mt. Vissoka Elha.

However, the highly limited exposure at the last-named locality gives no direct proof that the hydrothermal processes were either unusually strong, or did ever produce the large volumes of clay present in the volcaniclastics. Despite the generally low rank of rock alteration, it seems probable that in some parts of the volcano the hydrothermal processes may have produced much stronger alteration of the wallrock. Since the volcanic plug is evidently fault-bounded (Fig. 1), some stronger alteration is likely to have occurred along the faults and fractures which were the main channelways for solutions (*cf.* Harvey and Vitaliano, 1964; Whitebread, 1976, p. 14). The fault-controlled hydrothermal activity, by being focused primarily on faultlines, could well have produced some relatively narrow zones of densely altered wallrock. Such zones may not be actually detectable at Mt. Vissoka Elha because of the scarce exposure and dense vegetation cover. Besides, if the process was so, the marginal hydrothermal activity might well have caused alteration of the lava-sheets and adjacent pyroclastic deposits of the summit cone.

The inferred hydrothermal activity must not have been necessarily prolonged, high temperature, or unusually strong. For example, Crandell (1971) has shown that considerable portions of the Holocene lava-sheets and wallrock of Mt. Rainier summit cone, Washington, became altered through ordinary solphatarization to a loose, sandy, clay-rich material during a time period as short as the last 1000—2000 years.

The abovesaid hypothesis has also one other corollary implication. If strong hydrothermal alterations did really take place in some portions of the volcano, the process would have direct bearing on the rate of erosion. Because of the inherent structural weakness of hydrothermally altered rock, the rock becomes especially susceptible to large-scale failure and sliding. Moreover, if slides of altered rock were to occur in an active hydrothermal area, the slide material most likely would already be wet because of condensed steam and thus would readily become a debris-flow as the slide moved downslope.

## PALAEOGEOGRAPHY AND DEPOSITIONAL SETTING

The isolated, scattered occurrences of the volcaniclastics, as recognized in the study area (Fig. 2), are interpreted here as fragments of two extensive alluvial aprons which originally extended principally east and northwest of Mt. Vissoka Elha, respectively (Fig. 21, top). The provenance of volcaniclastic material and the palaeoflow data all are proof of this interpretation; palaeocurrents indicate that the drainage was mostly from the vicinity of Mt. Vissoka Elha, and was principally directed easterly and northwesterly (Fig. 21, top).

Thus, the two opposedly directed alluvial fans grew in concert, and were built by debris derived from a localized drainage area, corresponding essentially to the flanks of Mt. Vissoka Elha volcano. Climate was warm and humid, and the volcanic cone underwent strong weathering and denudation. The material was derived from remobilized, unconsolidated regolith of the volcanic cone and through erosion of the volcanic plug itself; this produced a clay-rich sediment consisting of both strongly altered and nearly fresh debris, intimately mixed with each other. Subordinate nonvolcanic rock fragments are of strictly local derivation too, and essentially correspond to the close neighbourhood of the volcano.

The volcaniclastic alluvial aprons are underlain by, and also locally interbedded with the polymict sediments derived from the basement rocks present in the vicinity of the volcano. These fluvial deposits, containing subordinate volcanic clasts too, were laid down by a braided-stream system which was essentially contemporaneous to and slightly younger than the volcanic activity, but presumably also partly pre-existed this latter. This extensive fluvial system reworked most of the primary pyroclastic cover off the volcanic cone, resulting in high dilution of the volcanic debris by non-volcanic clastic material. Any remnants of the primary pyroclastic cover, which was presumably relatively thin off the volcano, underwent strong weathering and thus lost their characteristic vitriclastic features too. The bentonite beds, which have been locally found as present at the base of the Tertiary succession (Dr B. Alexiev, personal commun., 1979), are thought to be the weathering products of this sort.

Noteworthy is the similarity in palaeocurrents noted between the basal polymict deposits and the overlying volcaniclastics. This implies that the deposition of this entire clastic succession did essentially take place within same palaeoslope settings. Moreover, the rugged prevolcanic surface with high-relief topographic features (which pre-existed the Tertiary molasse sedimentation was probably only partly subsumed by the initial (polymict) fluvial deposits, and was subsequently overridden and filled by the prograding volcaniclastic aprons.

The approns were extensive, outward thinning, fan-shaped wedges. Data on their original thicknesses are scant (Fig. 19), but from palaeogeographical reconstruction the thicknesses are inferred to have been greatest in fan-head areas, just near fan apexes; based on the present height of Mt. Vissoka Elha, these maximum original thicknesses are deduced to have been probably about 200 m and more than 350 m for the eastern and for the western apron, respectively. The easterly trending apron is thought, therefore, to have been relatively thinner of the two. These latter apron was probably about 3.5 km long and 3 km wide, while that trending northwesterly was more than 4.5 km long and 3 km wide (Fig. 21, top). So far there is no evidence to indicate whether these two alluvial prisms ever extended further out distally.

