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EROSIONAL AND DEFORMATIONAL STRUCTURES IN SINGLE SEDIMENTARY BEDS: A GENETIC COMMENTARY

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Abstract: Erosional and deformational structures from various sedimentary environments are discussed and natural examples are confronted with their experimental analogues. Particular attention is paid to deformation structures following convective pattern of motion. From experimental investigations and field studies it is concluded that many structures, which in current classifications are separated, may result from analogous or similar formative processes.

Abstrakt: W pracy niniejszej przedstawiono rozważania nad pochodzeniem struktur erozyjnych i deformacyjnych w nieskonsolidowanych utworach osadowych występujących w obrębie oraz na powierzchniach spągowych i stropowych ławic. Szczególny nacisk położono na struktury odkształceniowe powstałe w następstwie niestatecznego warstwowania gęstościowego. Rozważania oparte zostały na doświadczeniach przeprowadzonych dawniej przez autora i jego współpracowników a ilustracje struktur doświadczalnych zostały zestawione razem z ich naturalnymi odpowiednikami. Zarówno doświadczenia jak i badania terenowe prowadzą do wniosku, że wiele struktur które często uważane są za niepowiązane ze sobą genetycznie mają jednakowe lub podobne pochodzenie i przechodzą jedne w drugie.

Key words: Tool markings, scour markings, convective deformations, convolute lamination, experimental sedimentary structures.

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INTRODUCTION

This discussion is part of one-semester series of lectures given in 1989, when the writer was visiting professor at the Hebrew University in Jerusalem. The discussion is chiefly concerned with "interfacial structures" in clastic sediments (sensu Dzulynski & Simpson, 1966a) which are either produced on interfaces between one bed and the bed immediately below or above, and on interfaces within single beds. Such structures include current markings produced by current erosion, markings produced by depositing/reworking currents and interpenetration structures resulting from protrusion of one unconsolidated or semi-consolidated sediment into another.

Most of the structures to be discussed are important indicators of sediment transport and, inferentially, give an insight into the paleogeography of ancient sedimentary basins. They also are "way up" indicators showing the original position of strata in folded rocks. In this respect, the following discussion may be regarded as complementary to the excellent and already classic book by Shrock (1948).

Genetic interpretation of the features under consideration is based upon simple qualitative experiments, and pictures illustrating natural structures are compared with those showing their experimental replicas. Where the interpretations here presented differ from generally accepted opinions, the reader can refer to the relevant publications for alternative explanations. The literature on the problems discussed is voluminous and space does not permit due references to all relevant papers. In general, the references will be kept to the minimum necessary for clarity of presentation and to give the credit to the authors whose ideas are similar or identical with those here presented.

The material here offered is not entirely new. It has been published by the present writer and his coauthors in various publications, often with limited distribution. In the following discussion an attempt will be made to bring together this scattered material with emphasis on structures whose origin is still disputable. In this respect, the discussion is a critical reappraisal of earlier studies of the author and his coauthors, before the present writer changed his research interest to other topics.

The list of structures discussed is not intended to be exhaustive, and the paper presented is not meant as a comprehensive manual of sedimentary structures. The structures that are satisfactorily treated in text-books are omitted or only briefly mentioned.

The study of sedimentary structures has now become a matter of routine and application. Consequently, the problems under consideration may be of value to field geologists who use sedimentary structures as a key to paleogeographical reconstructions of ancient sedimentary basins. None of the structures is diagnostic of any specific sedimentary environment nor, in itself, proof of the action of turbidity currents. Turbidite sequences, however, provide particularly good conditions for the appearance and preservation of such structures. Consequently, most of those discussed come from turbidites of the Carpathian flysch.

The specimens, both natural and experimental, which are depicted in this publication are housed in the Geological Museum of the Institute of Geological Sciences of the Jagiellonian University in Cracow. The photographs of these specimens have already been published (e.g. Dżułyński & Ślączka, 1958, 1960; Dzulynski & Sanders, 1962; Dżułyński, 1963a; Dzulynski & Walton, 1965; Dzulynski & Simpson, 1966a) and reproduced in some sedimentological textbooks (e.g. Pettijohn & Potter, 1964). The photographs of natural specimens, however, were rarely brought systematically face-to-face with their experimental analogues. Such a confrontation is made in this paper.

Depending on the publication in which the papers authored or coauthored by the writer were printed, his name was given either in latinized form "Dzulynski" or in the Polish spelling "Dżułyński". This name is here given as in the papers referred.

The structures to be discussed have been divided into two groups: (1) erosional markings produced by scouring action of the currrent or by impact of transported objects (tools) upon a cohesive substratum and; (2) deformational structures resulting from interaction between the flowing and depositing current and the bottom, or from mutual interpenetration of soft sediments differing in density and kinematic viscosity. These two groups of structures will be treated separately, although there are transitions between them. The rest of this paper will be devoted to structures in single beds resulting from impact of heavy suspensions upon a stratified substratum.

EXPERIMENTAL TECHNIQUES

Inasmuch as the forthcoming discussion is based upon experimental investigation, some comments are needed on the experimental techniques applied (for details see Dżułyński & Walton, 1963, 1965; Dzulynski & Simpson, 1966a; Dżułyński, 1966).

Experiments designed to produce erosional markings were similar to Ph. Kuenen's classic experiments (Kuenen, 1958; Kuenen & Migliorini, 1950; Kuenen & Menard, 1952), concerned with intrastratal depositional structures in turbidites. However, instead of sandy suspensions used by Kuenen to simulate natural turbidity currents (abbreviated to t-currents), the artificial t-currents here under consideration were produced by pouring of watery plaster-of-paris mixtures into water-tanks with settled clay on the bottom. The first experiments of this type, were carried out in variously sized but, generally small flumes (Dzulynski & Walton, 1963). Larger flumes up to 2 m long, 0.15 m broad and 0.20 m high, were used in later experiments (e.g., Dzulynski & Simpson, 1966a). These flumes were provided with gates at their distal ends and were submerged in a large water tank. In each experiment the gate was first closed and muds or clay were settled from suspension to produce a bottom layer of suitable cohesiveness. Care was taken to keep the top surface of the settled clay at the level of the chute outlet and the threshold at bottom of the gate. Then the gate was opened and plaster-of-paris suspensions of varying density were introduced to produce artificial t-currents. Opening of the gate prevented the undesirable effect of a reflected backflow disturbing the deposited turbidite. The average current velocity in the experiments discussed, was 9-14 cm/sec.

After hardening, the plaster-of-paris turbidites were lifted and their undersurfaces washed to expose the markings produced. If the turbidite layer was too thin, and therefore fragile, the gate was closed again and additional plaster-of-paris suspension was introduced to settle upon slightly indurated top surface of the turbidite.

The bottom sediments used in the experiments were natural easily slacking clays and coarse technical kaolinite. Depending on the quality of clays, the settling time varied from a few to 24 hours. In the experimental production of tool markings, small objects were placed on the bottom clay near the chute outlet, prior to the introduction of plaster-of-paris suspensions. The objects included water-logged pieces of lignite, fish bones and hard plaster-of-paris discs or spheres. To obtain positives of current markings, the bottom clay was replaced by a layer of sediment deposited from watery plaster-of-paris suspensions containing small amounts of clay to delay the hardening. Then a t-current made of fine sandy or silty suspensions was released. After hardening of the substrate, which in this case took two or three days, the uncemented sandy turbidite was washed away and the positives of markings exposed. For specific purposes, other changes were made in the experimental procedure and these will be described at appropriate places in the following sections.

The techniques applied to obtain syndepositional deformations on bottom surfaces of artificial turbidites were essentially the same as those used to produce erosional markings. The plaster-of-paris suspensions applied, however, were denser than those which gave rise to tool-, and scour markings. As such suspensions began to settle, they formed a soft layer which would interact with the underlying clay, giving rise to deformational structures.

To produce experimental analogues of postdepositional deformations in the absence of current flow, diluted sandy and silty suspensions, mixed with small amounts of plaster-of-paris and clay, were gently introduced into water-tanks with a settled clay to obtain unstable systems with undisturbed members. The development of deformations was than initiated by tapping the walls of the tanks. This resulted in partial liquefaction or softening of the members. Addition of plaster-of-paris did not affect the rheotropic properties of other model materials during the experiments. It made it possible, however, to obtain relatively hard models which, when sliced, gave an insight into three-dimensional aspects of the structures. Additional information concerning the experimental techniques is given when specific structures have been discussed.

In principle, the experiments conducted to produce postdepositional deformations were similar to those designed by other authors to obtain load-induced structures in sediments (eg., Kindle, 1917; Kuenen, 1958, 1965; McKee *et al.*, 1962; Selley *et al.*, 1963), and to simulate igneous diapirs and related interpenetration structures (eg. Nettleton & Elkins, 1947; Ellwell *et al.*, 1960; Ramberg, 1971).

As in the case with experiments on erosional markings, those aiming at producing soft-sediment deformation structures were qualitative experiments. The changing kinematic viscosity (see later) turned out to be one of the most important factors controlling the deformation. The quantitative evaluation of this factor was beyond the capability and competence of the experimenters.

The experiments under consideration have been criticized on the assertion that particle fall-diameter and fluid viscosity were not proportional to each other and to the flow size (Allen, 1971b; Selley, 1988). In fact, the above mentioned parameters were proportional one to each other, and any change in their mutual relationships brought about the appearance of different structures. To satisfy the principle of mechanical similarity, the settling velocity of particles in suspension was scaled down by using materials having specific gravity much lower than that of natural sediments. Admittedly, the parameters in question were not expressed quantitatively and, in this respect, the experiments under consideration may be regarded as somewhat less rigorous. They have yielded, however, the exact replicas of practically all known erosional and deformational structures observed on bottom surfaces of natural turbidites. The experiments have also given insight into processes involved in the formation of sedimentary structures.

No attempt has been made to refer to fundamental laws of hydrodynamics. As is the case with natural turbidity currents, the suspensions and model materials used in the experiments discussed were heterogeneous. When such suspensions begin to settle there is no longer one mixture, but a great number of them, and most of the hydrodynamic laws applied for homogeneous liquids no longer are valid (R. Voughan in Chang, 1939). In this context it may be emphasized, that the real value of experiments in geology is not to produce scale models but to test and investigate certain concepts suggested by field observation and to stimulate the imagination of investigators (Johnson, 1970, p. 131). The aim of the experiments under consideration will be fulfilled if it will prepare the way to further more quantitative investigations.

EROSIONAL STRUCTURES ON INTERFACES OF BEDS

Minor inorganic markings on undersurfaces of clastic beds are referred to as sole markings (Kuenen, 1957).In the following considerations we shall use the term "erosional markings" to indicate sole markings produced directly by erosive current action, prior to collective settling of grains and their final deposition. Such markings, preserved in the form of moulds, have been divided into two categories: (1) tool markings made by the contact of transported objects (the tools) and the bottom and, (2) scour markings produced by the scouring action of the current (Dzulynski & Sanders, 1962a). Moulds of such markings on undersurfaces sandstones are among the most reliable directional structures.

Erosional markings originate in a soft cohesive muddy substratum whose essential property is that it allows the passage of a current without itself being washed away. An incohesive sandy substratum is not conducive to the formation and preservation of minor erosional markings, nor is a hard bottom surface. However, dense masses of water saturated sand, grinding against an earlier deposited sand ("mass-transport marks" of Ricci Lucchi, 1969), leave distinct scratches on the interface separating the welded sand beds (Pl. 3C). It is to be noted that erosional markings are not directly related to any of the Bouma divisions (Bouma, 1962). In most instances they were produced prior to the collective deposition of grains, and only coarse particles entrapped in erosional hollows bear evidence of that interlude which is not recorded in the overlying bed.

Apart from papers authored and co-authored by the present writer, the tool-, and scour markings discussed in this part have been described and illustrated by many authors in publications devoted to sedimentary structures (e.g., Pettijohn & Potter, 1964; Lanteaume *et al.*, 1967). Consequently, the descriptions of such marking will be given in an abbreviated form.

The tools

In flysch basins, the objects acting as tools are either: (1) allochthonous i.e. derived from outside the depositional basin or, (2) autochthonous i.e., derived from within the basin.

The first category includes displaced or redeposited shallow-water fossils, water-logged pieces of wood and exotic rock fragments or pebbles. The second category of tools includes objects which rest on the bottom surface or are buried in the uppermost, unconsolidated part of the bottom sediment. In flysch sediments such tools are hard remnants of free-swimming organisms (e.g. fish) and water-logged pieces of wood. The above mentioned tools are washed out from unconsolidated sediments and, when incorporated into relatively dilute currents, are transported along the bottom giving rise to various markings on contact with the bottom mud. Such tools are occasionally preserved at the downcurrent ends of the markings (see e.g., Dzulynski & Sanders, 1962a, pl. X, XI; Dżułyński & Walton, 1965, fig. 65, 66, 82, see also Pl. 6C & 10A in this paper). The unconsolidated fine-grained bottom sediment is assimilated and dispersed within the flowing t-current, and most of the clay material in sandy turbidites comes from this source (Enos, 1969).

With progressing bottom erosion, semi-lithified mud lumps or lihified shale fragments are ripped up from the bottom and act as tools (Dżułyński & Radomski 1955). Such fragments are occasionally preserved at down-current end of markings (Pl. 2). The presence of hard shales, derived directly from the bottom, indicates that transformation of some clays and muds into a hard sediment may proceed with astonishing rapidity. Parenthetically, it may be recalled that in recent deep seas, unusually rapid overconsolidation may occur at a depth of 0.5 to 1 m below the bottom, and is caused by a chemical intergrain bonding, although actual cementation is not yet visible. Such bonding is attributed to solution and redeposition of silica, calcium carbonate, iron or manganese compounds (Richards & Hamilton, 1967).

The mechanism by which consolidated pieces of mud are torn up is commonly related to differential scouring of the bottom. Crescentic depression which tend to develop on the upcurrent sides of more elevated and usually more resistant patches of the bottom sediment are undercut by vortex erosion. The edges of such patches form projecting overhangs which are then detached and carried away (Pl. 1A and Fig. 1). The illustrations show successive crescent markings

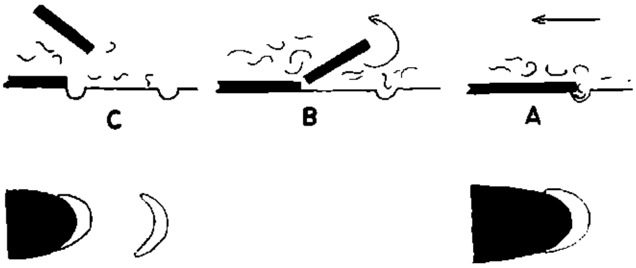

Fig. 1 Diagram showing formations of successive crescent marks. Comp. Pl. 1A

developed in front and along the sides of erosional remnants of once continuous thin layer of an already lithified siliceous shale within the still soft bottom sediment. A contributing factor in the detachment of undercut patches is the settling sand which is forcibly injected into the space under the overhanged shaly projections (Pl. 1B). Analogous situation is shown in Pl. 1C, in which the lower stratified component of an amalgamated bed was detached by a sinking slurry injected laterally along bedding surfaces. Such process may operate on a larger scale and may result in the formation of huge rip-up clasts (Pl. 49A, B) which, if dragged by the current, may produce large grooves.

It is to be noted that hard shales showing a distinct fissility, are ripped up from the bottom as platy fragments which, when rolled by the current, acquire discoidal shapes (Pl. 3B). On the other hand, soft, cohesive and structurally homogeneous muds are tomed out in the form of irregular lumps which, when rolled, become rounded (see also, Ricci Lucchi, 1969). By accretion of small pebbles, such mud balls are transformed into armored mud balls (Pl. 3A).

In spite of the ubiquity of tool markings on the bottom surfaces of turbidites, those with the tools preserved are rare. This is partly due to the fact that one tool may produce a very large amount of markings before it comes to rest. Another reason is that the tools are subject to destruction by repeated impingement against the bottom (Pl. 2B, C). Moreover, when lifted above the bottom, the tools may be prevented from sinking by the settling of sand grains. Indeed, shale fragments are commonly found embedded within the turbidite bed where they are concentrated at specific levels corresponding to drastic changes in density of the flowing and settling sandy suspension (Pl. 3B).

Types of tool markings

The formation of tool markings is a simple process and, in many instances, self-evident from the shape of the preserved mould. In this respect, the experiments have only confirmed the conclusions arrived at earlier from observations of natural markings.

The tool markings are generated by relatively dilute t-currents. In the experiments here discussed, such markings were produced by suspensions in which the ratio of plaster-of-paris to water was 1 to 5 per volume. With high-density currents the same tools were transported above the bottom and no longer came into contact with it (see later).

The tools carried by the current move by saltation, somersaulting, rolling, skimming and dragging. The shape of tool markings commonly provides information concerning the mode of transport. In such instances we may use genetic terms based upon the inferred mode of transportation and collision with the bottom. However, even with respect to such markins the applications of genetic terms may be fraught with error (see later).

For convenience of discussion, the tool markings are divided into discontinuous and continuous types. The former type is due to single or multiple impacts of tools against the bottom. The latter type is made by tools which, while moving, remain in contact with the bottom for an appreciable time interval (Dzulynski & Sanders, 1962a). There are all possible gradations and combinations between these two arbitrarily differentiated types. The inertial conditions which prevail during transportation at the base of t-current cause the transported tools to move along straight lines (Gilbert, 1914). This explains the observed linear alignement of tool markings. Such markings repeated at more or less regular intervals are referred to as skip markings (Dzulynski *et al.*, 1959; Dzulynski & Sanders, 1962a; Pl. 5 in this paper).

The most important types of discontinuous tool markings are as follows:

Prod markings (Dżułyński & Ślączka, 1958; Dzulynski *et al.*, 1959). These are asymmetric in longitudinal profile with gentle upcurrent and steep downcurrent sides (Pl. 4a, Fig. 2A). If, however, the tool hits the bottom at a high angle and then is dragged for a while leaving a short groove behind, the steeper slope of the marking is on its upcurrent side (Dzulynski and Sanders, 1962a, pl. XIII B). In some instances the striking tool is permanently arrested and preserved at downcurrent end of the markings.

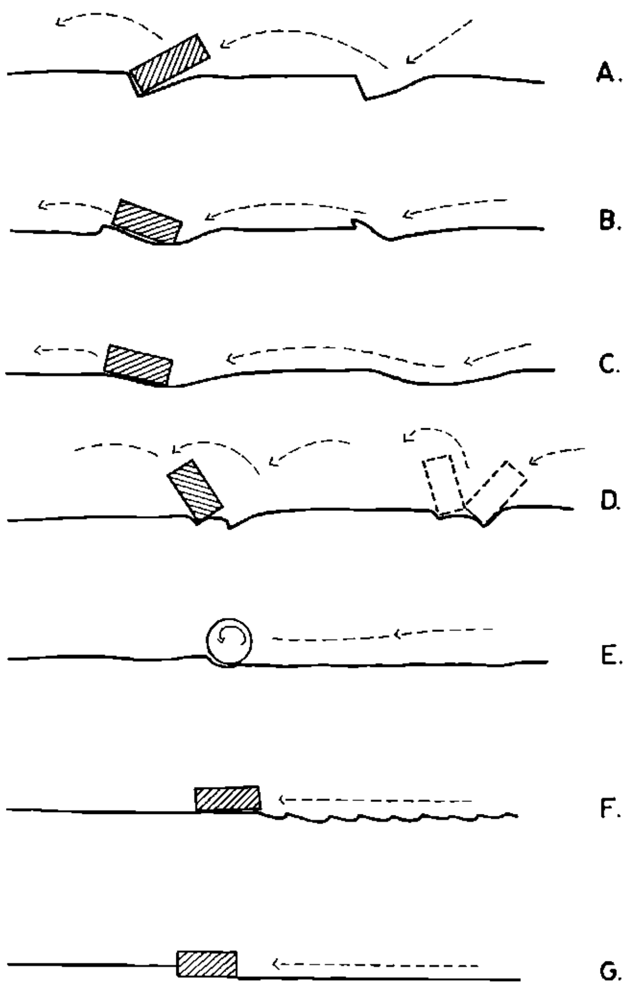

Fig. 2 Diagram showing formation of tool marks; **A** – prod marks, **B** – brush marks, **C** – bounce marks, **D** – saltation and somersaulting marks. **E** – roll marks, **F** – chevron marks, **G** – groove marks

Brush markings (Dżułyński & Ślączka, 1958). These markings show one or more moulds of overhanging mud bulges at their downcurrent end (Pls. 4b & 5). Brush marks are made by tools impinging upon the bottom at a relatively low angle without being halted on impact (Fig. 2B).