Primary dips of the volcaniclastics strata are commonly more than 20° (up to 30°) near Mt. Vissoka Elha and, in general, the strata dip away

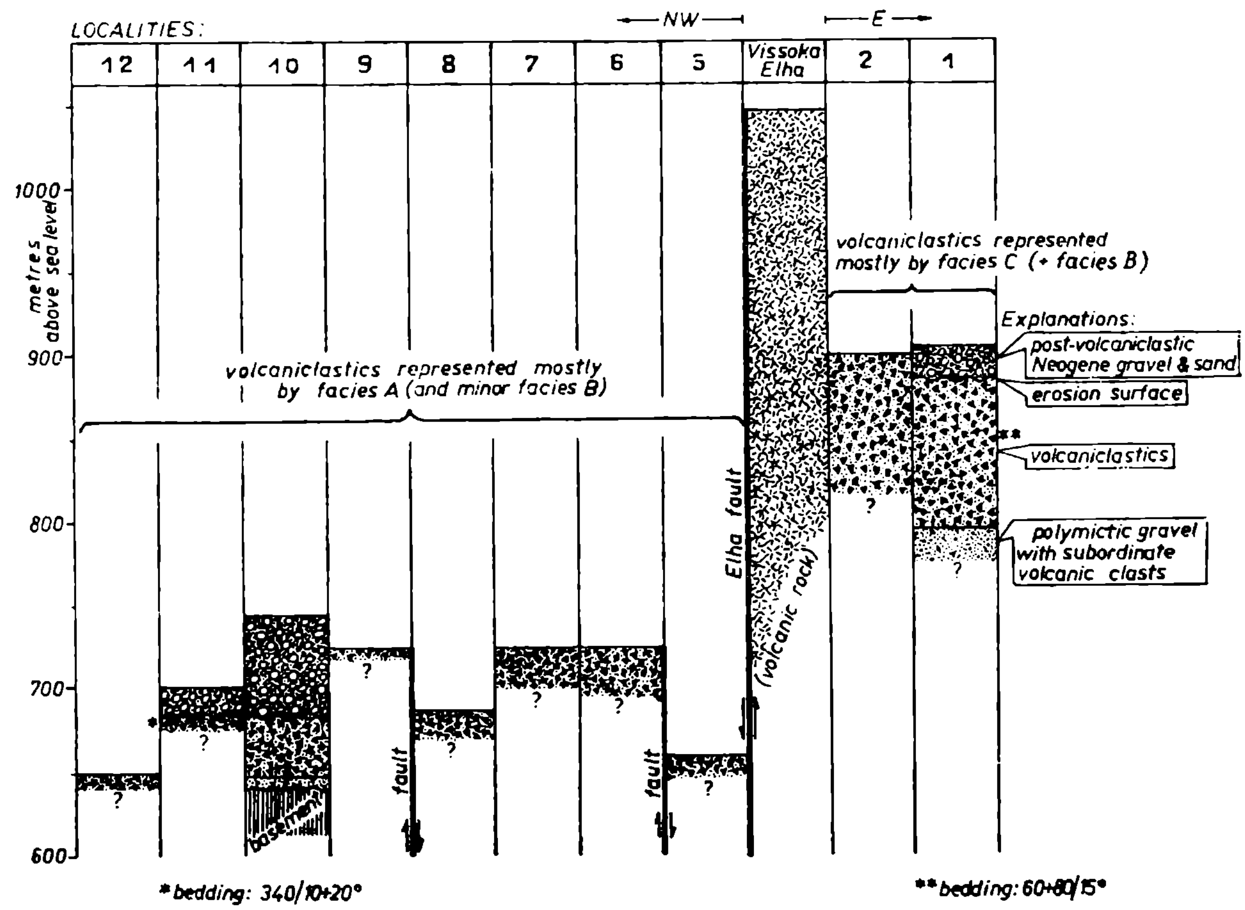



Fig. 19. Summary data on the present altitudes of individual outcrops, and their vertical relationships to Mt. Vissoka Elha (no horizontal scale; vertical scale only). Columns correspond to individual localities; the localities are numbered as in Figure 2

Fig. 19. Podsumowanie danych dotyczących wysokości poszczególnych odsłonięć nad poziomem morza oraz ich relacji wysokościowej względem góry Vissoka Ełha (bez skali poziomej; zachowana tylko skala pionowa). Kolumny diagramu odpowiadają poszczególnym odsłonięciom (numeracja jak na fig. 2)

from the last-named locality. The base of the volcaniclastic pile is at different altitudes in the region (Fig. 19), and this is primarily due to faulting. Deformation of the series is slight, and structural tilt is generally less than 10° except locally near faults. From this evidence, relatively high gradients are inferred for the depositional surfaces of the aprons (probably up to 15—20° in proximal reaches). Considering the direction of tilting (see Fig. 1), however, somewhat smaller values (probably not greater than 5—10°) must be deduced for the easterly trending apron.

Although having several features in common, the two alluvial aprons are of contrasting types. The dominating element in the northwesterly trending apron are debris-flow deposits (facies A), interbedded with subordinate stream-flow deposits (facies B) and occasional sheet-flood units (facies C). In contrast, the easterly trending apron is characterized by the predominance of sheet-flood deposits, accompanied by subordinate stream-flow deposits and occasional debris-flow units.

The alluvial volcaniclastic aprons are erosionally dissected and disconformably overlain by the extensive, thick blanket of Middle/Upper Miocene polymict deposits (the "Neogene gravels"). These are clast-supported, imbricated gravels which display typical aspect of fluvial sedimentation. The unusually coarse grain-size and highly erosional base of these gravels point into deposition by gravelly braided-streams of high eroding power. Their deposition probably took place mostly through flood-surges, with the cobble/boulder gravels being laid down principally within some broad, erosional troughs (see example in Fig. 7). As sedimentation progressed, however, these high-relief topographic features became gradually subsumed and filled, and some finer-grained clastics became dispersed by streams onto adjacent areas; these depositional relationships are well illustrated by the section of Stražarski Dol gorge (Figs. 9 and 20).

Noteworthy are the palaeoflow data derived from the abovesaid gravelly cover, which is erosionally stacked upon the volcaniclastics (Fig. 21, top). For these blanket gravels the drainage was towards the northeast, implying that these polymict fluvial deposits were demonstrably transported across the area occupied by the volcaniclastic aprons, at a high angle to their axes. This suggests an important change in the regional palaeoslope direction, and a corresponding change in the location

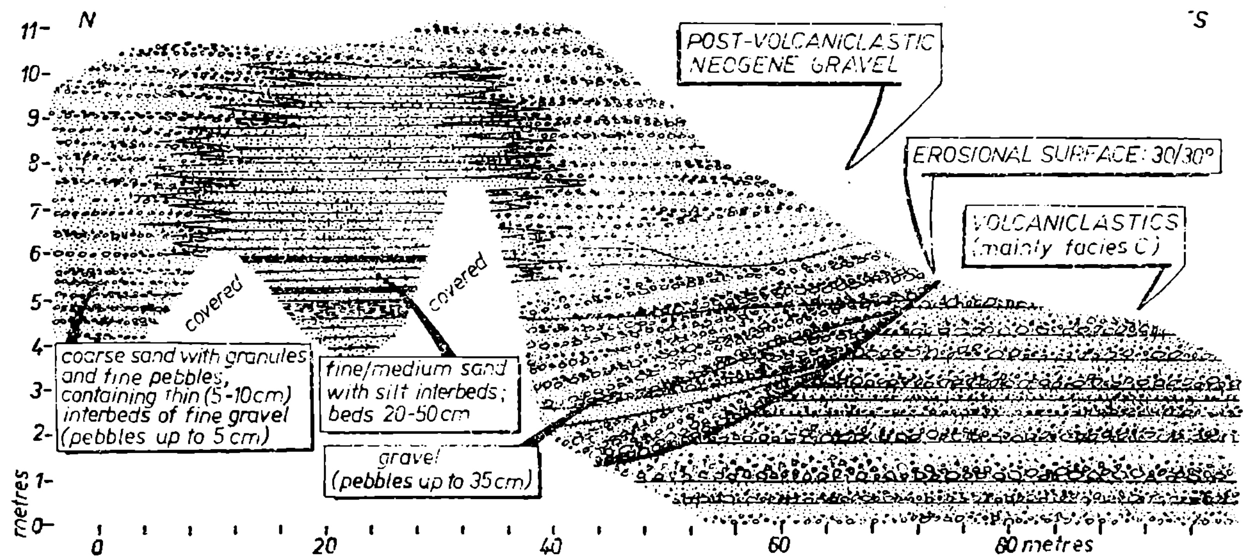



Fig. 20. Outcrop section showing the erosional contact between the volcaniclastic pile and the overlying "Neogene gravels", and the lateral variability in lithology of these latter deposits (note the vertical exaggeration). For outcrop location see localities no. 1 and 2 in Figure 9

Fig. 20. Szkic odsłonięcia ukazujący erozyjny kontakt między serią wulkanoklastyków a wyżejległymi „żwirami neogeńskimi", a także lateralną zmienność litologii tych ostatnich utworów (skala pionowa przewyższona). Lokalizacja zob. odsłonięcia nr 1 i 2 na fig. 9

of the drainage area. Thus, in the Middle/Late Miocene there took place an abrupt cessation of the pre-existing, local palaeoslope gradients which previously controlled the deposition the volcaniclastic aprons themselves.