Bounce markings (Wood & Smith, 1959) are more or less symetric in longitudinal profile (Pl. 4c, Fig. 2C). In contrast to brush markings, they lack the overhanging moulds of mud bulges at downcurrent end, and indicate only the current's orientation.

Saltation markings. Traces left by leaping and revolving tools are referred to as saltation markings (Dżułyński & Ślączka, 1958, Pls 4d & 10A, B, Fig. 2D). A specific type of saltation markings may be differentiated as somersault or "tumble" markings (Allen, 1984), when the tool hits the bottom alternately with its opposite edges.

The most common types of continuous tool markings are **groove markings**. The markings in question, named by Shrock (1948) "groove casts", were described by early investigators (e.g., Hall, 1843; Clarke, 1917). In most instances, the groove casts seen on the soles of turbidites resulted from dragging of individual objects or patches of semi-consolidated sediments (Fig. 2G). The tools preserved at the downcurrent end of grooves (Pl. 6C), as well as the experimentally produced grooves (Pl. 6A, B) provide evidence in this respect. The twisted, scratched and ruffled grooves (Ten Haaf, 1959; Dzulynski & Sanders, 1962a) and those which terminate like brush or chevron markings result also from dragging (Pls. 6, 7, 9A, B). The above mentioned ruffled grooves are flanked by oblique wrinkles which develop on one or both sides of the grooves. Such wrinkles join the groove downflow at an acute angle (Dżułynski & Sanders 1962a, fig. 6). The grooves mentioned may be indicated by a genetic term "drag marks". (Kuenen & Sanders, 1956; Kuenen, 1957). It may be noticed that, although typically developed "ruffles" join the grooves obliquely downflow, reverse orientations may occasionally occur. In such situations, the ruffles represent the fillings of oblique, shear-induced open fissures produced in cohesive substratum by grooving tool.

Roll markings (Krejci-Graf, 1932). Single groove casts, however, may be due to rolling (Fig. 2E). As noted, the platy shale fragments are generally transported in upright position and, by repeated impinging upon the bottom, may acquire discoidal shapes. Such tools may also produce single and narrow grooves similar to those made by dragging of sharp-pointed tools (Pl. 2A). Broader grooves were made in experiments in which plaster-of-paris t-currents rolled small clay spheres (Pl. 8B, C). Depending upon the absence or presence of a more resistant layer within the bottom sediments, the grooves are deep or shallow. Similar grooves on bottom surfaces of natural turbidites and shown in Plate 8A, might have originated in the same way. A specific type of roll markings are those produced by rolling of fish vertebrae (see later). In some instances, discontinous but regularly spaced markings may result from rolling of fossils provided with extending tubercules (Dzulynski & Sanders, 1962a; Seilacher, 1963).

Tool markings made up of closely spaced chevron-like features repeated many times at regular intervals are known as **chevron markings** (Dunbar & Rodgers, 1957). In longitudinal profile, each chevron is like a brush mark in that it shows a mould of an overhanging mud lip at the downcurrent end. In plan view, the markings are V-, or U-shaped, pointing with their apices in the direction of current flow (Pl. 9). Experimental investigations (Dzulynski & Simpson, 1966a) have confirmed the previous speculative suggestions that these markings are produced by objects skimming over a muddy surface (Fig. 2F). Between ruffled grooves and chevron marks there are all possible transitions (Craig & Walton, 1962).

The above mentioned tool-markings may be included into the category of deformational structures in which the entrainment of bottom sediments is combined with plastic deformation or shear fissures. In this connection it may be recalled that among a variety of reaction which occur between the flowing suspension and the bottom are also flow-transverse shear wrinkles which represent the effects of direct current drag on the hydroplastic bottom surface (Dzulynski & Sanders, 1962a).

Markings produced by fish bones acting as tools

Tool markings made by fish bones are limited to specific facies deposited under euxinic or semi-euxinic conditions (e.g. Oligocene Menilite Beds in the Carpathian flysch, or the Fischschiefer of the same age in the Alps. (Dżułyński & Ślączka, 1958, 1960; Pavoni, 1959; Dzulynski & Sanders, 1962a). The space devoted here to these markings may appear to be out of proportion to their practical value. Such markings, however, show the diversity of features produced by one specific object, depending on the manner of transportation and impact with the bottom.

The most characteristic are ring-shaped markings produced by somersaulting and saltating fish vetebra (Pls. 4d, 10A, B). These markings originate when the centre of the vertebra strikes the bottom with one edge of its articulate face and then rotates forwards so that the opposite edge is pressed against the bottom completing the circular impression. The downcurrent side of these markings is less deeply impressed than the upcurrent side because much of the momentum of the striking tool is dispersed during the first impact. The resulting asymmetry in longitudinal profile indicates the direction of flow. The conclusion concerning the origin of the ring-like impressions has been confirmed by the presence of vertebrae at the downcurrent end of such markings (Pl. 10A) and by experiments (Dzulynski & Walton, 1962, 1965; Pl. 10B).

If the vertebra centre skips over the bottom without somersaulting during each impact, the resulting feature is a series of semi-circular prod marks (Pl. 11C). The "club-shaped" markings (Dzulynski & Walton, 1965) and their experimental replicas originate from a combination of somersaulting and dragging (Pl. 11A, B).

When the centre rolls in continuous contact with the bottom and its rotation axis is in a plane parallel to the bottom surface, the resulting structure is a double-groove cast, occasionally with the vertebra preserved at the downcurrent end of the markings (Pl. 10C). Fish vertebra skimming along the bottom also produce chevron markings (Pl. 9C, D).

In flysch basins with euxinic conditions near the bottom, fish bones are the most common tools gouging the markings. Nevertheless, fish bones preserved at the downcurrent ends of tool markings are very rare. Where they do occur, however, they tend to appear in great quantities to form a real "fossil bonanza" ("fossil lagerstatte" in the meaning of Seilacher, 1993). The sudden deposition of large amounts of tools is presumably due to specific critical hydrodynamical conditions of the flow. It also indicates that the tools were transported in swarms (e.g., Mellen, 1956).

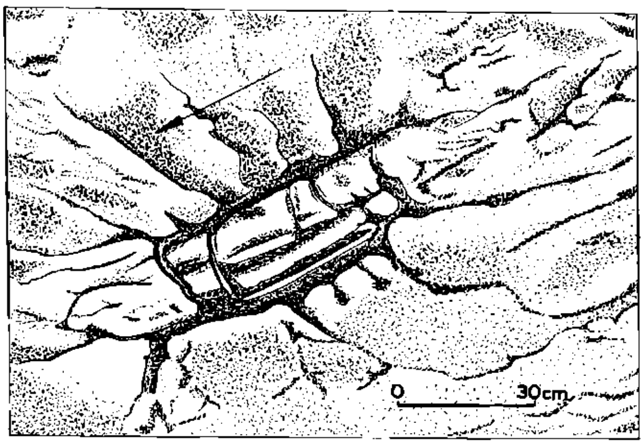

Fig. 3 Break-through prod mark. Upper Createaceous Flysch. Valea Leurzi. Rumania (after photograph)

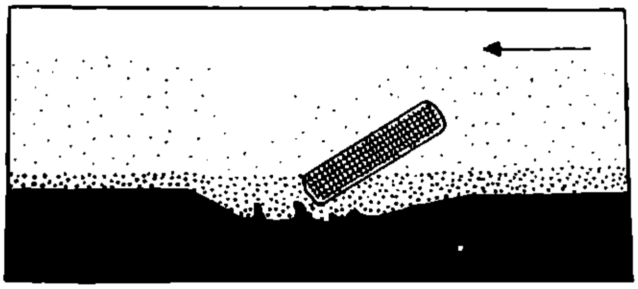

Fig. 4 Diagram showing formation of break-through tool marks

Break-through tool markings

With the onset of collective deposition of grains, the sole markings are gradually covered with a protective mantle of sand. In early stages of deposition, when the growing basal layer of sand is still thin, the transported tools may break through this mantle. They may gouge new markings in the underlying mud or deform and obliterate the earlier markings (Fig. 3). The impact of such tools may reactivate the scouring action and give rise to flute markings superimposed upon earlier features (Pl. 12A, B).

If the initial layer of settling grains is already too thick to be perforated by the proding tool, the impact of large tools may buckle down the base of accumulating turbidite. With close packing of grains and a deficiency of intergranular water, such down-buckling is commonly associated with the appearance of dilatation fissures (Dzulynski & Walton, 1965, fig. 96, see also Fig. 4).

Intrastratal tool markings

With progressing deposition, the interface between the settled sediment and the flowing suspension gradually shifts upwards away from the original bottom surface. The processes involved in the formation of sole markings may and, commonly do operate. The transported tools impinge upon the already deposited sediment. Such sediment, however, consists of incoherent grains in which the formation and preservation of markings is inhibited, unless the alternating deposition of finer grades brings about the appearance of more cohesive laminae that are firm enough to preserve the shape of markings. If this be the case, the tool markings may occur as syndepositional features on intrastratal parting surfaces (e.g. Hubert *et al.*, 1966).

Scour markings

The following general types of scour markings are here distinguished:

Obstacle scours (Dzulynski & Sanders, 1962a). The best known are current crescents (Peabody, 1947), developed in front of temporary or permanent obstacles (Pl. 1A, Fig. 1). These structures were first described by Hall (1843) and produced experimentally by Rücklin (1938). Morphologies of current crescents and the nature of the flow field associated with their formation are well known from a number of publications (e.g., Karcz, 1968; Richardson, 1968; Allen, 1984; Paola *et al.*, 1986).

Flute markings or flute casts (Crowell, 1955). These markings occur as moulds of elongate subconical hollows flaring out and merging with the bedding plane in downcurrent direction. The upcurrent ends of such markings are bulbous, sharp-pointed or gently rounded (Pls. 12, 13 & 14). The flute markings may reveal a distinct asymmetry and may begin with a spiral welt. General morphology of flute markings has been comprehensively discussed by many authors. Rücklin (1938) was the first to produce flute markings experimentally by passing a thin sheet of clear water over muddy substratum of unspecified structure. Allen (1968, 1971a, 1984) studied experimentally the development of flute markings formed through the solvent action of water and corrasion, by allowing water to flow over plaster-of-paris slabs and clay, in the presence of various obstacles or "bottom defects" (Allen, 1971a). Those defects that exceeded a certain critical dimension, with respect to the thickness of boundary layer, developed into flute markings parabolic in plan. Defects which fell below the critical dimension gave rise to elongate groove-like hollows. Allen (1971b) obtained variously shaped counterparts of natural scour markings, depending upon the duration of current action and the spatial distribution of bottom defects capable of initiating the turbulence. He treated the problem of scour markings in mathematical terms and the results of his experiments have been summarized in his excellent monograph (Allen, 1984). Dzulynski and Walton (1963) succeeded in producing small flute markings, using plaster-of-paris t-currents which they believed to be devoid of tools. In fact, however, the suspensions used in these experiments contained small lumps of congealed plaster-of-paris.

Flute markings are thought to originate from the combined action of vertical and sub-horizontal eddies. Allen (1971a) suggested that three classes of bottom irregularities initiate flute markings: (1) trace fossils, (2) forward or backward facing steps and, (3) objects distributed through the body of bottom sediment which differs in resistance from the enclosing mud. The significane of such irregularities or bottom defects, is confirmed by field observations and experiments.

The experiments carried out by Dzulynski and Simpson (1966a) have shown that there is yet another important factor promoting and controlling the development of flute markings, even on perfectly flat and homogeneous mud surfaces. This factor is the presence of tools in the flowing suspension. At the moment of impingement upon the bottom, the tool acts as a temporary obstacle generating vortices capable of scouring the bottom. In this connection, it is to be recalled that narrow sharp-pointed tools resting on the bottom with their longer axes oriented parallel to the direction of flow do not give rise to current crescents. The streamlines are gently split on the upcurrent end of such tools and reattach behind them to form a pair of horizontally oriented and counter-rotating spiral eddies. Along the central line of such double-eddies system, the relatively faster moving liquid above the tool is pushed downwards, increasing the flow velocity and scouring the bottom mud to produce elongate holes. The scouring is increased when the downcurrent end of the tool is asymmetric and, therefore, the reattached streamlines have different velocities. In such situation, also eddies with vertical axes are formed.

Returning now to tools transported by the current, it is to be remembered that elongate or platy objects, when suspended in the flowing suspension, tend to assume an upright position with their longer axes on a plane which is parallel to the main flow (Owen, 1908). Such tools strike the bottom with their narrow edges and, at the moment of impingement, behave in much the same way as the previously mentioned narrow obstacles. They generate vortices at the downcurrent end of the impinging tools. Such vortices persist for only a very short time but, given suitable consistency of the bottom and the appropriate density of the current, they are able to scour flute marks downcurrent of the proding objects (Dzulynski & Simpson, 1966a).

As noted previously, the tool markings are products of relatively dilute t-currents. With a slight increase in density of the t-current (under given bottom conditions and current velocity), the prod markings show recognizable flute markings at their downcurrent end (Pl. 13A, B). The indentations produced at the instant of collision, represent the bottom "defects" which once formed, may promote a further scouring after removal of the tool.

With a still higher proportion of plaster-of-paris to water (2:3) in the flowing suspension, the indentations become smaller. The scouring associated with tool impingement, facilitated by an increased hydraulic uplift, proceeds to the extent that an original prod mark is entirely obliterated by the growing flute marking (Dzulynski & Simpson, 1966a). Indeed, among the experimental and natural flute markings there are all gradations between tool markings modified by scour and flute markings bearing no obvious record of the tool impingement (Pl. 14). With still further increase in density of the suspension and the remaining conditions unchanged, the tools are transported above the bottom and no current markings are being formed.

In Allen's (1971a) experiments on scour markings, the obliteration of original bottom defects was a function of time. In experiments by Dzulynski and Simpson (1966a), the same effect was obtained by an increase in current density and, thus, its eroding power. The increasing erosion, in this case, was caused by decrease in the magnitude of the gravity force due to the buoyancy effect (see also Wright, 1936; Knapp, 1943). It is realized that, with respect to scour markings in flysch, both factors i.e., time and increased density, may cooperate and complement each other.

It should be pointed out that Allen (1971a) passed critical judgement on experiments by Dzulynski and Simpson (1966a). In his opinion, the structures obtained by the above authors cannot be interpreted as erosional markings because their production could not be observed and the truncation of bottom laminae was not reported (Allen l.c. p. 189). It is true that, the formation of markings could not be directly observed but the resulting structures were the exact replicas of natural flute markings. The absence of truncation of bottom laminae is not an argument against the erosive nature of the markings in question, because the experiments were made in flumes with non-laminated clay at the bottom.

Inasmuch as the interpretation of flute markings as presented by Dzulynski and Simpson (1966a) has not been generally accepted, it might be expedient to summarize the arguments in its favour. These arguments are as follows:

(1) There is a direct relationship between the amount of tools in the plaster-of-paris suspension and the amount of flute markings produced on the base of artificial turbidites. When only a few tools are transported, the flute moulds are scarce and commonly arranged in zones corresponding to the path of the skimming tools (Pl. 13B). When the tools are transported in isolated swarms, there is a zonal arrangement of flute markings separated by zones with fewer flutes, a phenomenon observed among natural markings. With a large amount of tools, the bottom surfaces of artificial turbidites are entirely covered with a multitude of flute markings (Pl. 14A, B).

(2) There is an obvious relationship between the size of tools and the size of flute markings. When large tools are used in experiments, the resulting flutes are large (Pl. 13C). Small tools give rise to small flute markings.

(3) The shape of experimental flute markings, at least to some extent, depends upon the shape of the tool and the way it hits the bottom, e.g., the platy or needle-shaped objects usually produce sharp-nosed flute markings (Pl. 13A & 14B, C).

(4) In the absence of tools, the experiments conducted under the previously mentioned bottom and flow conditions, failed to yield analogues of flute markings.

Summing up, there is a reliable experimental justification, supported by field evidence, that the presence of saltating tools in flowing suspension is an important factor controlling the development of flute markings on smooth and homogeneous mud surfaces. Thus viewed, tool markings and flute markings are not entirely contrasting types of current structures as previously regarded by Dzulynski and Sanders (1962a).

The problem of rolling tools as a possible factor promoting the development of narrow elongate flutes is open to question. The experimental elongate flutes shown by Dzulynski and Simpson (1966a, fig. 12) were produced in the presence of small rounded tools in the current. However, similar structures, resembling groove moulds, are also formed at downcurrent sides of small remnants of vertical organic tubes acting as stationary bottom defects.

As noted, the shape of experimental flute markings, among others, depends on the shape of the tool and the way it impinges upon the bottom. At the instant of impingement, the tool may deviate the streamlines in one or other direction, giving rise to asymmetric scouring. Consistent asymmetry of flute markings in plan view and their diagonal arrangement may appear where the current flow is curved (Dżułyński, 1965a, Fig. 12).

The moulds of flute markings, in vertical cross-sections, are either flat-bottomed or concave downwards (Pl. 14A, B). Flat-bottomed flute markings are formed in stratified substratum in which a thin soft top layer is underlain by a relatively hard layer that is strong enough to resist the current scour. In this respect, the experiments confirm the conclusions arrived at by field observations. Concave flute moulds are produced in unstratified and more or less homogeneous substrate in which the resistance to current scour increases gradually with depth.

Where the bottom mud contains thin silty laminae, the surfaces of flute moulds reveal minute terraces which record the position and number of such laminae. The silty laminae are more susceptible to lateral erosion than the enclosing mud. The scouring of these laminae produces bench-like recesses on the walls of the scoured hollows which are thus preserved in the sediment filling the flute markings (Dzulynski & Sanders, 1962a).

Attention should be called to moulds of rounded or, if closely packed, polygonal equidimensional scours which alone have no directional value. Markings of this type, indicated as "bulbous" or "pillow-like" markings (Dżułynski & Walton, 1965) were produced experimentally by Allen (1971a). They originate when a uniform and steady flow is tranformed into chaotic turbulent motion (Pl. 30). These features can be interpreted in terms of both, erosion and deformation phenomena and will be dealt with after convective polygonal, scaly and linear deformation structures have been discussed.

Transverse scour markings The structures described as transverse scour narkings (Dzulynski & Sanders, 1962a) are oriented with their longer axes at right angle to the direction of flow and have an asymmetric cross-section, being steeper on the upcurrent side (Pl. 15). They superficially resemble moulds of transverse ripples and, indeed, were first regarded as casts of current ripples by Dżułyński and Ślączka (1958, pl XXV, fig. 2). According to Dzulynski and Sanders (1962a), the markings in question are erosional features because they are observed to pass into regular flute moulds (Pl. 15B). A compromise view has been proposed by Dżułyński and Walton (1965) who argued that these features are the products of transverse flow-rolls combined with current shear. Structures resembling transverse scours were obtained experimentally under unspecified hydrodynamic conditions by Dżułyński and Walton (1965, fig. 41 B). The best replicas of these structures, however, were produced by Allen (1971a, figs. 91 & 93) who interpreted their origin in terms of "fluid stressing".

The presence of horizontal terraces on the side walls of incipient transverse markings, similar to the previously disussed terraces on the walls of flute moulds, appears to indicate that the structures in question, indeed, are erosional features produced by transverse eddies ("Wasserwalzen" – Salomon, 1926) in horizontally laminated substratum (Pl. 15A).

Rill markings. The structures described as rill marks (Dżułyński, 1963a; Dzulynski & Sanders, 1962; Dżułyński & Walton, 1965) consist of slightly sinuous and anastomosing upcurrent moulds of flow-parallel runnels which begin with sharp-pointed gouge moulds (Pl. 16A). The exact analogues of rill marks have been produced experimentally by Allen (1984). The origin of these features is not fully understood, though they may represent the effect of secondary cross-currents inherent to flowing suspensions (see later).