Since the northeasterly directed, flood-surge fluvial transport was markedly parallel to the direction of the Elha fault (comp. Fig. 21, top), it seems likely that the sediment dispersal at that stage might have been partly controlled by the axis of Železnitza graben. The overall palaeogeography and provenance of the gravelly cover imply Middle-Miocene structural control of the fluvial system, probably due to uplift in the sourceland to the south and southwest. Clast composition points into material derivation from the nearby ranges of Plana nad Vitoša Mountains.

## TECTONIC-SEDIMENTARY MODEL

The volcano of Mt. Vissoka Elha is a unique example of the Tertiary volcanism in Western Srednogorie. However, this volcano appears to have originated during an Early Oligocene episode of a more widespread emplacement of small, shallow, subvolcanic intrusions. Compositionally similar intrusive bodies are locally present in the Western Srednogorie, and are common in the nearby Kraištides; for example, Harkovska *(in press)* reports about thirty subvolcanic bodies of rhyolitoid composition as present between the vicinities of Trân and Divlia, Kraištides, and she refers to their age as Palaeogene. It is to be stressed, however, that no products of subaerial volcanic activity have been reported as associated with any of these igneous bodies. Thus, the Early Tertiary volcanic events in the region were most probably an accidental phenomenon, itself closely related only to some local, most favourable structural settings of the basement (a "one-chance" volcanism).

Presentday geological situation (Fig. 1) implies that the igneous body of Mt. Vissoka Elha has been emplaced along (sub-)vertical structural line of the contact between three separate basament blocks, namely the gneissic complex of Plana Mts., the andesites of Lozen Mts., and the diabase-phyllite complex of Vitoša Mts. (see also "Regional geological setting" earlier in text). The volcanic plug is markedly bounded by three important faults: the Elha fault, the Iskâr fault, and the Pasarel fault (Fig. 2). Presentday situation gives direct structural proof that both these and some adjacent faults have been active at least at the end of the Tertiary, and themselves post-date the molasse deposition. We believe, however, that these faults mark also the locus of syndepositional Tertiary faulting. This latter appears locus of syndepositional Tertiary faulting. This latter appears to have controlled the deposition of the Tertiary molasse succession in the study area, and in particular the development of the volcaniclastic aprons in question.

In order to contract a model for the volcaniclastics redeposition, it is desirable to consider the contrast between the two identified aprons. As discussed in the previous section, the two alluvial aprons grew in concert and the immediate cause of their growth was the erosion of the extinct and strongly weathered Elha volcano. The response to erosion

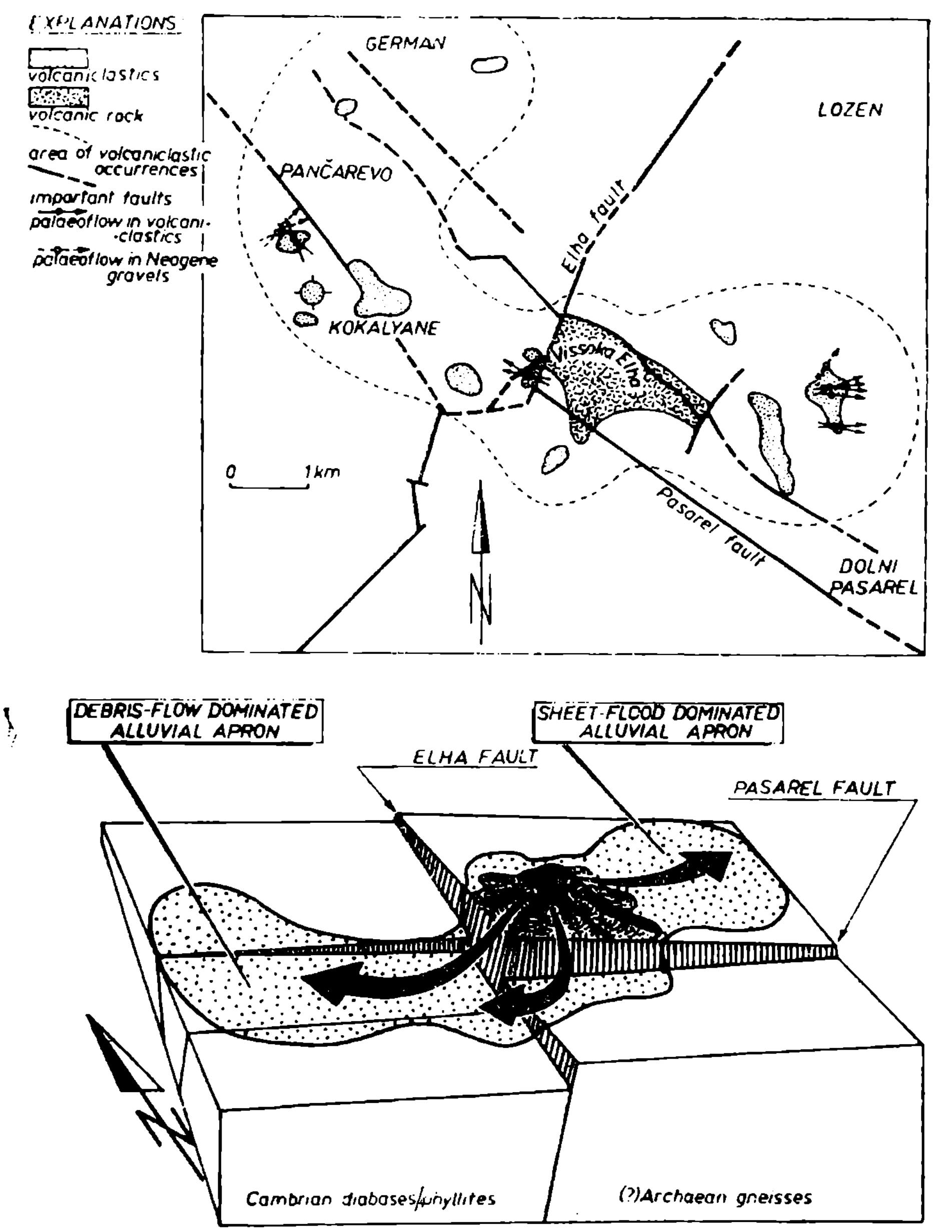



Fig. 21. Interpreted palaeogeography of the volcaniclastics *(top)* and a tectonic-sedimentary model of their depositional settings *(bottom)*; for details see discussion in text

Fig. 21. Interpretacja paleogeografii wulkanoklastyków *(u góry)* i tektoniczno-sedymentacyjny model warunków ich depozycji *(u dołu)*; szczegóły omówione w tekście pracy

in the west was the development of fairly steep, thicker alluvial apron dominated by debris-flow deposition, while the response in the east was thinner, lower-angle apron constructed predominantly by sheet-flood/ /stream-flood and minor stream-flow processes.

Thus, the two alluvial aprons, as developed on the opposite sides of eroded volcano, are in striking contrast, suggesting that there must have been some underlying control on their formation. In the present case it seems clear that neither a source-rock control, nor a simple climatic control can account for these differences. Taken together, however, the differences appear possible to explain in terms of a tectonic control over sedimentation. A suitable tectonic-sedimentary model is graphically shown in Figure 21 (bottom), and is discussed below.

We believe that the main vertical movements responsible for the sediment supply and apron growth were persistently located along the Elha fault, and partly also along the Pasarel fault. Normal faulting, with considerably more rapid (or more continuous) movement and subsidence along the Elha fault, did probably result in an eastward tilting of the entire basement block extending between Dolni Pasarel and Lozen, and a corresponding subsidence within the adjacent Železnitza graben (Fig. 1, 2, 21); the movement along the Pasarel fault was probably of more limited extent and importance, and even of lesser importance was the fault relief along the Iskâr fault. In the effect of faulting, the easily erodible volcano became structurally elevated and exposed to rapid erosion; since being surronuded by more resistent rocks, the entire volcanic cone suddenly became a prominent (though highly localized) source for sediment. The associated nonvolcanic rock fragments were locally derived from the upthrown margins of horst blocks.

In the west, phases of rapid subsidence along the Elha fault led to a sedimentary response of rapid progradation of the alluvial wedge and its considerable outward spreading across the Železnitza graben. The direction and highly elongate geometry of this fan suggest that the sediment dispersal was largely confined laterally, and probably controlled by the axis and relief of the Pančarevo graben (Fig. 21). At an earlier depositional stage the fan prism probably was banked against the boundary faults of this graben and, in effect, the sediment dispersal was preferentially downfan, rather than laterally across the fan. Later on, as the sedimentation progressed and the graben relief became gradually subsumed, some distal fan lobes started to be dispersed onto adjacent areas and so the direction of the fan growth became controlled principally by the axis and northeasterly inclined bottom of the Železnitza graben (Fig. 21).