Equally enigmatic is the origin of rill markings arranged in diagonally crossing sets (Dżułyński 1963a; Pl. 16B this paper). The pattern of these features is similar to rhomboidal pattern of bedforms investigated experimentally by Karcz and Kersey (1980) and described from recent beach surfaces (e.g. Otvos, 1965). It bears also some similarity to braided and overlapping patterns of channels on alluvial fans and in broad river valleys (comp. Rachocki, 1981).

Channel-fills (Williams, 1881)

Bottom surfaces of current-deposited beds commonly show variously sized and deeply incised erosion furrows whose amplitude may exceed the thickness of host beds. The moulds of such features, known also as "gutter marks" (Whitaker, 1973), may occur in the form of entirely isolated bodies (Pl. 17C). The side walls of such bodies are commonly covered with flute markings which indicate that the erosion furrows were not immediately filled with sand.

Large and laterally extensive furrows are due presumably to localized concentration of streamlines and increased bottom erosion (e.g. Whitaker, 1973; Flood, 1981; Aigner, 1985). Narrow furrows might have been carved by larger tools dragged by the current and then resculptured by minor vortices, prior to collective settling of sand particles. According to Whitaker (1973), the entirely separated gutter casts are produced where insufficient amount of sand is available to form a persistent bed. In proximal parts of sedimentary basins, however, the bulk of sand carried by strong currents might have been swept away leaving only the grains trapped in deeply incised gutters (Myrrow, 1992).

Channel fills have been chiefly reported from shallow-water sediments, deposited under subtidal conditions or in environments subject to intense storm activity. Isolated channel fills, however, occur also in deep-water flysch sediments (Wood & Smith, 1959; McBride, 1962; Walker, 1966). The examples here presented also come from deep-water sediments deposited in proximal parts of a flysch basin. The channel fills may be produced through by-passing t-currents and incompletely realized depositional episodes. In poorly exposed shaly flych, such episodes may escape attention of geologists and this, as well as the presence of laterally wedging out turbidites, may negatively affect statistical analyses concerning the frequency and periodicity of depositional episodes in flysch rocks.

Superposition and coexistence of various types of current markings

Although the time interval between current erosion and mass deposition of grains can be very short, it may last sufficiently long for the development of two or more generations of current markings superimposed one upon another. Such superposition is observed among natural and experimental current markings. Scour and tool markings may be formed simultaneously or one after another in any order. The successive generations of markings may cross each other at various angles (see Dżułyński, 1963a, pl. LVI & LVII; Dzulynski & Sanders, 1962a, pls. VIII B & X A). Such cross-cutting arrangements reflect local deviations from the main flow direction within a single current. In this context, it is to be recalled that while measuring the paleocurrent directions, it is necessary to take into account all directional structures in sufficiently large observational domains in order to eliminate local deviations. Such deviations, commonly motivated by bottom irregularities, are observed on the bottom surfaces of natural and experimental turbidites.

In flysch sediments, current markings may point to different direction of flow in successive beds (Pl. 17A) or even within a single bed. It is not out of place to mention here an unusual situation in which current markings pointing exactly in opposite directions coexist, side by side, on the same bottom surface of a turbidite bed (Pl. 17B). Such situation has been reported from an Oligocene turbidite section in the Polish Carpathians in which fine-grained sandstones deposited by currents coming from NW alternated with coarse and thick-bedded greywackes laid down by currents coming from the SE (Dżułyński & Ślączka, 1957). In this section, there is an amalgamated turbidite bed (now inundated by dam lake) whose lower part consists of a fine-grained arcosic sandstone and the upper part was made-up of a coarse greywacke. Both parts, representing tightly welded different turbidites, are separated by an irregular erosion surface. In places, the fine-grained portion was entirely removed by the current flowing from SE, which produced flute markings and deposited its load on the same mud surface that now delineates the sole of the amalgamated bed. In such places, flute markings pointing in opposite directions were observed It is to be noted that the above mentioned alternating types of sandstones differ drastically in their petrological and structural composition. The fine-grained sandstone show ripple cross-stratitification and a poor assemblage of most resistant heavy minerals. The coarse turbidites are massive and contain lithic fragments of igneous and metamorphic rocks. They also contain contorted clasts of fine-grained sandstones ripped up from the bottom and displaced backwards in the direction where the fine-grained material came from.

Although tool and scour markings may be formed simultaneously or one after another in any sequence, there are situations in which either of these two types of structures predominates. As previously indicated, the predominance of the one or other type depends, among other factors (e.g., number of tools present, flow velocity, cohesive strength of bottom sediment), upon the density of t-currents. Some examples of natural and experimental turbite soles showing the above mentioned coexistance are illustrated in Plate 18A, B. The structures depicted include features which will be discussed in the second part of this paper.

Relationship between the size of erosion markings and the thickness of beds

Intuitively one would expect large and heavily laden currents to produce large current markings. Indeed, large scour and tool markings tend to occur on undersurfaces of thick turbidites which are attributed to the action of large and powerful currents. There is, however, no direct relationship between the size of current markings and the thickness of the host-turbidite. The soles of thick turbidites may be covered with small markings and, vice versa, the thin turbidites may reveal markings whose depth is of the same order or even greater than the thickness of the covering bed. Such inconsistency is explained in many ways because many factors determine the presence or absence of current markings and their size. Inasmuch as a considerable part of tool markings in phanerozoic sediments is produced by wood fragments and hard skeletal elements of organisms, one would expect less tool markings in the Precambrian sediments than in sedimentary rocks deposited later. The Precambrian turbidites, however, reveal both tool-, and scour markings (e.g., Emery, 1964), but these were produced or motivated, only inorganic hard fragments such as shale fragments derived from the bottom or allochtonous pieces of rocks.

Distribution of erosional markings

Erosional markings may develop on intrastratal discontinuities and on the top surfaces of large mud clasts enclosed in thick turbidites (Dżułyński, 1963a, fig. 38).

Sole markings are generally absent on bottom surfaces of turbidites deposited on rapidly lithified siliceous muds. Significantly, such turbidites are generally devoid of dispersed clay particles and commonly represent pure quartzites. Fluxoturbidites (Dzulynski *et al.*, 1959), laid down in proximal parts of depositional basins, are also devoid of minor sole markings. The absence of markings is here due to the high density of the flow, whereby smaller tools are transported above the bottom. Moreover, the great erosive power of such flows removes unconsolidated muds exposing more lithified muds or sands of the subjacent fluxoturbidite. In the latter case, the fluxoturbidites become amalgamated, but the interface between them may show parallel groove markings or striations produced by shearing of moving sand masses upon the sandy substratum ("mass transport markings" in the meaning of Ricci-Lucchi, see also Pl. 3C in this paper).

Erosional markings tend to occur in proximal parts of sedimentary basins. Farther away from the source of clastics, the turbidites become progressively impoverished in erosional markings. This is due to the fact that turbidity currents lose their erosive power and most of the transported tools with distance.

With the introduction of the fan concept into flysch sedimentology, some authors suggested that the terms "proximal" and "distal" be abandoned with respect to turbidites (e.g., Nelson *et al.*, 1978). This is justified if the notion of facies is associated with these terms. However, inasmuch as experimental studies confirmed the proximal-to-distal decrease in the amount and size of sole markings, both terms retain their usefulness as indicators of the distance from the source of clastics (Walker, 1980; Macdonald, 1986).

DEFORMATION STRUCTURES

We come now to structures whose pattern is, to a considerable extent, determined by the geometrical and dynamical characteristics of the settling suspensions and the mechanical properties of the bottom sediment. Some of such structures may be interpreted in terms of both, erosion and deformation the processes which, by no means, are mutually opposing and may overlap each other in time and space. For example, steady currents moving over incohesive sand may set the grains in motion and redistribute them, particle by particle, according to the flow pattern. The redistribution is one of the attributes of erosion. On the other hand, currents of the same character, but flowing upon viscous or plastic substratum, may deform it without much if any entrainment. As long as the pattern of deformation reflects the pattern of streamlines, the resulting structures may show morphologies similar or identical with those produced in loose aggregates of grains.

We are interested here with structures resulting from instability in density stratification. Specifically, we are interested in structures reflecting more or less regular convective patterns of motion. In many instances it is impossible to designate such structures in terms of their time of formation vis-a-vis the time of deposition (Dżułyński & Walton, 1963). Such structures may form simultaneosly with deposition, e.g., when a sandy suspension flows over and sinks into a hydroplastic substratum. They may also originate as post-depositional features under subaqueous or subaerial environments and in buried sediments. In the latter case, the deformation is consequent upon softening or liquefaction of the already deposited sediment. Germane to the problem under consideration is reversed density gradient i.e., the superposition of a heavier material upon a lighter one. Indeed, many of the structures to be discussed are generated by reversed or negative density gradient. Such structures, however, may also originate along interfaces across which there is an unbalanced positive density gradient. This occurs when the lower and denser material becomes softened or liquefied and the overlying lighter material sinks to the level of hydrostatic equilibrium. Similar deformations also occur in dichotomous layered systems with positive density gradient, whose top surfaces are loaded and the material of the upper layer is pressed down into the subjacent layer. The reversed density gradient, however, can be created in dichotomous systems with positive density gradient by liquefaction of the lower member (Allen, 1977).

In agreement with Cegła and Dżułyński (1970) and Anketell *et al.* (1970), simple dichotomous systems with reversed or unbalanced density gradient will be referred to as (ba) systems, where (b) will stand as the symbol of the higher member and (a) of the lower member. (ab) will stand as the symbol for positive, i.e., non reversed density gradient.

Our chief concerns are: (1) deformational structures on undersurfaces of beds which can be directly and geneticaly correlated with deformations in the uppermost part of the subjacent bed; (2) deformations within the bed and; (3) deformations on top surfaces of beds, consequent upon mutual interpenetration of subjacent sediments.

The structures to be discussed fall, at least partly, into the category of "rheotropic structures" (sensu Conybeare & Crook, 1968). This term is used here in its purely literal sense as an indication of a pronouced increase in mechanical mobility of otherwise solid sediments.

Dispensing ourselves from any attempt to create a new classification, we shall concentrate on specific structures using conventional and descriptive terms whenever possible, with restricted use of genetic terminology (comp., Rodgers, 1954; Mills, 1983). The leitmotiv of the following considerations will be the structures on bottom surfaces of turbidites which are generally included into the category of sole markings. The recognition of processes involved in the formation of such markings will help in clarifying the origin of genetically related intrastratal and top-surface deformations. A complete review of structures resulting from instability in density stratification is out of question. The following considerations are directed along selected lines with attention focussed on structures showing a more or less regular convective pattern of deformation. The succeeding discussion is a retrospective analysis based upon qualitative experiments by Dżułyński (1963a, 1965a, 1966), Dzulynski and Walton (1963), Dzulynski and Simpson (1966), Dżułyński and Radomski (1966) and Anketell *et al.* (1969, 1970).

Environments promoting development of unstable (ba) systems

Unstable (ba) systems may occur in any environment. There are, however, environments which are particularly conducive in this respect. Such environments include deep-, and shallow-water turbidite basins. The instability arises here from rapid episodic and alternate deposition of coarse and fine-grained clay-rich sediments. Although it is not of direct consequence to the forthcoming discussion, some comments on the flysch facies are here not out of order, because all the previously discussed erosional markings and most of what will be discussed in the following sections, concerns the structures in flysch turbidites. The term "flysch", introduced by Studer (1827), has been repeatedly misused, subject to arguments (for references and details, see: Dżułyński & Smith, 1964; Enos, 1969; Hsü, 1970) and, rejected as no longer usable (e.g. Eardley & White, 1947; Miall, 1984). The argument commonly stems from the fact that the flysch sequences are implicitly or explicitly identified with turbidite sequences by geologists with little or no experience with the flysch facies and the Alpine orogeny. As indicated by Fairbridge (1958), the term flysch is extremely valuable designation for which there is no substitute in literature. If we set aside this designation we are left with thousands of local litho-stratigraphic terms to indicate a specific facies which is present in the majority of orogenic belts. The flysch is a deep-water marine facies of widespread horizontal extension and great thickness, whose coarse components were deposited chiefly by gravity mass-movements, among which turbidity currents played a dominant role. Implicit in this definition is the presence of deep-water hemipelagic and, occasionally, pelagic sediments. The depositional environment of flysch may be envsioned in the form of marine basins flanked, at least from one side, by tectonically active borders. There is a good reason to suppose that such borders consisted of step-fault terraces which might have acted as final preparation sites and transient reservoirs for clastics, prior to their subsequent transfer into the flysch basins. In such reservoirs, the clastics could have been subject to a prolongued action of longshore currents and waves. The terraces envisioned would be characterized by tectonic instability manifested in terrace tilting and, consequently, in sporadic debouchement of the previously accumulated clastics. Intuitively, thrust faults, would appear to be more prolific producers of clastics then normal faults (extensive zones of factured and crushed rocks). To explain vast quantities of sand in flysh sequences, appeal may be made to erosion of unconsolidated or weakly consolidated sedimentary rocks in source areas of clastics and to suitable climatic conditions. In younger flysch zones account should be taken to redeposition of older flysch sediments (auto-cannibalistic stage of development).

The terrace tilting (Dzulynski, 1979) may be considered as only one of several possible debouchement mechanisms. The mass movements resulting from slope failures due to overloading and/or spontaneous liquefaction, triggered off by earthquakes (the flysch is a syntectonic facies) might have occurred concurrently and initiated turbidity currents (Morgenstern, 1967).

Turbidite sequences, however, also occur in shallow-marine and lake deposits invaded occasionally by sediment-laden river floods. The lacustrine turbidites differ from those in flysch sequences only by the presence of features diagnostic of shallow-water environment, e.g. the presence of wave ripples, subaerial drying cracks, foot imprints of birds and land animals or roots of plants (e.g., Panin, 1962; Jerzykiewicz & Norris, 1994). The pelagic sediments which in flysch facies indicate the temporary cessation in delivery of clastic material may, in shallow-water turbidite sequences, be replaced, among others, by autochthonous coal seams.

Another favourable environment is provided by subaerial sedimentary basins (playa, temporary lakes) in semi-arid climatic zones with seasonal rainfalls. In such an environment, the instability in density stratification is generated by rain waters depositing coarse clastics upon dry mud. At the beginning, the dry mud supports the weight of accumulating coarse material. The deformations follow a downward infiltration of water and the succeeding softening of mud. Such deformations are facilitated by the scarcity of vegetation and by the presence of an impervious substratum below the active surficial soil layer (McCallien *et al.*, 1964).

Conditions promoting the development of unstable density stratification also occur in periglacial environments. Such conditions include; (1) alternate deposition of coarse and fine clastics by glacial streams, (2) deposition of coarse clastics by melt waters upon frozen mud followed by thaw;

(3) scarcity of deeply-rooted vegetation; (4) the presence of permanently or temporarily frozen ground which ensures abundance of water in active soil during thaw and; (5) frost heave which is among the processes producing reversed density gradient.

The Pleistocene periglacial zones of Central Europe were much more favourable to the formation of unstable density stratification than the present-day polar regions. The following reasons account for this: (1) the Pleistocene ice-sheets extended far into the lower altitudes where relatively warm summers and diurnal fluctuations of temperature induced more melt-waters than it is the case in higher altitudes. Consequently, the changes in the discharge of melt-waters were thus more drastic and frequent; (2) in Europe, the Pleistocene continental glaciers invaded areas covered with weakly consolidated clay-rich Pliocene sediments. Therefore, the glacial streams were more heavily loaded with fine-grained sediments than the present-day streams emerging from continental ice-sheets. The best conditions for the formation of sedimentary ba systems were provided by extensive outwash-plains with countless lakes, ponds and braided rivers. The outwash accumulations were built up chiefly during the melting of continental ice-sheets. Consequently, the deformational structures resulting from unstable (ba) systems were more likely to be generated during the recession of glaciers (Butrym *et al.*, 1964; Cegła & Dżułyński, 1970).

Conditions promoting the formation of unstable (ba) systems also occur on the contact surfaces of magma bodies differing in density and viscosity. The resulting deformations are analogous to those observed in sedimentary rocks.

General principles

As a preliminary to the following discussion some general principles, pertinent to the problems under consideration, are desirable.

Trigger systems

Sedimentary (ba) systems contain a certain amount of potential energy stored during their accumulation. Such systems may remain in a state of metastable equilibrium as long as the bearing capacity of the lower member is sufficient to support the overlying member. If, however, the strength of the lower member is reduced, e.g., by liquefaction and/or yields to superior load, the energy stored will give rise to deformation across the interface separating members differing in density. Inasmuch as the work on the deformation is expended by the stored energy, even insignificant impulses or "trigger forces" may result in conspicuous disturbances. The nature of impulses may be irrelevant to the pattern of deformation and may not be recognizable from the resulting structure.

Structural homogeneity vs. heterogeneity

A body is said to be structurally homogeneous if any discontinuities that exist in it are penetrative features, i.e., are small and repeated at distances so small as compared with the size of the body that they can be considered to pervade it uniformly and be present at any point (Turner & Weiss, 1963). Such discontinuities do not affect the pattern of deformation. Non-penetrative features, on the other hand, influence the pattern of deformation. The recognition of structural homogeneity and heterogeneity, in a statistical sense, will serve as one of the principal elements in differentiation of the deformations under consideration.

Kinematic viscosity

Kinematic viscosity is a measure of mobility of fluids and is expressed by equation $k = f/d$, where f = coeficient of friction and d = density. The relative differences in kinematic viscosity of the layers involved in deformation are important factors controlling the pattern of plastic and viscous deformations.

Spontaneous liquefaction

In loosely packed water-saturated muds, silts and sands, the previously mentioned transformations from solid to liquid conditions is referred to as spontaneous liquefaction (Terzaghi, 1956). Such transformation results from temporary loss of intergranular contacts between self-supported particles, accompanied by an increase in hydrostatic pressure. The liquefaction is commonly induced by shock or overloading and may affect simultaneously the whole sedimentary layer. It may also spread in the form of a moving liquefaction front. The manner in which the liquefaction affects the sedimentary layer may be one of the factors determining the pattern of deformation. It should be borne in mind, however, that the majority of deformations here discussed occur in sediments showing plastic or soft behavior. Between liquefaction and plastic deformations there are gradations. Contrary to liquefied sediments, those which deform by plastic flow retain the vestiges of their primary structure.

Dilatancy

Granular aggregates in condition of dense packing resist deformation and fail by fractures, if there is no liquid available to fill the increased proportion of voids (see, Reynolds, 1885; Mead, 1925; Brodzikowski 1981). When the available liquid is sufficient to fill the expanding voids, then the aggregate deforms by plastic or viscous flow.

Pattern of temperature induced convective movements

Inasmuch as a considerable part of the forthcoming discussion will be concentrated on deformations following incipient convective patterns, it would be appropriate, at this point, to include a few comments on this subject.

The term "convection" is applied to regular movements induced by reversed density gradient in fluids which are uniformly heated from below (Benard, 1901). In such a situation, a low density layer is generated at the bottom and the instability thus produced sets up the fluid in motion. The pattern of motion depends on the absence or presence of horizontal shear between the top and bottom of the fluid layer (see, Graham, 1934; Chandra, 1938). In the absence of horizontal shear, the instability in density stratification leads

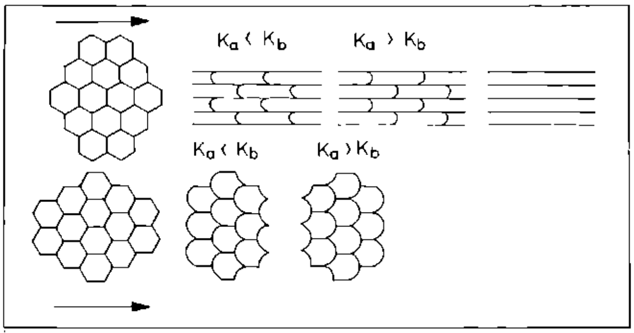

Fig. 5 Transformations of hexagonal convective pattern in presence of horizontal shear. Explanations in text. Modified after Graham (1934)

to the appearance of cell-vortices which, under conditions of perfect homogeneity of the members involved, acquire hexagonal outlines in plan view. When the upper zone of the fluid layer has a lesser kinematic viscosity than the lower zone, as is the case with heated gas, the cell-vortices show descending centers i.e., the motion is downwards along the axes and upwards along the sides of the cells. If the lower member has lesser kinematic viscosity than the upper one, as is the case with liquids, the motion is upwards along the axes and downwards along the sides of the cells.