The fanhead segment of the northwesterly trending apron was banked against the active plane of the Elha fault. The growth of this apron may thus be accounted for by a simple model of an elevated basement block (itself containing a localized and highly effective source of material supply) bounded by active faultline scarp with a fan-shaped wedge of alluvial deposits prograding from it (Fig. 21, bottom). At the Elha fault the apron growth was principally related to tectonic adjustment along the fault plane, and was stimulated by pulses of tectonic activity. Characteristic geomorphic expressions of such tectonic settings, which are usually accompanied by numerous earthquakes, include a pronounced fault scarp and abundant material supply from fault-generated avalanches and landslides. Moreover, movement along the boundary fault must have inevitably led to uplifting of the fan-head sediments, and these were then reworked again. Tectonically triggered landslides and avalanches, together with the abundance of easily erodible material, all led to the predominance of direct tectonic control over the source/climatic control on sedimentation. The prevalence of debris-flows and the development of steep, proximally thick, alluvial wedge are natural and expected results of such tectonic-sedimentary setting (*cf.* Steel, 1976, 1977).

In contrast, the easterly extending apron was associated with the dip-slope of the upfaulted basement block bordered by the Elha and Pasarel faults. Thus, the apron was laid down upon the surface of gently warped, relatively passive, easterly tilted block at the elevated corner of which there was located the weathered volcanic cone itself (Fig. 21). Here the fanhead area was probably banked directly against the eastern flank of the volcanic cone, rather than against any prominent fault plane. However, the abovesaid basement block was probably mildly faulted too (*e.g.*, Iskâr fault and Zdravčov fault), and so there might have been also some minor, low-relief fault scarps cutting the contemporaneous sediments of the prograding volcaniclastic wedge itself.

In such tectonic-sedimentary setting the material supply, though still abundant, must have been much less catastrophic than at the Elha fault scarp to the west, and so here the direct role of the source/climatic control on sedimentation appeared relatively more important. In other words, here the role of this latter appears to have been less strongly muted by the tectonic control than it was in fact at the Elha fault on the opposite side of the eroded volcano. Any low-relief topographic elements which pre-existed the redeposition of volcaniclastics were probably easily subsumed and overridden by the prograding apron, and so the sediment dispersal appeared less confined here. The response to such setting was the development of sheet-flood/stream-flow dominated, lower-angle alluvial fan (*cf.* Steel 1976, 1977) on which the sediment dispersal was almost equally vigorous laterally as downfan.

In summary, the major effect of faulting was to produce extensive alluvial aprons of epiclastically reworked volcaniclastics, with contrasting associations of sedimentary facies on the opposite sides of structurally elevated, extinct volcano. The contrasts between the aprons are understandable in terms of a sedimentary response to two different, though mutually related, structural contexts of one and the same tectonic situation. In the light of the herein presented model, the two different kinds of sedimentary response seem not surprising, and are much like the head and tail of one coin.

As already discussed above, the development of the volcaniclastic aprons was followed by an abrupt change in the depositional setting, caused by a sudden uplift in the sourceland to the south and southwest *(i.e.,* in the mountain ranges of Plana and Vitoša). Main palaeoslope suddenly became northeasterly, and huge masses of coarse gravel and sand were stripped away from the hinterland by high-energy flood surges and laid down over the vast area extending north of the Plana Mts. These polymictic fluvial gravels and minor sands were spasmodically deposited within a relatively short time-period, probably during Helvetian/Tortonian.

Taking together, tectonism was the major control of the Tertiary molasse sedimentation in the region. It operated mainly by a control over: *(a)* the magma emplacement and local volcanism; *(b)* the gradient of the topographic surfaces across which sediments were transported; *(c)* the rate of source-area uplift and erosion; *(d)* the rate of basin-floor subsidence; *(e)* the rate of the later uplift in the Plane and Vitoša ranges, and thus also the degree of the erosion and preservation of the pre-existing volcaniclastic aprons; and *(f)* the overall sediment dispersal pattern and, indirectly, the mode of sediment transport and deposition. The sourcerock control and climatic control on sedimentation, though themselves important agencies too, appear to have been largely subdued by the primary effects of the tectonism.

Late Tertiary tectonism produced also the presentday horst-and-graben structure of the region. At this late stage, the broad areas of Tertiary molasse sedimentation became dissected into a number of apparent "sub-basines" containing preserved remnants of the molasse; at this point it is worth noting that the degree of the basin dissection decreases with the distance from the Plana/Vitoša range. As already discussed above, however, the presentday fault-bounded features are not merely grabens and halfgrabens whose age post-dates the Tertiary molasse; most of these features were actually tectonically controlled, "early-stage" basins of the Tertiary sedimentation, which, though largely overswept at the late stage of this latter, became renewed again during a final phase of the region development.

## REGIONAL IMPLICATIONS

From the present study there comes also one other important implication to the Tertiary history of the region, namely the two-stage development of the molasse succession. As shown by the study area, at least some of the Tertiary successor basins in the region apparently were developed and filled in two stages. At an earlier depositional stage, comprising (?Late) Oligocene and Early Miocene, a number of small, variously isolated basins came into existence in the region due to horst-and-graben tectonics; an initial phase of this stage is probably dated by the subvolcanic intrusions and local valcanism related to the fault-controlled magma emplacement. These small basins were gradually filled with sediments derived from strictly local sources, which corresponded to the upthrown margins of the basement blocks; the alluvial aprons composed of epiclastically reworked volcaniclastics are an example. These "early-stage" basins were relatively shallow, and often poorly defined.

At a later depositional stage, comprising Middle and Late Miocene, some broad areas of the basement underwent rapid uplift and erosion (as examplified by the uplift of the Laramian plutonic massifs of Plana and Vitoša), and became sourcelands to the basins originating at their foreland. In effect, the pre-existing "early-stage" basins became incorporated within much more broader basins receiving sediment from the newly uplifted sourcelands. Marginal parts of these extensive basins were spasmodically filled with coarse clastics, but the sedimentation appeared less vigorous towards their centres, where extensive brown coals were able to develop in many areas. Later on, during the Pliocene, these "late-stage" basins underwent tectonic dissection again, and so became split into o number of grabens and halfgrabens separated by variously exposed basement blocks. The tectonic dissection appears to have been largely controlled by the tectonic plan of the pre-existing, "early-stage" basins.

The above-said model for the development and filling of the Tertiary successor basins of the region, in implying two major stages of the molasse sedimentation, is well-evidenced by the characteristic bi-partition of the molasse succession in the study area (Fig. 4). In this latter respect, the present study area appears to be of crucial importance to understanding of the problem. The model presented here seems to be vaild for at least some areas of the Tertiary sedimentation in Western Srednogorie, though may probably appear not easy to prove in the areas which lack any marked difference in the composition of the two respective molasse portions (such as that between the volcaniclastics and the overlying polymict deposits here).

Having arrived at this point, our final suggestion is that the depositional history of the Tertiary molasse in Western Srednogorie is to be re-examined with the use of the herein presented model as a local norm, in order to obtain a regional model possibly through distillation and synthesis of many local examples.

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## STRESZCZENIE

Na obszarze badań — obejmującym SW część Gór Łozen i przyległe partie pasm górskich Płana i Witoša — spod pokrywy osadów trzeciorzędowych odsłaniają się lokalnie skały wulkanogeniczne (fig. 1). Tworzą one kilkanaście izolowanych wystąpień, odpowiadających na ogół erozyjnym rozcięciom („oknom") w samej pokrywie trzeciorzędowej; największe z tych wystąpień, to otoczona skałami krystalicznymi wychodnia wulkanitów na górze Wysoka Ełha (fig. 2). Wspomniane wystąpienia skał wulkanogenicznych interpretowane były przez większość badaczy regionu, jako (silnie zwietrzałe) wychodnie subwulkanicznych intruzji o geometrii „pni" — stanowiących przypuszczalnie apofizy centralnie położonej intruzji Wysokiej Ełhy. Bardziej kontrowersyjny był natomiast charakter petrograficzny tych „intruzji" (trachyandezyty?, ryodacyty?, biotytowe dacyty?) oraz ich wiek (późnokredowy?, trzeciorzędowy?). Niniejsza praca przedstawia gruntowną reinterpretację wszystkich wspomnianych poglądów.

W rzeczywistości na omawianym terenie skały wulkaniczne odsłaniają się jedynie na górze Wysoka Ełha, podczas gdy wszystkie pozostałe wystąpienia rozważanych skał to utwory wulkanoklastyczne (fig. 2). Z przeprowadzonych badań wynika, że Wysoka Ełha stanowi pozostałość (ryodacytowy *neck*) po istniejącym tu kiedyś, a później głęboko zerodowanym, ryolitoidowym wulkanie. Erozyjnemu ścięciu uległ przy tym nie tylko cały stożek wulkaniczny, ale także górna (plagioryolitowa) partia samej intruzji zasilającej pierwotnie wulkan. Erozję wulkanu poprzedziły jego przeobrażenia hydrotermalne oraz silne wietrzenie.

Produktem erozji wulkanu są właśnie wspomniane wcześniej wulkanoklastyki. Buduje je materiał lito- i krystaloklastyczny (przy braku zachowanych składników witroklastycznych), zmienne domieszki materiału niewulkanicznego oraz duże ilości materiału ilastego (produkt wietrzenia); materiał litoklastyczny odpowiada składem plagioryolitom, z przejściem do ryodacytów. W wulkanoklastykach stwierdzono częstą obecność zwęglonego (spalonego) materiału organicznego w postaci transportowanych fragmentów pni drzewnych, gałęzi oraz detrytusu.