In the presence of a relatively small horizontal shear between the top and the bottom of the fluid layer, the polygonal pattern is transformed into a scaly pattern (Fig. 5). Depending upon the orientation of the polygons in respect to the direction of horizontal shear, the scales are arranged in rows or in an "en-echelon" pattern. If the kinematic viscosity of the upper zone is smaller than that of the underlying lighter zone, the scales face with their convex sides up-flow. With reversed viscosity distribution, the scales face with their convex sides down-flow (Fig. 5).

With a relatively high rate of horizontal shear as compared with the vertical vector of motion, the cell-vortices are transformed into longitudinal double-rolls with opposite

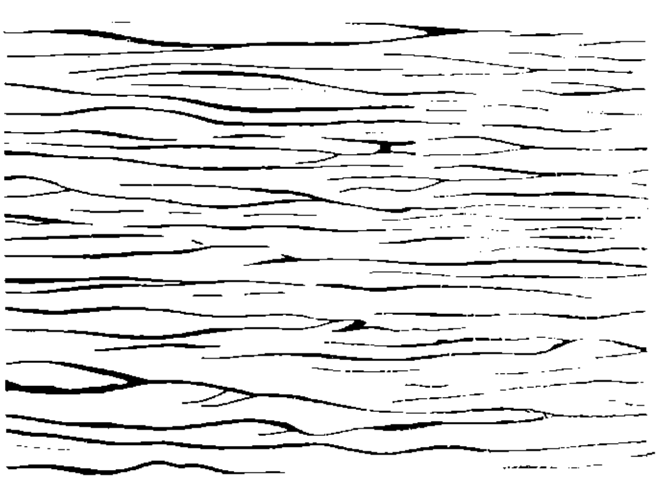

Fig. 6 Pattern of convective rolls in experiments on fluids heated from below. Horizontally mobile (ba) systems. Modified after Graham (1934) and Chandra (1938). Explanations in text

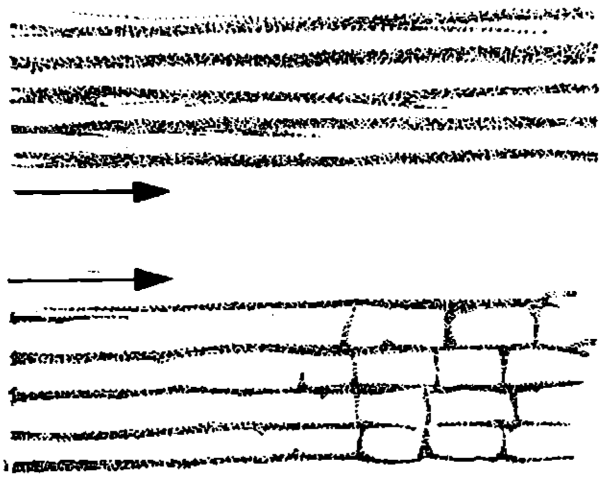

Fig. 7 Patterns of clay suspensions moving across interface between saline and fresh water. Polygonal and square patterns form where forward motion stops. Explanations in text. After Dżułyński (1966)

sense of rotatation in adjacent rolls which, visualized by suspended smoke particles, occur in the form of shear-parallel stringers (Fig. 6).

With decreasing shear velocity, the motion in rolls breaks up into polygonal or rectangular compartments. The size of convective cell-vortices and rolls in dichotomous convective systems is roughly proportional to the joint thickness of the members involved in deformation.

Convective pattern of motion in isothermal (ba) systems

The previously mentioned patterns find their close analogues among the patterns of sedimentary structures resulting from instabilities in density stratification. We shall discuss these analogies more fully in the context of specific structures. Some comments are here needed on convective patterns of motion in flowing suspensions. Such patterns may be demonstrated by simple experiment in which a diluted semi-transparent clay and/or coal-dust suspension is gently introduced into a glass tank containing hypersaline solution beneath a layer of fresh water (Dżułyński, 1966; see also Sonnenfeld & Hude, 1985). The hypersaline solu-

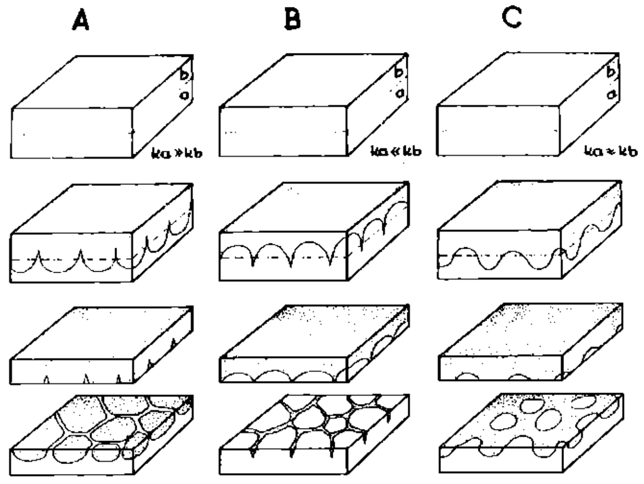

Fig. 8 Incipient deformations in non-mobile bipartite (ba) systems comprising statistically homogeneous members deforming by viscous or plastic flow. **A**, **B** and **C** show configurations resulting from different initial ratios of kinematic viscosities (k_a/k_b). After Cegla & Dżułyński (1970) and Anketell *et al.* (1970)

tion is slightly colored by potassium permanganite in order to make the interface between the saline and fresh water visible. The clay suspension released from a chute placed at the level of this interface, moves across it in the form of a suspended density current or an interflow. Being slightly heavier than the hypersaline solution, the suspension, sinks slowly while moving forwards. This gives rise to spiral cross-currents in longitudinal double-rolls, whereby the clay particles are arranged into flow parallel stringers (Fig. 7). The spiral motion is occasionally made visible by the finest clay particles which, when caught by cross-currents, describe helicoidal trajectories. The bulk of the particles are pushed forwards along straight courses. The pattern of longitudinal stringers is resolved into a network of polygonal cell-vortices when the forward motion ends, provided that there is still suspension available to ensure unstable density stratification (Fig. 7). The steady flow in longitudinal rolls represents the first stage in a laminar-turbulent transition process (Sparrow & Husar, 1969). It is, strictly speaking, neither laminar nor turbulent and has also been described as "subturbulent" (Dżułyński, 1966). Throughout this paper, the deformations following convective patterns will be referred to as "convective deformations".

Factors influencing the pattern of sedimentary convective deformations

It appears from the foregoing considerations that the presence or absence of lateral shear between the members of unstable systems is an important factor controlling the pattern of convective deformations. The (ba) systems in which there is no lateral shear will be referred to as horizontally non-mobile systems and those in which such shear exists, as horizontally mobile systems (Anketell *et al.*, 1970). For the sake of brevity, these will hereafter be referred to as non-mobile and mobile systems repectively. The above distinction will serve as a basis for subdivision of convective structures discussed in this paper. The manner in which the disturbance proceeds along the interface between (a) and (b) is yet another factor which exerts influence on the pattern of convective structures in (ba) systems yielding by plastic or viscous flow. Two extremes will be considered: (1) the disturbance affects simultaneously the whole interface and, (2) the disturbance progresses laterally along the interface between a and b. It is realized, however, that between these two extremes are gradations.

The structures in question will be discussed taking into account the previously indicated controlling factors which, to a considerable extent, are interdependent. Inasmuch as the materials of different composition may behave in similar ways under deformation, we shall first discuss the structures in general terms, referring to specific properties rather than to specific composition.

Incipient convective deformations in dichotomous non-mobile (ba) systems

Consider a simple two-member (ba) system deforming by plastic or viscous flow in which the deformation affects the whole interface between (a) and (b) simultaneously. Depending on the ratio of kinematic viscosities k_a/k_b, the interface may be deformed into one or another of three types of incipient deformations (Fig. 8).

Type A. Where $k_a > k_b$, the material of the upper member descends in a series of convex-down lobes. The movement proceeds downwards, along the axes of lobes and upwards, along their sides. The material from the lower member is pinched up between the sinking lobes to form sharp-crested ridges (Fig. 8A).

Type B. Where $k_a < k_b$, the lighter material of the lower member rises in the form of a series of concave up lobes. The ascending motion is in the center of the lobes and the sinking along their sides. In this situation the upper layer material is pinched down between the rising lobes in the form of sharp-crested pendants (Fig. 8B).

Type C. Where k_a and k_b are approximately equal, the interface is deflected sinusoidally as downward moving lobes from the upper layer interfinger with similarly sized upward moving lobes from the lower layer (Fig 8C).

The above configurations, produced simultaneously with deposition or as post-depositional features, have been indicated as central, marginal and neutral, respectively (Artyushkov, 1965). The genetic relationship between the viscosity distribution and the type of incipient deformation may be demonstrated experimentally in the following way (Dzulynski & Simpson, 1966a, see Fig. 9). A plaster-of-paris suspension containing relatively high proportion of sand particles is released to flow as turbidity current upon clay substratum. The presence of sand particles increases the kinematic viscosity of the suspension. Consequently, the convex-up lobes (type B) are formed in the proximal part of the flume. Farther away from the chute, the amount of sand particles decreases and the convex-up lobes produced on the bottom surface of the resulting turbidite pass through deformations of type C into convex-down lobes (type A).

When the rising or descending lobes come into mutual

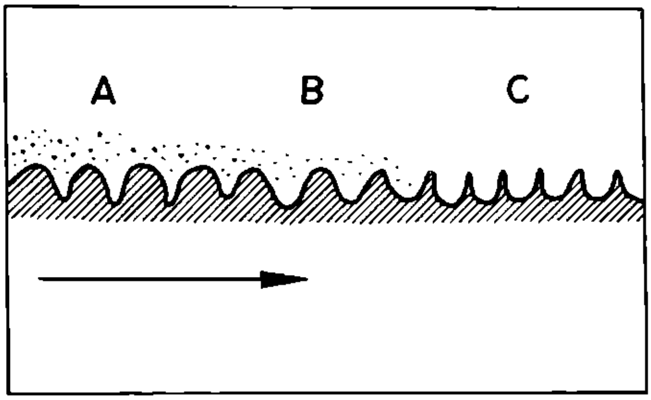

Fig. 9 Diagram showing vertical section through artificial turbidite showing upright lobes passing, with decreasing amount of sand in suspension (*dotted*), through sinusoidal to pendulous lobes. After Dzulynski & Simpson (1966a)

contact, they acquire polyhedral or hexahedral shapes. The latter shape is produced under conditions of perfect homogeneity of the members involved in deformation. In case of Type A, the material of the lower member is squeezed upwards and attains its greatest vertical extension at triple points i.e., in places where three adjacent polyhedrons meet. In the case of Type B, it is the descending material of the upper member that at triple points attains its greatest vertical extension. This is due to the fact that the triple points receive the largest amount of the material pushed aside by expanding dome-like lobes. The significance of this observation will be discussed later. The size of incipient structures is roughly proportional to the combined thickness of the members involved in deformation.

The incipient deformations A, B and C merit particular attention as basic forms for a series of derivative structures. They may also be preserved permanently in the form of discrete sedimentary structures (see below). This may be due to; (1) the expulsion of water and the attendent increase in strength or, (2) when descending or ascending lobes are barred from further motion by a resistant layer or the top surface of an evolving (ba) system. For such incipient deformations, the presence of differential loading is not obligatory. If the vertical movements are not arrested, the evolving (ba) system may go through a succession of deformational stages shown in Fig. 10. The transformations depicted, represent a general trend of deformations that may eventually lead to the establishment of positive density gradient - a situation rarely achieved under natural conditions. Each stage of the above transformations may be stabilized in the form of a discrete sedimentary structure.

The previously mentioned simple relationship between the form of deformational structure and kinematic viscosity does not hold true for the above mentioned transformations. This is due to the fact that the viscosity changes during deformation and the deformational structures produced, represent already non-penetrative features which influence the pattern of successive transformations.

SEDIMENTARY STRUCTURES SHOWING NON-MOBILE CONVECTIVE PATTERNS

Features on bottom surfaces of beds

Syndepositional structures of type A on undersurfaces of current-deposited sandstones, generally referred to as load casts (Kuenen, 1953), are characteristic of subaqueous deformations because the settling water-saturated suspension is, in most instances, more mobile than the settled bottom clay. In such situations, the incipient deformations seldom if ever go through the succession of transformations depicted in Figure 10, because the settling sand is prevented from a further sinking into the lower bottom layers by the appearance of increasingly more consolidated sediments.

Natural and experimental load casts corresponding to type A may be bulbous or flattened when in contact with a more resistant bottom layer (Pls. 19 & 20). The flattened load casts may reveal bilobate contours in vertical cross-sections with small hollows in the center which represent moulds of point injections of clay. Such point injections arise from a composite nature of larger compartments in which the movement takes the form of incipient ring-like vortices (see later).

The size of the compartments is roughly proportional to the joint thickness of the layers involved in motion. In most instances, it is the lowermost part of turbidite beds that is involved in deformation, because the expulsion of water following deformation and the attendant compaction stiffen the muddy substratum and the lower portion of accumulating turbidite. Thus, the appearance of relatively small load casts on the undersurfaces of thick-bedded sandstones indicates that the load casts were formed prior to deposition of the whole bed, but after the accumulating sand has reached the thickness corresponding to the height of the flame structures between the load casts. In general, the size of load casts vary from small "pimples" (Pl. 21A, B; see also Tanaka, 1970, pl.

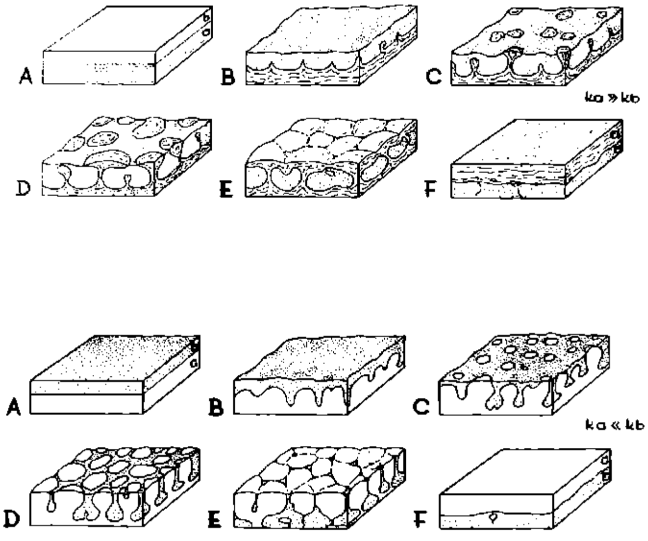

Fig. 10 Selected stages of deformation in non-mobile (ba) systems for different ratios of kinematic viscosities. After Cegła & Dżułyński (1970) and Anketell *et al.* (1970)

VII 2) to features measuring many tens of centimeters in diameter. In plane view, the contours of load casts may be roughly polygonal or rounded (Pls. 19 & 20). Such contours indicate that the load casts in question were formed in the absence of unidirectional flow at the bottom. Very irregular contours may point to limited multilateral flowage of the settled but unconsolidated suspension ("flow casts" in the meaning of Shrock, 1948; see also Prentice, 1956; Kuenen & Prentice, 1957).

Load casts of type A on bottom surfaces of sandstones may also originate as post-depositional features. In experiments, this occurs in (ba) systems whose members are loosely packed and prone to shock-induced liquefaction. The morphology of such structures is identical with those produced simultanously with deposition. However, while the syndepositional load casts represent in general single deformational episodes, the postdepositional ones may result from repeated impulses and may pass through the succession of transformations depicted in Fig. 10. Load deformations of type B and C may occasionally occur on bottom surfaces of current deposited beds, however, in sub-aqueous environments such deformations are rare. The above mentioned load deformations should not be identified with those produced by unequal loading and exhibiting forms which may be correlated with non-penetrative structures of the overlying layer (Kelling & Walton 1957).

Intrastratal deformations

Intrastratal polygonal deformations of type A may occur within sandstone beds on intrastratal discontinuities covered with thin films of finer sediment, which represent short-lived temporary lulls in the settling suspension. The structures of this kind, as seen on exposed parting surfaces, occur in the form of polygonal or square ridges, commonly associated with longitudinal ridges (Pl. 36A).

Intrastratal deformations known as "trendless-pattern convolutions" (Ten Haaf, 1956) or "casque-shaped" convolutions" (Anketell & Dżułyński, 1968a) are closely related to the above mentioned features. Such deformations develop in dichotomous beds whose upper member is rippled, laminated or massive, and whose lower member is horizontally laminated (Pl. 22), provided both members are statistically homogeneous. The laminated sediment of the lower member is squeezed upwards in the form of narrow sharp-crested anticlines, between the downpending bowl-shaped bodies produced by sinking of the upper member. Consequently, in all vertical cross-sections, the structure reveals the "cusp and pin-up" pattern (term after Selley *et al.*, 1963), which is characteristic of convolute lamination. At triple points, the upward squeezed material may acquire mushroom-shaped forms.

The structure in question, is produced in the absence of horizontal shear by sinking of the upper member to an equilibrium level, after much if not all of the bed has been deposited. The deformation is initiated by softening or partial liquefaction of the lower member, resulting as a consequence of overloading. The ubiquity of casque-shaped convolutions in thick sequences of strata makes other causes (e.g. earthquake shocks) less likely. Parenthetically, it may be added that in the Cretaceous Inoceramian Beds of the Carpathians, the majority of sandstones in thick rock sequences may show this kind of deformation.

The above discussed type of convolutions might have been described from different localities by many authors. In most instances, however, the illustrations of apparently analogous structures are given in one vertical section only, leaving thus the true shape of the deformation and hence its characterization as casque-shaped convolution conjectural.

The casque-shaped convolution should not be confused with "ripple load convolutions", resulting from sinking of regularly distributed large ripples into laminated substratum (see later).

Deformations of type B inside sedimentary units showing reversed density gradient, occur in systems whose lower members are subject to liquefaction resulting in a relative decrease in kinematic viscosity. The rising columns of the less dense material may be arrested in their ascent and, if closely spaced, acquire polyhedral shapes before reaching the top surface of evolving (ba) systems. Pl. 23A shows the upper surface of the deformed lower member with "frozen" ascending columns. The structure depicted was produced experimentally by depositing a loess layer upon the substratum made of plaster-of-paris/water mixture and clay. The deformation was shock induced and the surface exposed by washing away the loess. The same kind of deformation, however, may be initiated by other impulses, e.g., overloading. Natural equivalents of such surfaces are known to exist among subsurface periglacial structures (Richmond, 1949; Rohdenburg & Mayer, 1969) and are rightly regarded as a prelude to the formation of polygonal soils. Similar, though isolated columns (pillars), have been reported from the Tertiary Molasse of Switzerland (Dzulynski & Smith, 1963, fig. 5) and produced experimentally in unstable (ba) systems (Dzulynski & Walton, 1963, pl. XXIII b). It is to be emphasized the the surface shown in Plate 23A is a mirror image of polygonal load casts on the bottom surfaces of natural sandstones and experimental turbidites (Pl. 19). Viewed this way, the polygonal soils may be considered as reverse analogues of polygonal load casts (see later).

A specific case of deformation is provided by non-mobile (ba) systems in which member (b) consists of cohesionless grains and the viscosity of the member (a) is greatly reduced. In such situation, the instability takes the form of closely spaced descending stringers of isolated grains (Dzulynski & Walton 1963, pl. XXIIa; see also Fig. 11). The structure is analogous to narrow "fingers" which develop at an unstable interface between two fluids moving vertically through a saturated porous medium (Wooding, 1969). It is to be noted that similar structures were also obtained experimentally by Kuenen (1965, 1968) who introduced the term "settling convection" for this and similar types of deformation. Presumable analogues of the above mentioned deformations appear to be the "auto-injections" described from Aberystwyth Grits of Wales by Wood and Smith (1959) or the vertical sheet structures described by Laird (1970) and interpreted by him as due to convective movement initiated by differential settling of grains. As yet another possible

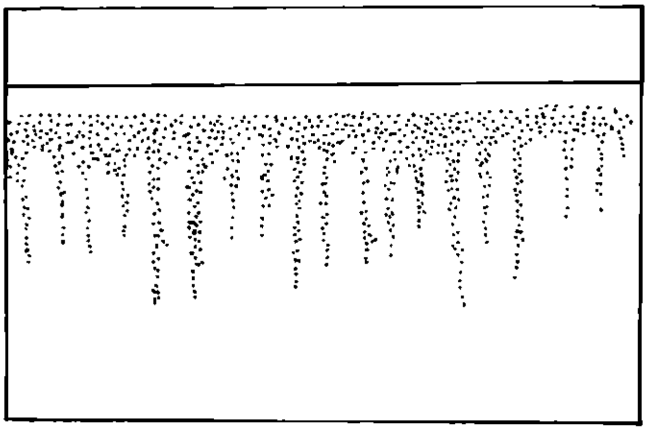

Fig. 11 Vertical striping produced experimentally by sandy threads sinking into very soft mud. Experimental structure. After photograph in Dzulynski & Walton (1962)

natural analogue one can mention the "helictic flame" structure, described by Brodzikowski and Haluszczak (1987, fig. 4) from Quaternary glacio-deltaic sediments of Poland and interpreted as shock initiated post-depositional deformations.

Top surface expressions

If the upper member of an unstable (ba) system is thin, the previously mentioned polyhedral columns may break through it, producing on the top surface of the system, first, roughly circular patches which correspond to the highest parts of the rising lobs (Fig. 12B, C). When the flow of the lighter material continues, the patches grow larger pushing aside the material of the upper member. In a further development, when the margins of expanding circular patches come to mutual contact, the resulting structure shows polygonal pattern of pendent wedges (Fig. 12D, Pl. 23B). If the upper member contains platy fragments, these are set up in upright position in the previously mentioned wedges bordering the polygonal compartments. The structure bears a striking similarity to polygonal soils described from cold climatic regions (Pl. 24A).

The formative process, through which the polygonal structures were produced in experiments, has been suggested as an alternative explanation of natural sorted polygons formed under subaerial conditions in dichotomous soil mantles (Dżułyński, 1963a, 1966; see also Butrym *et al.*, 1964). Artyushkov (1965) came to similar conclusion on the basis of theoretical considerations combined with field observations. In the above interpretations, the processes leading to the appearance of polygonal soils are analogous to those that give rise to convective load deformations induced by unstable density stratification in soft water-laid sediments.

It is to be noted that the concept of convective movements has already been suggested to account for polygonal soils (Low, 1925; Gripp, 1927; Gripp & Simon 1933; Mortensen, 1932). In most instances, however, it was based on the assumption that such movements were set up by differences in density of water at 4° and 0° C. Such differences, however, turned out to be insufficient to initiate convective

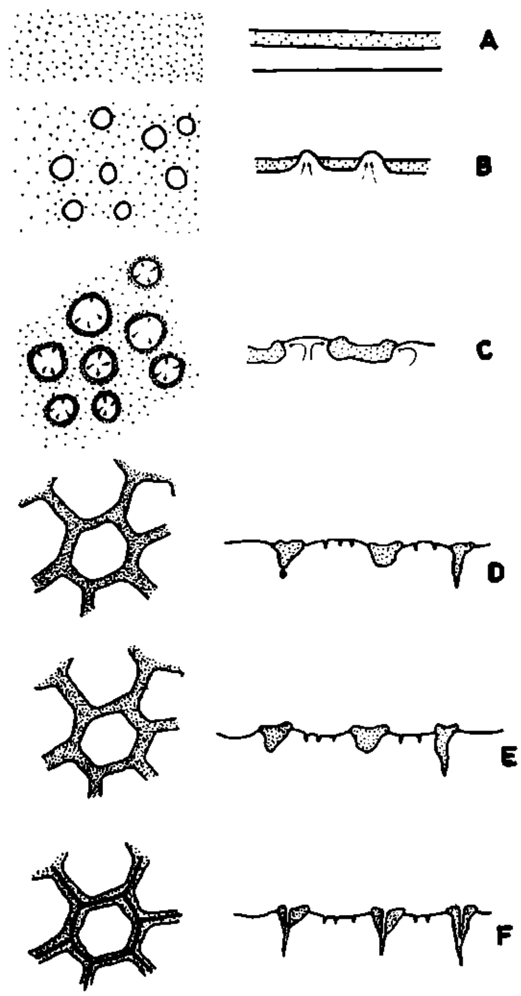

Fig. 12 Formation of polygonal soils from instability in density stratification in plan view (*left*) and cross-section (*right*). Based on experiments. Upper layer; sand and lower clay. After Dżułyński (1963b)

movements in clastic (ba) systems. Consequently, the concept of aqueous convective currents has been abandoned in favour of multigelation hypotheses (for references see; Washburn, 1969). In this context, it is to be noted that Sorensen (1935), in his important though not sufficiently exploited publication, pointed out that the superposition of coarse debris upon finer materials is not the result of processes which lead to the formation of polygons but vice versa. During the polygon-forming processes, the coarse rock debris is being pushed horizontally and sinks along the sides of polygonal compartments. Sorensen also emphasized the importance of the upward flowage of water-saturated fine materials in the formation of natural sorted polygons (see also; Gripp, 1952).

The convective concept of polygonal soils, as here presented, does not require a direct interference of frost action in the formation of such soils, although such action may play the role of impulses starting the deformational process (Jahn & Czerwiński, 1965) and provide preconditions for the formation of sorted polygons. The concept fits well with what is known on conditions attending the formations of polygonal soils. The developmental stages of experimental polygons (Fig. 12) also are identical with those reconstructed for natural occurences on the basis of field observations (Elton, 1927).

Inasmuch as the genesis of polygonal soils and, in general, the origin of "patterned grounds" (term after Washburn, 1956) is still disputable, some additional comments on this subject are not out of place here. The explanations concerning the origin of polygonal soils have three basic assumption in common: (1) the polygonal soils and related structures are features indicative of cold climatic regions and are in one way or another consequent upon the action of frost, (2) the structures are obviously associated with dichotomy and reversed density gradient in stone-rich soils affected by deformation and; (3) they tend to develop on flat surfaces in water-logged sediments or soils (e.g. Hamberg, 1915; Sharp, 1942; Czeppe, 1961). It is to be noted that polygonal soils, once formed, are relatively stable and, in most instances, inactive (see also Washburn, 1969).

The first contention is not necessarily true. It is the second and third contentions which are the key-notes of the structures in question. Under periglacial conditions the sorted polygons, tend to form on flats, where the dichotomy and reversed density gradient are produced by frost heave or deposition of coarse clastics upon frozen mud. The deformation is initiated by the onset of thaw.

The sorted polygons, however, are not limited to periglacial zones and do not provide an independent test of cold climatic conditions. As already noted, the unstable (ba) systems, and thus the preconditions for the formation of sorted polygons, may also occur in warm semi-arid sub-aerial environments with seasonal rainfalls. One of the most instructive examples, in this respect, has been recently provided R. Soja (in preparation) who described well developed sorted polygons from the subtropical region of Sivaliks in India (Pl. 24B). These polygons were formed on an alluvial plain covered with dry mud on which, at the onset of rainy season, the swollen river waters laid down a layer of gravel of approximately uniform thickness. The deformation, which finished with the appearance of sorted polygons was started by infiltration of water into the mud and its consequent softening.

The possibility that sorted polygons may be formed under subtropical conditions was briefly mentioned by Romanovsky and Cailleux (1942). Such polygons were thought to originate on impervious muddy substratum. The above mentioned authors, however, explained the formation of polygons in terms of temperature motivated density differences in the materials involved in deformation. Hunt and Washburn (1960) also described polygonal structures, resembling those in periglacial environments, from evaporitic sediments. In this connection, reference should be made to brief information by Kaiser (1965) on one field experiment in Boulder, Colorado in which a gravel layer was laid down on weathered crystalline material. Small polygons were developed after 5 years. However, no explanation nor details were given on these polygons. Recently, Ahnert (1994), obviously unaware of relevant literature, described sorted polygons produced in the absence of frost action on parking lots and other places, where loose gravels were spread on "unobstructed groud surfaces". Ahnert, basing on theoretical model experiments, considered these polygons in purely statistical terms without specifying physical conditions and processes contributing to and attending the formation of the structures in question.

Polygonal structures similar to sorted polygons also occur on recently exposed surfaces of near-shore sediments in front of temporary innudated canyons emptying into the Dead Sea in Israel. The surfaces are covered with a carbonate cemented crust underlain by soft and mobile clays. The extrusion of clays and the consequent breaking of the crust resulted in the formation of roughly polygonal compartments with ascending centers (type B). The platy fragments of the crust, accumulated along the margins of polygons, became oriented in roughly upright position. Where the exposed sediment consisted of several crusty layers with clays between them, the resulting structure was less regular and was similar to some Pleistocene "cryoturbations" in thin-bedded limestones with marly intercalations. Analogous polygonal patterns have been described and interpreted in terms of reversed density gradient by Dionne (1971) from recent tidal flats along the east coast of James Bay in Canada.

The polygonal structures evolving from incipient deformations of type B are rarely encountered in subaqueous environments. The manganese nodules, however, reported by Lonsdale (1950) from the floor of Atlantic ocean at depth of 3,000 m and arranged in a polygonal network, might have been formed in unstable (ba) system, in the previously suggested manner.

In the hitherto discussed polygonal soils, the motion was upwards in the center and downwards along the periphery of incipient compartment (type B). It should be realized, however, that polygonal patterns may also develop from incipient configurations of type A, in which the motion is downwards in the center and upwards along the periphery of evolving "cells". In such situations, the coarser and heavier material of the member b is concentrated in the center of polygons (as in "stone-centred polygons" of Rozansky, 1943).

If the upper member of an evolving (ba) systems is not homogeneous in statistical sense, or the vertical movements are associated with limited multilateral horizontal shear, the top surface expressions of deformations are irregular.

In fairness to earlier investigators, it should be pointed that differential or unspecified loading was invoked to account for polygonal soils and related periglacial structures by several authors (e.g., Elton, 1927; Mortensen, 1932; Dewers, 1934; Sharp, 1942; Klaczyńska, 1960; Halicki, 1960; Czeppe, 1961; Bülow, 1964). These explanations closely approach the concept here advocated which, in fact, has already been accepted by some investigators (e.g. Rohdenburg & Meyer, 1969; Cegła, 1973; Eyles & Clark, 1986).

Drop structures

In sorted polygons of type B, the maximum concentration of the top material is in triple points. This induces a

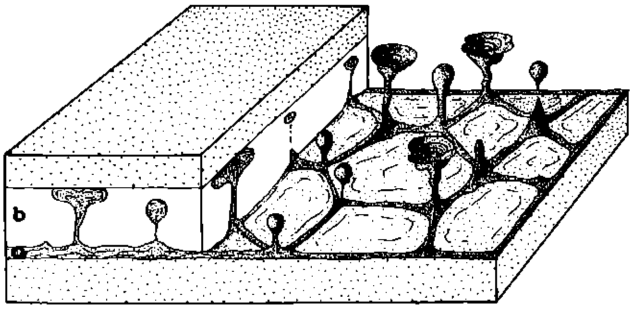

Fig. 13 Formation of polygonal soils from instability in density stratification in plan view (*left*) and cross-section (*right*). Based on experiments. Upper layer: sand and lower clay. After Dżułyński (1963b)

more rapid and deeper sinking of the material (b), giving rise to the appearance of drop structures (Pl. 25A). On encountering a resistant layer, the drops become flattened. The polygons with such "frozen" drop structures are referred to as "rooted polygons" (Romanovsky & Cailleux, 1942).

In polygons of type A, it is the lower member material which, at triple points, is squeezed farthest upwards (Fig. 13). The natural equivalents of such reversed drop structures have also been described (e.g. Lea, 1990; see Fig. 14).

The drop structures may also originate from concentration of heavier material in the centres of polygonal compartments, if the motion is downward along the axes of such compartments (Pl. 25C, D). The concentration of material

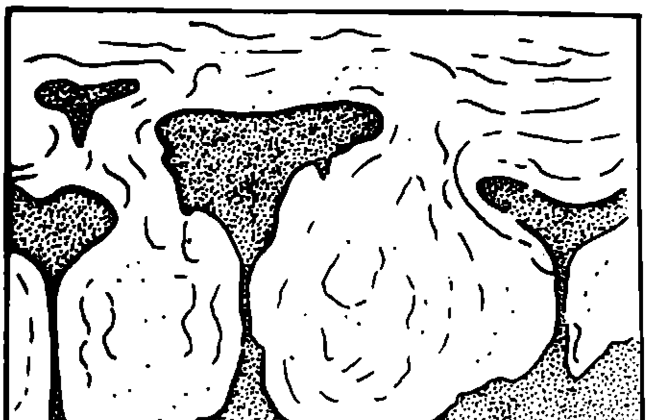

Fig. 14 Upward injected drop structures in Pleistocene sands; organic silt – *dotted*, sand – *white*. Modified after Lea (1990)

towards the center, induces there more rapid sinking. The resulting drops commonly lose their connection with their parent layer or the top surface of (ba) system and acquire the shape of isolated rounded bodies.

It is realized that drop structures may develop independently of sorted polygons. In heterogeneous (ba) systems, such structures may form in places where the heavier upper layer is thick enough to promote more intensive sinking.

Fossil drop structures have been described from various

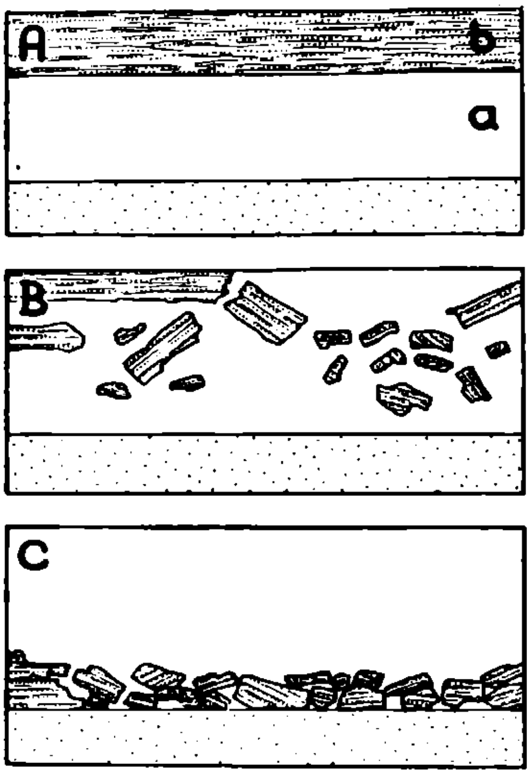

Fig. 15 Deformations in (ba) system in which (*b*) shows brittle behavior and (*a*) is prone to liquefaction. After Anketell *et al.* (1970)

environments, including Pleistocene glacifluvial sediments (e.g., Steusloff, 1950; Maas & Müller, 1954; Steward, 1963; Selley, 1964; Pękala, 1964; Kłysz, 1975). Such structures may also reveal the previously mentioned bilobate contours in vertical cross-sections which tend to form when a drop of liquid penetrates into another liquid with which it is immiscible. Experimetal analogues of drop structures have been repeatedly investigated by several authors from physical point of view and in order to explain the injection phenomena in sedimentary and and igneous rocks.

Deformations in non mobile (ba) systems showing partly brittle behaviour

In sedimentary (ba) systems in which the upper member shows brittle behaviour, the disturbance consequent upon liquefaction of the lower member may result in collapse and fragmentation of the upper member (Anketell *et al.*, 1970). Depending on mechanical properties of the upper member, its fragments may be angular, subangular or crumpled. In many instances, the collapsing platy fragments become rotated by 90°. Such rotated fragments are known to occur among natural structures. On "freezing" of the liquefied memember, the fragments become suspended in the fixed mass and the resulting structure resembles that of intraformational breccias (Fig. 15B).

If the viscosity of the lower member is sufficiently reduced or its state of liquefaction continues for an appreciable time interval, the fragments of the upper member may accumulate at the base of the system to form a rubble bed as shown in Fig. 15C.

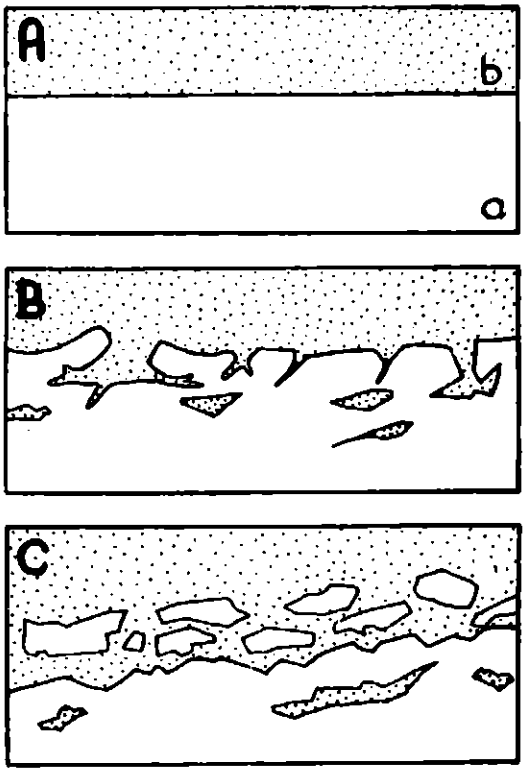

Fig. 16 Deformations in (ba) system in which (*b*) behaves as quick sand and (*a*) approaches plastic limit. After Anketell *et al.* (1970)

In anticipation of what will be discussed in the last but one chapter of this paper, reference is made to yet another experiment described by Anketell *et al.*, (1970). In this experiment, the upper member of an unstable (ba) system consisted of a heavy sandy suspension, and the lower one of settled kaolin clay approaching its plastic limit. Under the weight of the accumulating upper member, the lower low density clay layer underwent fragmentation. The opening fractures were filled with, and enlarged by descending sandy suspension (Fig. 16). Some of the detached clay fragments were lifted and incorporated into the structureless mass of sandy material. Natural analogues of the above mentioned "in situ fragmentation" were described by Covan and James (1992).

Convective deformations in non mobile (ba) systems in which instability propagates laterally

The hitherto discussed deformations affected more or less simultaneously the whole interface separating layers differing in density and kinematic viscosity. In non-mobile (ba) systems, the deformations arising from instability may affect instantaneously only a part of the interface between the members and then spread progressively over the area initially unaffected. Such case was investigated by Anketell and Dżułyński (1968b) in an experiment in which a thin and soft layer of plaster-of-paris turbidite was deposited upon a clay layer to produce an unstable (ba) system. The system was then subject to a slight single tap on the shute through which the plaster-of-paris suspension was introduced into

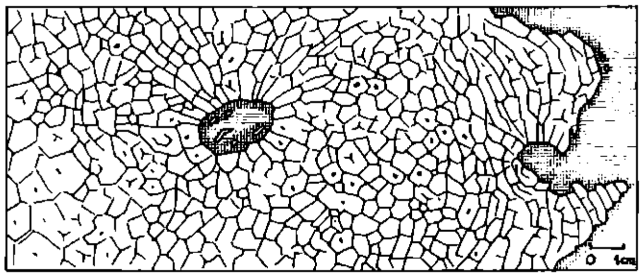

Fig. 17 Plan view of deformation pattern resulting partly from intermittent propagation of liquefaction front in non-mobile (ba) system in which $k_a < k_b$. Polygon covered area affected by instantaneous liquefaction of low density lower layer (*a*). Transverse pattern trending parallel to margin of intact upper layer (*dotted area*) produced by intermittent propagation of disturbance. After Anketell & Dżułyński (1968b)

the flume. In the proximal part of the flume, the disturbance was followed by instantaneous liquefaction of the clay and the appearance of clay polygons (Fig. 17). The distal part of the turbidite was at first unaffected and an irregular sharp boundary was formed between the disturbed and undisturbed turbidite. Then, however, the deformation was observed to spread slowly and intermittently toward the distal end of the tank, breaking the undeformed plaster-of-paris layer into elongated clay compartments parallel to the boundary between the deformed and intact area (Figs. 17 & 18). Because of its higher kinematic viscosity, the plaster-of-paris was squeezed into narrow pendent wedges between broader clay bodies. This process is shown schematically in Fig. 18A. Natural features showing similar morphologies are to be found among some periglacial structures. This question, however, requires further field investigations.