Wulkanoklastyki zostały zdeponowane przede wszystkim po północno-zachodniej i wschodniej stronie erodowanego wulkanu, formując dwa rozległe stożki aluwialne (fig. 21). Stożki te różnią się między sobą zespołem facji sedymentacyjnych, a także geometrią. Stożek rozciągający się ku północnemu-zachodowi uformowany został przez grawitacyjne spływy rumoszowo-błotne *(debris-flows)*, którym towarzyszyły podrzędnie procesy fluwialne *(stream-flows)* oraz sporadyczne zalewy warstwowe *(sheet-floods)*. Natomiast stożek rozciągający się ku wschodowi utworzony został w głównej mierze przez zalewy warstwowe *(sheet-floods)*, którym towarzyszyły nieco bardziej intensywne procesy fluwialne *(stream-flows)* oraz rzadkie spływy rumoszowo-błotne *(debris-flows)*. W pracy przedstawiono szczegółową analizę sedymentologiczną odmian wulkanoklastyków (facje) stanowiących produkty wspomnianych trzech typów depozycyjnych. Z paleorekonstrukcji stożków wynika ponadto, że pierwszy ze wspomnianych stożków aluwialnych miał pierwotnie większą miąższość (ponad 350 m) niż stożek drugi (około 200 m) oraz cechowało go także bardziej strome nachylenie powierzchni depozycyjnej (do 15—20° w części proksymalnej).

Wspomniany kontrast facjalno-geometryczny między stożkami aluwialnymi oraz ich układ przestrzenny wyjaśniono w pracy szczegółowo w kategoriach syntektonicznej sedymentacji kontrolowanej przez czynne uskoki (fig. 21). Zwrócono też uwagę na wpływ klimatu oraz przeobrażeń hydrotermalnych na sam przebieg procesów erozji i sedymentacji.

Na omawianym terenie seria wulkanoklastyków zalega zgodnie na polimiktycznych żwirach piaszczystych, które same zawierają także podrzędne ilości wulkanicznych klastów plagioryolitowych. Ta ostatnia cecha oraz lokalna obecność warstw bentonitowych w spągu (prawdopodobnie produkt wietrzenia tufów) wskazują, że wspomniane osady polimiktyczne mogły być w swej znacznej części zdeponowane w czasie działalności wulkanu Wysoka Ełha (wiek zob. niżej). Miąższość tych „spągowych", fluwialnych utworów polimiktycznych jest rzędu kilkudziesięciu metrów.

Seria wulkanoklastyków przykryta jest niezgodnie (niezgodność erozyjna) i przekraczająco przez ekstensywną pokrywę fluwialnych, polimiktycznych „żwirów neogeńskich" (fig. 1), której miąższość sięga 300—400 m. Strop samej serii wulkanoklastycznej jest przy tym silnie erozyjnie porozcinany (różnice reliefu do 200 m), a jej pierwotna miąższość znacznie zredukowana (do 100—150 m). Pochodzenie materiału i kierunki paleoprądów wskazują, że wspomniane „żwiry neogeńskie" deponowane były od strony wypiętrzonych plutonów laramijskich Płany i Witošy (transport ku NE; fig. 21).

Z ustaleń paleobotanicznych wynika, że wspomniana wyżej „pokrywa neogeńska" jest wieku środkowo-późno-mioceńskiego (najprawdopodobniej helwet-torton). Natomiast radiometrycznie datowany wiek „intruzji" Wysokiej Ełhy odpowiada wczesnemu oligocenowi; byłby to zatem także wiek wulkanu oraz w przybliżeniu wiek polimiktycznych osadów podścielających omawiane w pracy wulkanoklastyki (por. wyżej). Wynika stąd, że wietrzenie, erozja i redepozycja samych wulkanoklastyków musiała mieć miejsce w pośrednim przedziale czasowym (zob. fig. 18).

Przeprowadzone badania mają szereg istotnych implikacji regionalnych. Wynika z nich bowiem fakt lokalnej trzeciorzędowej działalności wulkanicznej na obszarze zachodniego Srednogoria; wulkan Wysoka Ełha byłby pierwszym dotychczas zidentyfikowanym tego przykładem.

Na uwagę zasługuje ponadto charakterystyczna dwudzielność trzeciorzędowej serii molasowej na badanym terenie; jej część dolna (dolny oligocen — dolny miocen) to seria wulkanoklastyczna przechodząca w spągu w cienki pakiet osadów polimiktycznych, a część górna (środkowy — górny miocen) to gruboziarnista, polimiktyczna pokrywa żwirowa. Dwudzielność ta wskazuje bowiem na dwuetapowość rozwoju i wypełniania trzeciorzędowych basenów molasowych w tym regionie Srednogoria, co byłoby faktem nowym w geologii regionu. Zaproponowano w pracy pewien ogólny model rozwoju tych molasowych basenów.