In similar experiments with more viscous clay ($k_a > k_b$), the configuration was reversed in that the underlying clay formed narrow injective ridges and the overlying plaster-of-paris was downbuckled to form canoe-shape bodies whose axes were oriented at right angle to the spread of liquefac-

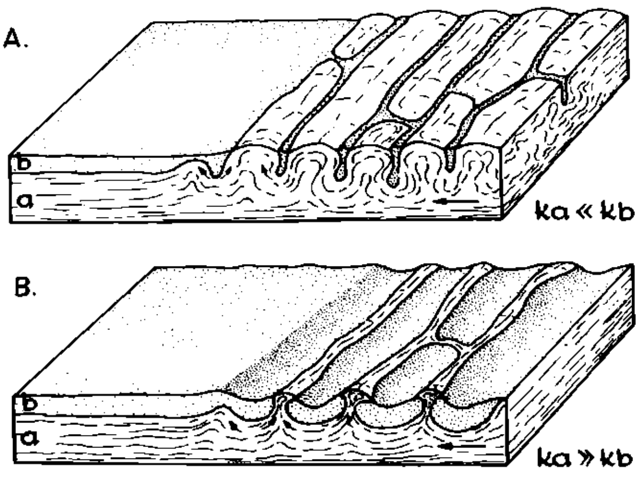

Fig. 18 Schematic presentation of deformation pattern in non mobile (ba) system in which instability propagates laterally. Direction of propagation shown by arrow. After Anketell *et al.* 1970

tion (Fig. 18B). Morphologically similar structures are known to occur in the form of semi-isolated bodies in shales and on bottom surfaces of natural turbidites. Whether or not, these natural structures were formed in the way indicated by the experiment, remains open to question. The structures of similar morphologies may also belong to the scheme of transformations evolving from polygonal compartments acted upon by a slight horizontal shear (Dżułyński & Simpson, 1966b; see later) and there is no simple set of criteria to distinguish one from another.

Other deformations resulting from instability in non mobile (ba) systems

Some of the structures previously discussed are recently explained in terms of fluidization. The role of fluidization in the formation of such structures must not be overlooked (see e.g., Lowe, 1975), notably, in view of recent experiments by Nichols *et al.* (1994). However, most of the deformations here under consideration fail to reveal evidence of transfer of sediment particles by fluid drag which is restricted to vertical conduits and results in break-through pipe-like features (e.g., Brencheley & Newall, 1977; Mills, 1983) as is the case with cylindric structures produced by springs pouring through stratified sands (Havley & Hart, 1934; Dionne, 1973). Moreover, in considering synsedimentary deformations it is not easy to differentiate between liquefaction and fluidization. Intruding sandy suspension behave as one homogeneous liquid and it is not water that drags the sand grains.

The previously discussed structures resulted chiefly in simple bipartite (ba) systems with more or less homogeneous members, deforming chiefly by plastic flow. The deformations in multilayered unstable systems and/or reflecting a plurality of deformational events, are more complicated, notably when the layers involved differ in their mechanical properties, texture, structure and water saturation (see; e.g., Brenchley & Newall, 1977).

The multiple (ba) systems may occur: (1) in a single bed produced by intermittent but essentially one depositional episode and, (2) within in a sequence of beds comprising several discrete sedimentary episodes which are then amalgamated into one new bed.

With respect to the first case, reference may be made to the well known "dish-and-pillar" structure (e.g., Wenthworth, 1967; Staufer, 1967; Lowe & Lo Piccolo, 1974). The structure in question consists of concave-upward lamination disrupted by vertical pillars attributed to explosive water escape and/or eruption of fluidized sediment. The dish-and-pillar structure may be due to overloading or earthquake motivated collapse of stratified beds containing sandy intercalations which are prone to liquefaction and transformed into quick-sand. The concave-up shape of disrupted fragments of layers results from violent discharge of the quicksand and/or foundering of upper members into more viscous layers below. The dish-and-pillar structure, as seen in plan view, shows roughly polygonal pattern and in this respect appears to be related to deformations arising from type A of incipient deformations. It is, however, the presence of relatively thick sandy intercalations, prone to liquefaction and confined between less permible or impermible clay-rich laminae, that provides conditions to vertical quicksand injections.

More complicated deformations occur when a sequence of alternating unconsolidated beds of differing characteristics is subject to abrupt gravitational collapse, producing thus a "new bed" (see Selley *et al.*, 1963). Less coherent and lighter layers are thereby squeezed upwards between descending heavier units. The disrupted fragments of such units, foundering into soft and/or liquefied substratum acquire turned up or even infolded margins. Such structures known as "pseudonodules" (Macar & Antum, 1949-1950) or "ball-and-pillow" structures (Smith, 1916) were produced experimentally by Kuenen (1958). Where the pseudo-nodules are uniformly concave upward and show little or no rotation, essential descent in place is indicated.

In many instances, the deformation in such beds may proceed consecutively and it is possible to discern earlier deformational stages. In other instances, the deformations pertaining to different stages are projected onto each other in such a manner that the recognition of the preceding stages is hardly possible. Particularly complicated structures arise during deformations involving layers differing drastically in competence and when horizontal shear is included into the deformation. Irregularities also arise from the fact that metastable aggregates of grains may contain portions which resist liquefaction.

It is not necessary to dwell further on the above mentioned structures as they are illustrated and described in many publications (e.g., Selley *et al.*, 1963; Rizzini & De Rosa, 1969; Dionne, 1971; Brencheley & Newal, 1977; Brodzikowski & Haluszczak, 1987; Brodzikowski & Van Loon, 1985; Davenport & Ringrose, 1987).

CONVECTIVE DEFORMATIONS IN MOBILE (BA) SYSTEMS

Convective deformations in mobile (ba) systems may be considered in terms of two component vectors V_p and V_h, representing the movement perpendicular and horizontal, i.e., parallel to the interface between (b) and (a). Two situations are feasible here: (1) the perpendicular and horizontal vectors are approximately of the same order and, (2) the horizontal vector is greater then the vertical one ($V_p < V_h$). The resulting features will be referred to as "scales" and "longitudinal ridges" (abbreviated to l-ridges) respectively. Although these two types grade one into the other, they will be dealt with under separate headings, taking into account the ratio of kinematic viscosities k_a/k_b and assuming that the deformation involves simultaneously the whole interface between (a) and (b).

Scales on bottom surfaces of beds

Variously shaped and arranged moulds of scaly ridges were described by Craig and Walton (1962) from the Silurian flysch of Scotland as erosional features affected by

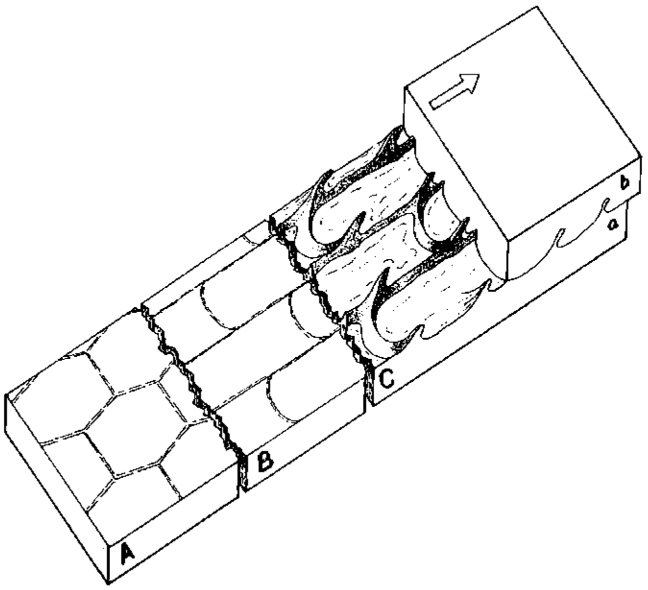

Fig. 19 Transformation of hexahedral pending lobes ($k_a > k_b$) into scales arranged in rows. After Anketell *et al.* (1970). Comp. Pl. 28B, C. Explanations in text

loading. The above mentioned authors emphasized the intimate association of these structures with the moulds of longitudinal ridges.

Considering the origin of the structures in question in terms of two component vectors, one can envisage that such structures originate by means of two processes operating separately in two successive stages. In the first stage there is a vertical movement which ends with the appearance of pending or rising polyhedral columns (as was the case with non mobile (ba) systems). In the second stage, the polyhedrons produced are acted upon by a slight horizontal shear. The simplified vision, where the two stages are considered separately, is admissible for clarity of presentation. It should be borne in mind, however, that these two processes operate simultaneously. In sedimentary (ba) systems, the lateral shear is effected by current drag or gravity creep on gently sloping surfaces. Let us consider the scaly deformations produced experimentally by sandy suspensions flowing upon a substratum made of watery mixtures of plaster-of-paris and clay ($k_a > k_b$). Suppose that the polygonal compartments formed in the first stage of deformation and bordered by squeezed up ridges of the substratum (type A), are oriented with respect to the direction of lateral shear as shown in Figure 19. The diagram is somewhat idealized in that the compartments are hexagonal in plan view. If acted upon by horizontal shear, the compartments are attenuated parallel to the direction of shear. The clay ridges which are oblique to this direction are straightened. Those which are transverse, are bulged outward, with their convexities up-shear. This behaviour occurs because of the clay wisps which project from triple points into the overlying slowly moving and settling suspension. The wisps, whose tops extend into zones of higher shear velocity, are dragged forwards and draw with them those parts of transverse ridges which are close to the wisps. The least affected by drag are the central portions of transverse ridges. Being the lowest, these portions are not dragged as much as the other parts of the ridges. Consequently, the transverse ridges are curved and acquire U-shaped forms oriented with their convexities upcurrent.

The diagram in Figure 19, showing the positive aspect of scaly structures arranged in rows, was based on experiments in which a fine-grained sandy t-current was allowed to flow along bottom sediment made-up of clay and plaster-of-paris mixture. The resulting relief was exposed after hardening of the bottom sediment and washing away the sandy turbidite. This relief is shown in Plate 26C. The surface depicted on the right side of the photograph shows I-ridges with wispy protuberances. These ridges correspond to the stretched out flow-oblique sides of incipient polygons. The protuberances extending from original triple points are inclined in the direction of shear. The specimen depicted on the left side of the photograph shows the same surface after the uppermost thin layer of congealed substratum has been brushed away and the remaining sand grains removed from the furrows. The brushing exposed arcuate ridges which correspond to curved flow-transverse sides of polygonal compartments. Plate 26B shows moulds of scales, arranged in rows, as they are seen on bottom surfaces of artificial plaster-of-paris turbidites and Plate 26A illustrates the same type of structures on the sole of natural turbidite.

If the shear direction is as shown in Figure 20, the hexahedral compartments acted upon by current drag are arranged into an "en echelon" pattern. This time, however, the central portions of oblique sides of the compartments are least affected by shear, in contrast to flow-parallel sides. Plate 27A and B show moulds of the structures in question as they are seen on the bottom surfaces of artificial and natural turbidites. The scales arranged in en echelon pattern were described by Ten Haaf (1959) as "squamiform load casts".

From the foregoing considerations it appears that polygonal compartments developed in mobile sedimentary (ba) systems are drawn into scaly pattern as is the case in experiments on heated fluids in the presence of a slight horizontal shear (Fig. 5).

Scaly structures on bottom surfaces of natural turbidites are reliable directional criteria facing with their convexities up-current. In vertical cross-sections, they are bulbous or flat-bottomed, depending upon the absence or presence of a resistant layer below the active (ba) system.

The two types of scaly patterns indicated above, correspond to two arbitrarily selected arrangements of incipient polyhedrons with respect to the direction of horizontal shear. The arrangement of incipient polygonal compartments, however, is variable. Consequently, the resulting structures are less regular and there are various combinations of scaly patterns. An example of such combination is shown in Plate 28 which illustrates the natural and experimental scales. In general, the spatial distribution and arrangement of polygons and scales may be determined by bottom irregularities and changes in the character of flow.

Scaly structures may be transformed into transverse structures when compressed in front of an obstacle. For $k_a > k_b$, the resulting structures take the form of convex-down

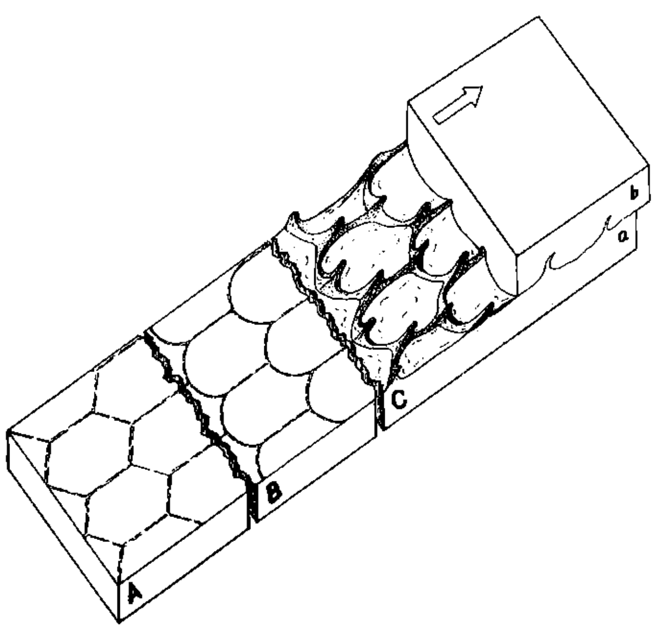

Fig. 20 Transformations of hexahedral pending lobes (ka>kb) into scales arranged in en echelon pattern. After Anketell *et al.* (1970). Compare Pl. 29. Explanations in text

square bodies whose longer axes are oriented at right angle to the direction of shear (Dżułyński & Simpson, 1966b; see Pls. 29, 30A 31, 32 in this paper). Depending upon the primary arrangement of scales, such bodies are distributed alternately or set in parallel rows (Pl. 31, Fig. 21).

The transverse scaly structures, here under consideration, compare to the scheme of convective deformations known from experiments on heated fluids (Chandra, 1938). The natural structures shown in Plates 29 A and 30A belong presumably to the same scheme of transformation.

Scaly structures bear morphological similarity to flute moulds and are commonly identified as scour structures. Here again, the distiction between scour and deformation is hazy. As noted, the erosion and deformation are complementary processes depending upon mechanical properties of the bottom. If the bottom is ductile rather than loose, the resulting structure is dominantly of deformational origin, although small amount of erosion cannot be excluded. Becuse of the role played by sediment intrusion, some of the scaly structures have been recently indicated as "flute-like marks" (Keighley & Pickerill, 1994). The present writer would not wish to use this term, for only further confusion would result.

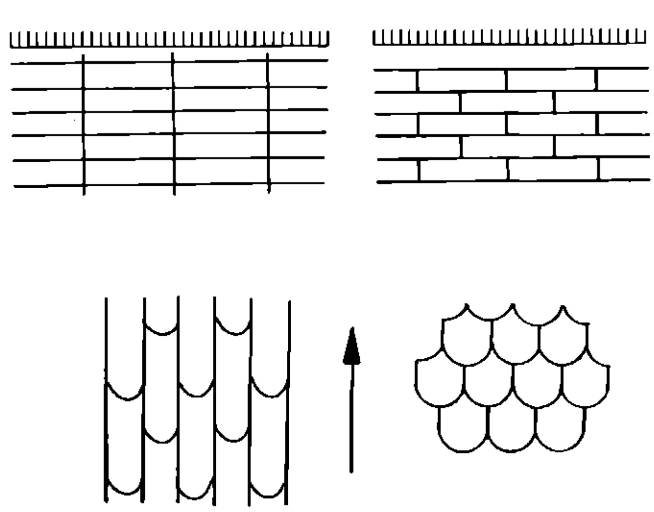

Fig. 21 Idealized diagram showing transverse deformation patterns resulting from pressing of scaly structures against obstacle. Based upon experiments. Comp. Pls. 31, 32. Explanations in text

Scaly structures on top surfaces of (ba) systems

Scaly structures which develop from incipient deformations of type B, face with their convexities downcurrent (Fig. 22). Such orientation is explained by the fact that, contrary to the previously discussed scales, it is the upper member material that is pushed aside by rising lobes of the underlying less viscous and lighter material to produce downhanging polygonal wedges (not ridges!). At triple points, this material forms pendents whose ends extend to zones of highly reduced or no shear. Consequently, the more shallow wedges between the triple points are pushed forwards in preference to the pendents and their central portions run ahead of the remaining portions (Fig. 22). Natural equivalents of the above mentioned structures, known as "stone garlands" (e.g., Gripp, 1927), occur on gently sloping surfaces in "periglacial" regions (Fig. 23).

Parenthetically, it may be noted that analogous pattern of pending wedges was obtained in experiments in which plaster-of-paris t-currents were flowing over slightly congealed gelatine substratum (Dzulynski & Walton, 1963, their pl. XX B).

L-ridges on bottom surfaces of beds

In mobile (ba) systems in which the lateral component of motion predominates, the flowing suspension tend to organize itself into flow-parallel double rolls showing helicoidal cross-currents with opposite sense of rotation (Fig. 24).

The helicoidal or spiral flow is of diverse origin (see

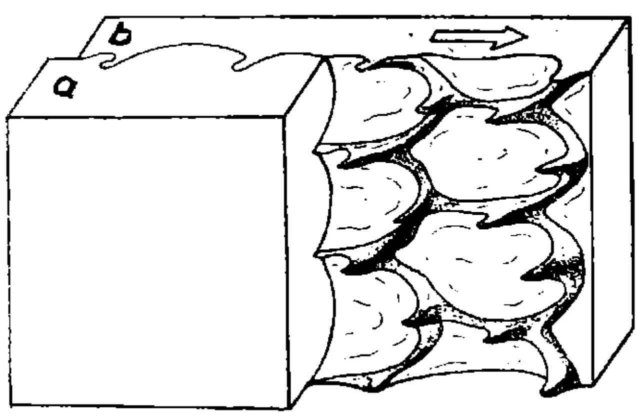

Fig. 22 Transformations of hexahedral up-rising domes (ka<kb) into scales arranged in an en-chelon pattern. After Anketell *et al.* (1970). Explanations in text

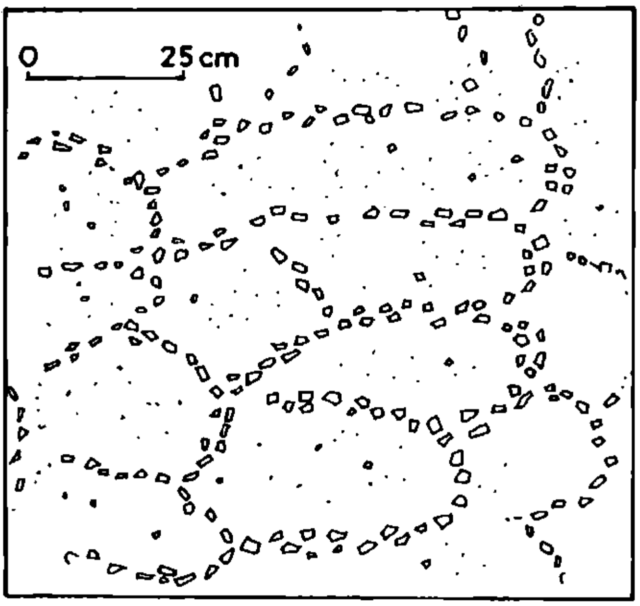

Fig. 23 Sketch of stone garlands. Plan view. Modified after Sharp (1942)

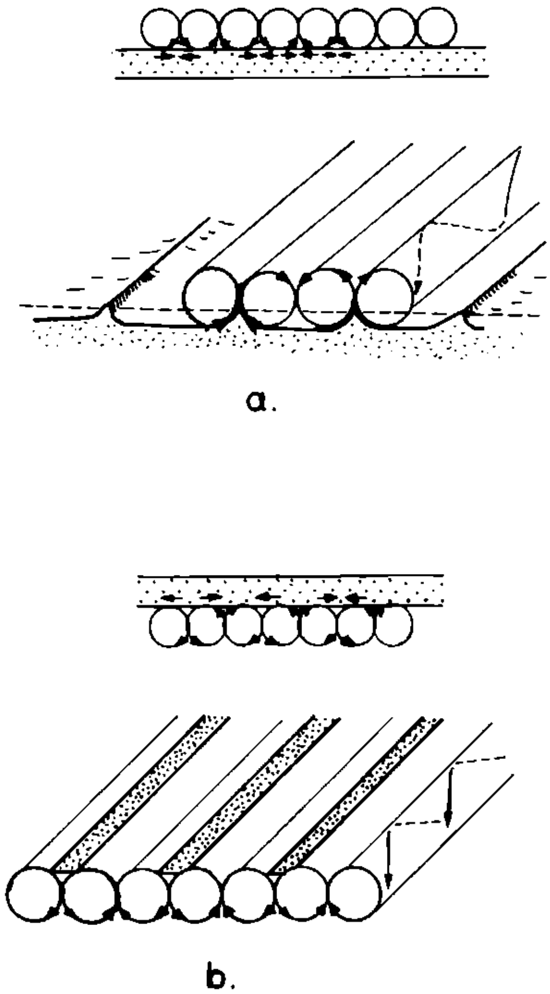

Fig. 24 Schematic presentation of development of experimentally produced current ridges (a) and sorted stripes (b). After Dżułyński & Walton (1963), Dżułyński (1966). Explanations in text

e.g., Vanoni, 1946; Matthes, 1947; Kolar, 1956; Karcz, 1970; Allen, 1971a, 1984; Wilson, 1972; Best, 1992). It may be stimulated by bottom irregularities, curvature of the flow, sudden increase in current velocity, generated as the boundary effect and, caused by density gradient.

In flowing suspensions, the preservation of spiral flow for long time intervals requires a continuous expenditure of energy. Vanoni (1946) attributed this energy to the presence of suspended load and the resulting secondary circulation (spiral cross-currents). This idea was adopted by Dzulynski and Walton (1963) and Dżułyński (1966) to explain the origin of ridges preserved as moulds on the base of turbidites, in terms of reversed density gradient. Such gradient may originate when the flowing suspension sinks into a soft substratum. It also may be generated in the boundary layer close to the bottom where the current velocity is reduced and coarser grains, which higher up remain in suspension, settle onto the floor. In such situation, there appears a layer of lighter and more dilute suspension close to the bottom. Consequently, the sediment-laden suspension tends to sink down, and this, combined with forward flow, sets up the spiral circulation. With settling suspensions, the instability in density stratification is expected to occur repeatedly in the immediate neighborhood of the bottom.

If the kinematic viscosity of the suspension is less than that of the cohesive ductile substratum, the surface of the substratum is warped into narrow flow-parallel ridges separated by broader troughs (Fig. 24 a). Such a structure was first obtained experimentally by H. Casey (1935). Whatever the origin of spiral motion, the flow pattern in longitudinal stringers is characteristic of steady and uniform flows.

Moulds of ridges and intervening furrows on bottom surfaces of sandstones (Pl. 32) have long been recognized as the result of current action. They were first attributed to helicoidal flow in double rolls by Dzulynski and Walton (1963) and Dżułyński (1966); a concept developed and expanded by Allen (1971). The structure is generally regarded as purely erosional and/or depositional feature (e.g., McBride, 1962; Karcz & Kersey, 1980; Flood, 1981, 1983: Allen, 1984). As noted, such opinion is justified where a granular sediment is acted upon by helicoidal flow and the grains are capable of independent motion. The cross-currents produced, redistribute the sand particles to form flow-parallel ribbons (for details and references see e.g., Karcz, 1970; Grass, 1971; Grass *et al.*, 1991). However, when the helicoidal flow is superimposed upon a ductile substratum, the main ridge-forming process is the deformation in which individual particles have limited or no freedom of movement. This is observed where the substratum, in which the l-ridges develop, is laminated. In such situation, the laminae are warped into sharp-crested anticlines and rounded synclines, but retain their identity (see later).

The l-ridges are striking because of their marked parallelism, constant size and wave length (Pl. 32). Closely spaced sets of ridges are indicative of relatively rapid flow and relatively low density of the settling suspension. The ridges may bifurcate up-, and down-current. The upcurrent

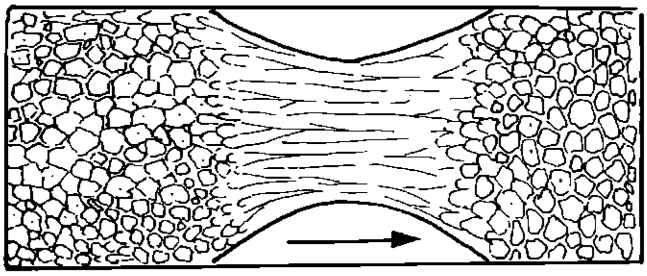

Fig. 25 Polygonal ridges and scales passing into l-ridges in narrow passage of flume. After Dżułyński & Simpson (1966b)

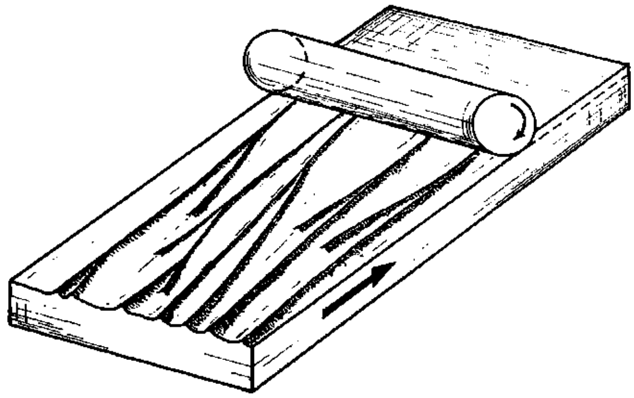

Fig. 27 Diagram showing formation of dendritic ridges by rolling cylinder

bifurcation occurs when the streamlines are pressed together by lateral constraints (Fig. 25). In such situation, some of the double-rolls are thrust upward out of place. On the other hand, where the suspension spreads laterally (e.g. in places where the passage widens), the streamlines tend to separate and the overlying double rolls are pressed downward. If this be the case, the resulting ridges bifurcate downcurrent (Dzulynski & Walton, 1963; Folk, 1977).

Moulds of l-ridges showing distinct upcurrent bifurcations are known as dendritic ridges (Pl. 33) or "syndromous load-casts" (Ten Haaf, 1959). They were produced experimentally, by the slow advance of the crenulated front of water upon an exposed, soft and superficially aerated surface of clay (for details see, Dzulynski & Walton 1963, 1965). The clay was pushed aside between the double rolls forming ridges along the lines of ascending cross-currents and the advancing front became scalloped reflecting the motion towards a medium of higher kinematic viscosity (air !). The scallops were split into still smaller lobes (analogous to fractal partition) between which minor tributary ridges were formed (Fig. 26). Such ridges joined the main ridges obliquely, giving rise to upcurrent bifurcation. These upcurrent bifurcations, however, were not produced by upward thrusting of double rols, as was the case of the previously discussed l-ridges. This may be an argument in favour of polygenetic nature of bifurcation.

The results of the above mentioned experiment cannot be directly applied to explain the origin of natural dendritic ridges. Such ridges do not originate in the front of an advancing t-current, as previously suggested (Dzulynski & Walton, 1963), but away from it (Dżułyński, 1965a; Allen, 1984). They belong presumably to the endmost generations of sole markings. A vague suggestion can be made that the already condensed "traction carpet" is pushed forwards by the current drag, and produces dendritic ridges in much the same way as the impact induced ridges are generated (see below). The "fractal partition" of secondary lobes as described in this section may, however, throw light on the development of frondescent marks (see later).

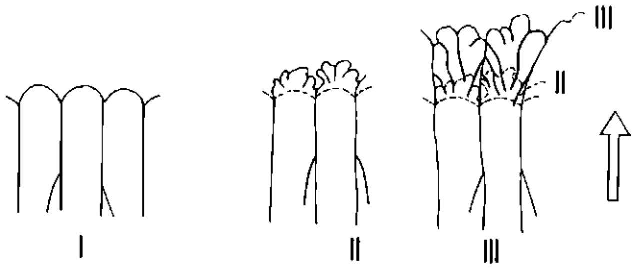

Fig. 26 Schematic diagram showing showing formations of dendritic ridges by secondary crenulation of lobes of advancing water front. Explanation in text Modified after Dżułyński & Walton (1965)

Dendritic ridges, in system with positive density gradient may also be produced in purely mechanical way by pressing the exposed surface of a settled clay by rolling cylinder (Anketell *et al.*, 1970; Folk, 1977). The rolling of the cylinder inflicts an unidirectional horizontal shear and this, combined with loading, results in the appearance of dendritic ridges on the top surface of the lower clay layer (Fig. 27). Natural equivalents of such impact induced ridges occur bottom surfaces of some prod marks (Fig. 28). The structure depicted in Pl. 11C, resulted from instantaneous loading and horizontal shear produced by the the fish vertebra striking the bottom mud (Dżułyński *et al.*, 1972).

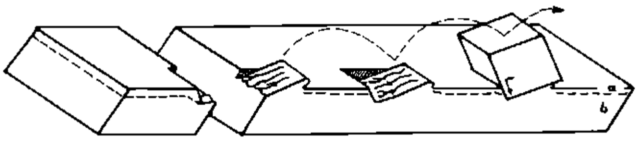

Fig. 28 Diagram showing impact induced dendritic ridges. Comp. Pl. 11C. After Dżułyński *et al.* 1972

Interrelation between polygonal, scaly and longitudinal ridges

Moulds of polygonal, scaly, and longitudinal ridges on bottom surfaces of sandstones are intimately interconnected and pass one into another (Craig & Walton, 1962). Plate 34 shows the undersurface of natural and experimental turbidite in which the moulds of troughs reveal traces of arcuate sides of scaly structures.

With current-induced ridges, any change in their pattern reflects the change in the pattern of streamlines. In this respect, such ridges may serve as convenient means of visualizing the pattern of streamlines. In general, the pattern and distribution of ridges depend upon: (1) the pattern of streamlines inherent to the flow and, (2) the shape and distribution of bottom irregularities.

As indicated previously, the transformation of l-ridges into polygonal, pillow-like and transverse structures may occur in front of obstacles (Pl. 30), It also may occur in on the downcurrent sides of bottom undulations or ripples (Fig. 29), where the smooth steady flow breaks into turbulent cell-vortices (Dżułyński & Simpson, 1966b) and in areas of interference between deflected and undeflected flow (Pl. 35A, B). Admittedly, however, the flow-transverse zones of l-ridges, alternating with zones showing polygonal structures may also reflect transverse oscilations or pulsations inherent in natural flows (e.g., Matthes, 1947; Folk, 1976; Karcz & Kersey, 1980).

The control of bottom irregularities upon the pattern of ridges and scales is also shown by other experiments. For instance, the l-ridges exhibit an approximately radial convergence toward the center of roughly circular bottom depression (Fig. 30). With flow-parallel elongated bottom depressions, the ridges and scales reveal downcurrent convergence towards the axis of the depression (Fig. 31). This pattern is analogous to the "fleur-de-lys" pattern of scales described by Craig and Walton (1962, fig. 2c). Finally, the l-ridges also may occur in flow parallel zones alternating

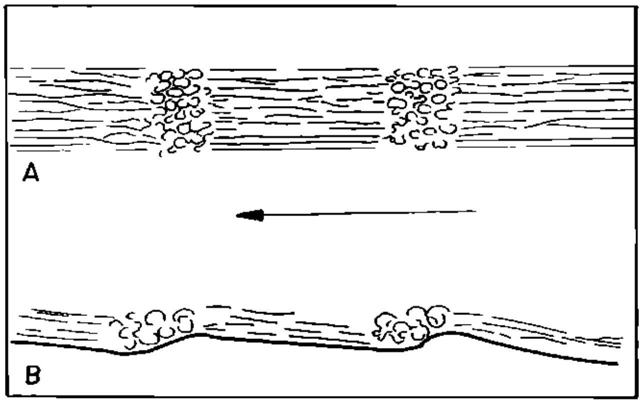

Fig. 29 Sketch showing formation of l-ridges on upcurrent side of ripples and polygonal structures on their lee side

with zones showing polygonal ridges (Pl. 35). Such pattern corresponds to alternating zones of relatively rapid and highly reduced flow velocities at the bottom.

Postdepositional l-ridges and scales on load-deformed bases of ripples

Special type of short l-ridges occur on load-deformed bases of ripples (Pl. 35C). It is to be recalled that ripples deposited on soft substratum may be subject to deformation resulting in downbuckling of their undersurfaces, whereby the maximum sagging is underneath the crests of ripples (Dżułyński & Ślączka, 1959; Dżułyński & Kotlarczyk, 1962; compare also Craig & Walton, 1962, pl. IV, and Dżułyński & Walton, 1965, fig. 104 and 105). The moulds of the ridges discussed, occur exclusively on the upcurrent

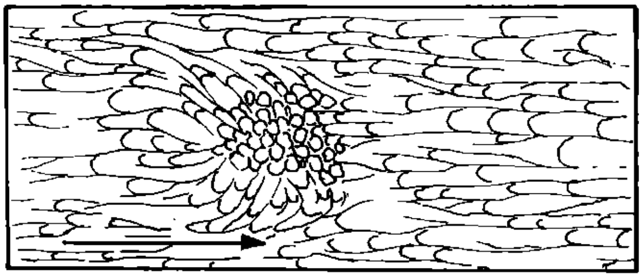

Fig. 30 Radial convergence of l-ridges and scales around circular bottom depression. Modified after Dżułyński & Simpson (1966b)

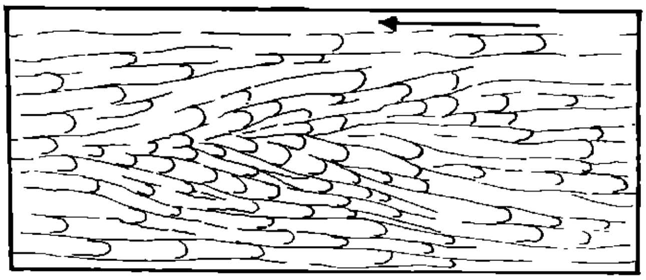

Fig. 31 Fleur-de lys pattern of l-ridges and scales developed in flow-parallel bottom depresion. Modified after Dżułyński & Simpson (1966b)

side of load-deformed base of ripples, and in this respect, are reliable directional indicators. They terminate abruptly and/or break into small knobby, scaly or reticulated pattern of ridges along the line of maximum downbuckling.

The ridges in question, appear to have been made simultaneously with the deformation of ripples through partial liquefaction of the lowermost foreset laminae. The liquefied material was pressed into the underlying muddy substratum and slightly pushed along the sand/mud interface towards the region of maximum sagging. This brought about the appearance of ridges or ridge-and-furrow structures on the upcurrent side of the down-buckled base of ripples. On the lee side of deformed ripple bases, the formation of ridge-and-furrow structure was impeded by the foreset laminae which intersect the bent sole of ripples at this side and obstruct the flow of liquefied material.

Argument in favour of the above explanation was provided by an unpublished experiment made in 1961 by E. K. Walton and the present writer. The experiment was carried on in a water tank with rubber bottom in which a layer of clay was overlain by a layer of sand mixed with plaster-of-paris. The rubber bottom was bent down by means of a handle attached in the middle of the rubber bottom to produce a roughly U-shaped deformation. The sagging was accompanied by a shear and mutual interpenetration between sand and clay, as well as the formation of a dilatant crack in the middle the deformed sand layer. The exposed base of this layer revealed "ridge-and-furrow" structure fringing on both sides the dilatant crack. Inasmuch as the clay had lesser kinematic viscosity, it penetrated into the overlying sand to form round-crested ridges separated by narrow sharp-crested pendents (an arrangement reversed to that observed on the base of deformed natural ripples). The symmetrical distribution of structures with respect to the dilatant fissures was due to structureless and homogeneous character of the sand layer.

Intrastratal l-ridges

The simplest form of intrastratal l-ridges are linear sharp-crested ridges (Pls. 36 & 37A) showing "cusp and point-up" pattern in transverse cross-sections. The size range of these structures is measured in milimetres and centimetres . In rare instances the amplitude of ridges attains several tens of centimetres. Care should be taken not to confuse the ridges under discussion with linear wave ripples.

Of particular interest are ridges which develop in laminated sediments (Pl. 37B). The ridges in successive laminae follow the contours of ridges in the preceding laminae. This indicates that the ridges are deformation structures and result from the tendency of flowing suspensions to move in the form of double-rolls with opposite sense of rotation. Such rolls create alternating zones of upward suction and downward pressure which deform the hydroplastic substratum into flow parallel "synclines" and "anticlines". The above described ridges may be regarded as current structures produced by flow of suspension, accompanied by intermittent rhythmic accretion of new laminae upon the earlier deposited laminae. As long as the conditions promoting the development of ridges remain unchanged, the axes of "anticlines" persist in the same vertical planes. This means that the flow-parallel ridges, once formed, influence the location of subsequent ridges, as is the case with sand streaks produced on friable substratum (comp. Best, 1992). It may be noted that structures identical with the previously mentioned deformational ridges were observed in sloping, laminated cave muds and rightly interpreted as non erosional features by Bull (1978).

As far as the formative process is concerned (spiral motion), the deformation ridges discussed are analogues of erosion-induced "accumulation" ridges on incohesive granular substratum or on hard substratum covered with a thin layer of sand (Karcz & Kersey, 1980).

To the same category of structures belong ridged convolutions (Pl. 38A) which tend to form in mobile (ba) systems, whose lower member is laminated and the upper one rippled, massive or laminated, both members being structurally homogeneous in statistical sense (Anketell & Dżułyński, 1968a). The downsinking of the upper member, combined with unidirectional movement along the interface between (a) and (b), gives rise to longitudinal (flow-parallel) undulations which, in transverse cross-sections, reveal "cusp and point-up" pattern. In flow parallel cross-sections, the interface between the members is flat or gently undulated, occasionally with vertical knobs.

The convolution mentioned is synchroneous with deposition of the host bed as a whole, but postdates the deposition of members involved in deformation. The horizontal shear along interfaces separating the members is, presumably, due to current drag.

Longitudinal undulations in massive sandstones

While the concept of convective motion may still be invoked to explain the above described ridged convolutions, some troubling questions arise with respect to flow parallel undulations illustrated in Plate 38B. The undulations involve thick massive sandstone beds deposited in an erosional channel. The top surface of the host bed is sharply delineated with short upward injected sandstone dykes, indicative of partial liquefaction of the upper member. It is not clear whether the sandstone bed represents separate depositional episodes welded together into one "new bed", or reflects changes or lulls within one depositional act. The flow transverse wrinkles on interfaces separating the undulated layers, reorientation of grains and apparent schistosity of fine-grained intercalations between the members, inconsistent with the trend of tectonic features, bear evidence of limited flow parallel shift of the upper member.

Inasmuch as the sandstone in question fills a relatively narrow erosional channel and the undulations are neither regular nor areally widespread, there is room for argument that such undulations are longitudinal folds resulting from lateral pressure to which soft masses of sediments are always exposed when sliding into a narrow passage (see later). The previously discussed minor experimental l-ridges, however, also originate when the suspension in excess, enters a narrow passage.

"Besko type" convolutions

We conclude our survey of flow parallel deformation ridges with brief comments on some unusual type of structure which, for the lack of a better term, is here indicated as "Besko type" convolutions, after the name of locality in the Polish Carpathians, where this structure was first described (Dżułyński & Ślączka, 1958).

The structure in question develops from flat-lying thick laminae in the lowermost part of a turbidite bed. Higher-up, the the laminae become thiner and are warped into regularly spaced longitudinal and transverse round-crested ridges (Pl. 38A).

From intersections of these ridges, vertical conical knobs arise. Still higher up, the relatively large ridges become vaguely delineated, but the conical knobs are still recognizable. The growth of these knobs affects the subsequent laminae which are rised without interruption by the growing cones. Towards the top of the bed, the internal layers become thiner and are thrown into more densely spaced flow-parallel undulations. In places, such undulations are deformed plastically by the growing knobs, which also here do not break the laminae but arch them upwards. Close to the top of the bed, the deformations vanish and the parting surfaces become flat with only localized disturbances.

Experiments with plaster-of-paris suspensions coloured by coal dust, and flowing over a soft substrate (mixture of clay and plaster-of-paris) provided an insight into the development of the pattern discussed. The experiments mentioned were not entirely realistic in that the unstable system (ba) was produced by two separate t-currents following shortly one upon another. The first current resulted in the formation of the soft white substrate which was then covered with another artificial and dark turbidite. The resulting structure, as seen in plan view, is shown in Plate 39B. The left side of the structure depicted represents a horizontal section close to the interface between the two members of the system, exposed after the removal of much of the second dark turbidite. The exposed surface reveals the pattern of longitudinal and transverse ridges developed on the interface separating the two successive depositional episodes. From the intersection points of these ridges arise conical protuberances whose horizontal sections are seen on the right side of the specimen. The right side represents a higher section of the second turbidite exposed after the removal of its uppermost layer.

The pattern of longitudinal and transverse ridges is similar to that revealed by parting surfaces of the natural structure. The vertical injections which rise from intersection points of the ridges correspond to the previously mentioned vertical knobs. Significantly, both the natural and experimental knobs are vertical and do not show downcurrent deflection. In experiments, the rise of knobs was initiated by a slight tapping of the flume, shortly after deposition of the second turbidite. Whether the tapping only stimulated the further growth of the already formed protuberances or acted as the main formative agent, remains open to question. Also open to question is the origin of transverse ridges, which might have resulted from shock-induced wave or may represent undulation effected by the current action. The vertical orientation of conical knobs, however, is not in accord with the action of effective current shear.

The Besko type convolution is not commonly encountered. Somewhat similar deformations, however, have been described by Ian Rolfe (1963) from Old Red Sandstones and attributed to air-heave and by Nougier (1964) from periglacial sorted stripes.

Linear deformations on top surfaces of (ba) systems

As indicated, with lesser kinematic viscosity of the lower member, the more viscous material of the upper member is pushed aside to form pending wedges. Linear features of this type, which form on top surfaces of mobile (ba) systems, have been designated "sorted stripes" (Washburn, 1956). Such features (Pl. 40A) tend to develop on sloped surfaces in cold climatic regions and their origin is disputed (for references and details see: Washburn, 1969). In experiments, the analogues of sorted stripes were produced in the following way (Dzulynski, 1963, 1965, Butrym *et al.*, 1964). Clay settled in a water-filled tray was covered with a thin layer of coloured plaster-of-paris deposited from suspensions or sieved through water unto the clay surface. After the water was removed, the tray was slightly tilted and held in an oblique position. On tilting the tray, the strips were formed spontaneously on the sloping surface. The rising clay diapirs were seen to transform into elongated tongues, whereby the heavier and coarser material of the upper member was pushed aside to produce pending longitudinal wedges (Pl. 40B). The above experiments offer an alternative explanation for the origin of sorted stripes. It is realized, however, that these features may originate in more than one way.

FRONDESCENT MARKS

Specific structures resembling foliating leaves which occur on bottom surfaces of turbidites are known as "cabbage leaf" structures (Kuenen, 1957), "feather-like hieroglyphs" (Książkiewicz, 1958), "deltoidal hieroglyphs" (Birkenmajer, 1958) or "frondescent marks" (Ten Haaf, 1959). The characteristic spreading "foliage" of these structures is believed to be directed downcurrent which, in most instances, is true (Pls. 41A, & 42A).

The origin of frondescent marks is disputed. Książkiewicz and ten Haaf considered them to be current marks. Birkenmajer regarded these structures as due to the flowage of sand into depressions of the bottom accompanied by the movement of the underlying lutite. Allen (1984), basing on experiments by Einsele *et al.* (1974) and morphological similarity between frondescent marks and the "plumose" or "herringbone" structures on joint surfaces of rocks, suggested that such markings were formed by tearing off strips of mud from the bottom by strong currents. Aalto (1995) interpreted frondescent markings to be moulds of scours produced by two oppositely rotating corkscrew eddies which were present in the liquid that eroded mud during flute formation. Dżułyński and Ślączka (1958), came to the conclusion that frondescent marks were produced by local liquefaction of sand, the conclusion confirmed by later experiments (see below).

Any plausible explanation of frondescent marks must take cognizance of the following observations: (1) these features belong to the last generation of sole markings, (2) although in most instances the structures in question foliate downcurrent, they also may diverge radially from point sources and, in some instances, upcurrent, (3) the body of frondescent marks is a structureless sandy mass devoid of any depositional structures such as grading, lamination or cross-lamination and, (4) they may occur, as entirely post-depositional features, on bedding surfaces of source beds of sandstone dykes.

An insight into the origin of frondescent marks was provided by experiments in water-filled glass tanks with a transparent gelatine substratum at the bottom (Dzulynski & Walton, 1963). The most instructive were experiments in which the relatively strong gelatine substratum was overlain by a warmed-up thin layer of soft gelatine, with or without water above. When the cooling produced a more resistant skin on the top of this layer, a plaster-of-paris suspension was poured over it. In places, where this skin was broken, the plaster-of-paris suspension was seen to flow in the form of flaring stringers within the soft gelatine. The penetrating suspension produced elongate tongue-like bodies on the undersurfaces of plaster-of-paris suspension, which were characterized by flaring and scalloped margins (Pls. 42B, 43B). Where the suspension penetrated the weak layer along a broad front, then the resulting structures had broad crenulated fronts. Where sinking at a point occured and the impetus imparted by the current flow maintained the motion of the suspension, the interpenetrating plaster-of-paris crept forwards in the form of fan-like flaring overlapping lobes, each with scalloped margins. The axes of scallops bifurcated downflow. In the absence of unidirectional forward motion, the plaster-of-paris suspension spread radially from points of injections.

The experimental structures are morphologically identical with natural frondescent marks. The observed formative processes, were not much different from those involved in the formation of longitudinal and dendritic ridges. The horizontal spread of downsinking suspension which brought about the appearance of artificial frondescent marks, took the form of familiar double-rolls or stringers. The essential factor, however, was differential sinking of the stringers after breaking through the surface of gelatine. As the stringers sank, they became separated from their neighbours. Having little confinement at their sides, the sunken and thus isolated stringers become unstable. They send out lateral scallops from which new tips may shoot out, in much the same way as was the case with dendritic ridges (Fig. 32). The axes of scallops were inclined in the direction of the spreading plaster-of-paris. The sinking stringers were followed by another portion of viscous suspension so that the contact with the parent suspension, and thus the turbidite, was not lost.

The structures discussed are synchroneous with deposition. They may, however, also represent post-depositional phenomena. The radiating markings, for instance, were obtained experimentally from point injection of already deposited but still very soft plaster-of-paris under the above indicated conditions.

The results of experiments fit well with field evidence and may explain the occasional occurrence of frondescent marks marks on the surfaces of source beds of sandstone veins. Such source beds are, otherwise, devoid of any primary structures (Dżułyński & Radomski, 1956). It is to be noted that, in many instances, the natural frondescent marks start from prod marks. Such a situation is analogous to point-injection of very soft plaster-of-paris by breaking a skin of the jelly layer by means of a needle prick.

The frondescent marks, as seen in vertical cross-sections, may be bulbous or flat-bottomed (Pls. 41 & 42). The former originate when the upper soft portion of the lower member of (ba) system is relatively thin. Frondescent marks may occasionally spread upcurrent. This occurs when the flowing suspension intrudes into the underlying mud at the downcurrent end of step-like bottom defects (Dżułyński & Walton, 1965, fig. 92).

Summing up, frondescent marks appear to be the result of the injection of liquefied sediment from the base of the settling turbidite into the underlying clay layer. In most instances, the spreading "foliage" is directed downcurrent. This occurs when the impetus of the current drives forwards the liquefied sediment. Such situation, however, is not always the case. Consequently, the frondescent marks alone are not reliable indicators of current direction.

Morphologically very similar to frondescent marks, but genetically different are "feather structures" (Woodworth, 1897; Książkiewicz, 1968) or "plumose markings" (Parker, 1942) on joint surfaces of sandstones and splitting surfaces of soft cohesive sediments (Cegla & Dżułyński, 1967). The plumose markings are occasionally observed on the surfaces of ripped up fragments of semi-consolidated sandstones enclosed in dense sediment flow deposits or subaqueous slumps. Such structures, indeed, record the tearing of strips of mud from a relatively stronger sediment, as envisaged by Einsele *et al.* (1974) and Allen (1984). In the present writer's opinion, however, these structure should not be identified with sole markings referred to as frondescent marks.

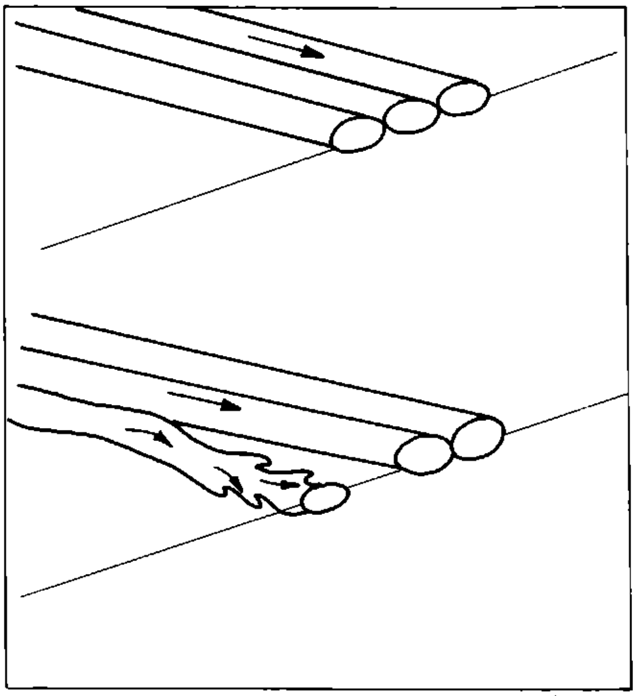

Fig. 32 Sketch showing formation of frondescent marks. After Dzulynski & Walton (1963)

CONVOLUTE AND CONTORTED LAMINATION

The term "convolute bedding" (Kuenen 1952) or "convolute lamination" (Ten Haaf, 1956, Sanders, 1960) is generally applied to more or less continuous intrastratal disturbances that die out both upward and downward within a given sedimentary bed, attaining its maximum in the middle of the bed. Strictly speaking, the term should be applied only to deformations which, in cross-sections, are characterized by regularly spaced, narrow sharp-crested "anticlines" separated by wider and rounded "synclines" ("cusp and point-up structure" of Selley *et al.*, 1963). In this respect, the structure is a reliable top and bottom criterion (Signiorini, 1936). The structures which do not show the cusp and point-up pattern should be referred to as contortions (e.g. Dott & Howard, 1962).

On the scale of the host bed, the convolute lamination is a synsedimentary deformation, accomplished before the final deposition of the host bed. As indicated by Einsele (1963), the convolute lamination is formed during the transformation from metastable to stable conditions and only in this short time interval, are the formative processes able to produce such deformations in fine-grained, laminated and water-saturated sediments. Although the force of gravity plays a dominant role in such processes, the formative mechanism may vary.

The origin of convolute lamination is still disputable. The once widely accepted explanation that this type of structure is produced by sediment creep on inclined bottom surfaces is now definitely abandoned. Consequently, the absence of lateral displacement of the deformed laminae has been included into the definition of convolute lamination. However, as shown previously (ridged convolution), the bedding parallel shear may be a factor controlling the pattern of convolute lamination. Some authors attribute the convolute lamination to expulsion of water, gas or quicksand (e.g. Steward, 1956; Jan Rolfe, 1963; Wunderlich, 1967; Boer de, 1979). Indeed, water-saturated granular aggregates, during deformation and compaction, give off much of the interstitial water, but such dewatering is the result and not the cause of convolute lamination. This is indicated by the observed continuity of the deformed laminae, which is a distinct feature of convolute lamination. The laminae are thinned, streched and, finally, may wedge out, but are not disrupted.

Such hydroplastic deformations do not form in response to water or air escape.

The prerequisite for the formation of convolutions is fine lamination and hydroplastic behaviour of the sediment involved. The lamination facilitates plastic deformations, because the adjacent laminae tend to slip by each other in opposite directions. The presence of lamination makes also the deformation visible. The finely laminated sediments yield easily to all kinds of deformations, including those inflicted by current-generated pressure and suction or load imposed by the settling suspension (e.g., Emery, 1950; Holland, 1959; Dott & Howard, 1962; Selley *et al.*, 1963; Sullwold, 1959; Davies, 1965; McKee & Goldberg, 1969; Allen, 1977).

As noted, when the laminated substratum becomes cohesive and ductile, the current shear, which otherwise would tend to redistribute the particles individually, generates undulation or streaked-out protuberances. The undulations once formed, may tend to become magnified through further depression of troughs by accumulating sediment and by rising of adjacent ridges (Ten Haaf, 1956). With ripple cross-lamination, such ridges represent the crests of ripples, and the above mentioned processes give rise to convolutions which are casually connected with intensification of the ripple crests by current drag. This type of deformation has been adequately dealt with by several authors (Kuenen, 1953; Sanders, 1960; Sutton & Lewis, 1966) and is mentioned here merely for completness' sake.

Reference has already been made to convolutions in structurally homogeneous dichotomous (ba) systems resulting from instability in density stratification and following convective patterns. Due to variations in texture, structure and composition, normal to many natural sediments, the structural homogeneity is not commonly realized in nature. Consequently, the resulting structures are less regular, though there is still a tendency toward the appearance of cusp and point-up pattern.

The cusp and point-up pattern, however, may also result from regular distribution of non penetrative features. This is, for instance, the case with "ripple-load convolution" (Dżułyński & Ślączka, 1965) which results from sinking of large semi-isolated ripples into a soft horizontally laminated substratum (see also Dżułyński & Walton, 1965, fig. 105).

Inasmuch as the sinking is greatest below the ripple crests, the plano-convex ripple bodies are transformed into planoconcave structures with flat top surfaces. The underlying, and originally horizontally laminated sediment, is squeezed upwards between the deformed ripples to form commonly, mushroom-shaped "anticlines". The distribution of such anticlines may be regular or quasi-regular, provided the ripples are distributed at more or less regular spaces. Parenthetically, it may be added that the here described load-deformed ripples are occasionally "homogenized" by early diagenetic ankeritization and transformed into concretionary bodies in which the primary structure is partly or entirely obliterated. In such cases, the mechanical deformation is accompanied by physico-chemical replacement induced by connate waters flowing towards the region of maximum deformation (Dzulynski & Smith, 1963).

Among contortions which do not fall into the category of convolute lamination sensu stricto there is "crinkled lamination" or "curly bedding" (Sutton & Watson, 1960). The structure consists of irregularly contorted and, commonly, discontinuous lamination. It does not reveal the cusp and point-up design and is locally disrupted by sandy injections.

The crinkled lamination here under discussion, mantles the tops of thick and coarse turbidites (Pl. 43A, B). The chaotic jumble of contorted laminae is, in places, haphazardly interpenetrated by and interfolded with injected coarse sand veins from the subjacent coarse turbidite.

An insight into the possible origin of the deformation in question was given by an unpublished experiment, carried out by the present author many years ago. In this experiment, a dense watery suspension of sand, silt and plaster-of-paris was released to flow in a water tank with settled laminated clay at the bottom. The clay was rised by the advancing front of of the t-curren to form dilute clouds os suspension above the dense moving slurry. In the waning stage of of the flow, the clouds began to settle intermittently upon not yet stabilized and slowly moving mass of coarse sand. Under such conditions, the deposition from the settling clouds resulted in the formation of crinkled laminae (Pl. 43C). It is tentatively suggested, that the deformation shown in Plate 43 (A, B) might have been formed in a similar way.

Although it is not in direct relationship to the subject discussed in this paper, brief comment is made on contortions which originate in a sequence of soft horizontal layers or laminae which is pressed down at one end by obliquely imposed burden. In this context, reference is made to experiments by Rettger (1935) and Bucher (1956), in which the obliquely imposed load threw the laminae into a series of folds and small thrusts.

The structure varied from overturned folds near the source of pressure to almost symmetrical folds away from it. Significantly, the deformations decreased upward and downward from a maximum attained in the middle of the sequence (Bucher, 1956). In this respect, the deformations were similar to convolute lamination, but differed from it in the absence of cusp and point-up pattern.

Deformations corresponding to Bucher's experiments occur in nature. On a small scale, such deformations are observed in association with ice-, and plant-drifted dropstones or rock fragments falling from the roofs of karst caverns unto soft stratified substratum. On a larger scale, such deformations are exemplified by folding of thin layers in the lacustrine Pleistocene Lizan formation in Israel (e.g. Arkin & Michaeli, 1896). These deformations resulted presumably as a consequence of a sudden lowering of the lake level and loading of the marginal parts of the fine-grained laminated sediment by huge landslides following the lake level fall. On much larger scale, similar deformations are observed in front of nappes thrust upon molasse sediments in the foreland of the Carpathians (Krzywiec, 1994).

Upper boundaries of covoluted and contorted beds

As indicated previously, the convolutions and related contortions commonly die out upwards, passing into undisturbed horizontally laminated sediment. In other instances, the convolute lamination may pass gradually into rippled sediments showing the same composition and transport direction as the remaining host bed. In places, the ripples may show minor posterior deformations indicative of deposition upon not yet entirely stabilized substratum. It is to be noted that the convoluted and rippled parts of the bed are laid down by the same current, the rippled portion being deposited by the current's tail.

In amalgamated beds, the convolutions may be truncated by erosion surfaces and covered with a new turbidite. Such surfaces are commonly irregular and bear clear evidence of scouring (Pl. 44A). Of particular interest are the convolutions or contortions which are truncated by sharply delineated flat surfaces devoid of any evidence of current scour, and covered with horizontally laminated fine-grained silt or mud (Pl. 44B). The truncating surfaces of this kind were produced experimentally by Anketell *et al.* (1969) in the following way. Unstable horizontally laminated sequence was produced in water tanks by alternate deposition of variously coloured clays, silts and fine sands, settling from dilute suspensions. The deformation started spontaneously with the deposition of a critical layer that raised the total weight of the column beyond the limits of its bearing capacity. In other experiments the deformation was triggered of by a slight shock. The deformational process followed the well known pattern of density controlled deformations (Pl. 44C). As long as the deformation occured within the sediment, the laminae, although streched and thinned, preserved their identity. On reaching the sediment-water interface, however, the rising columns disintegrated, producing clouds of suspension. The interface was transformed into flat surface truncating the folded laminae (Pl. 44C), and the previously mentioned suspension clouds settled on this surface to form a set of fine horizontal laminae. In the experiment under consideration, the deformations were irregular because the requirement of structural homogeneity was not satisfied. The point in question, however, was that such truncated surfaces originated in the absence of unidirectional current erosion. The surfaces of this kind are encountered among natural convolutions (Pl. 44B). It is realized, however, that flat surfaces, truncating the deformed laminae of turbidites may be produced by bottom currents (not t-curents) which sweep the bottom under conditions of non deposition.

DISTURBANCES PRODUCED BY IMPACT OF HEAVY SUSPENSIONS

We come now to structures produced by the impact of heavy suspensions upon horizontally stratified bottom sediment. Such impact may result in the formation of new beds showing slump-like deformations. Without entering into the general subject of slump and sediment-flow deposits, described and illustrated in numerous papers, we shall concentrate on some selected problems.

Let us begin with one experiment described by Anketell and Dżułyński (1968b) in which diluted and variously coloured plaster-of-paris suspensions, mixed with small amounts of clay, were successively released to form a sequence of laminae on top of a clay layer. Then a dense plaster-of-paris suspension was allowed to flow over the still soft sequence. The flowing suspension threw the laminated sediment into a series of flow-transverse structures showing "pseudonodular" cross-sections (Pl. 45A, B). The corrugated sequence became truncated and amalgamated with the newly deposited turbidite, without being incorporated as clasts into its matrix. The clay wisps, rising from the layer underlying the laminated sequence, were vertical or deflected down-, or upcurrent. The transverse structures of the type discussed are occasionaly observed attached to bottom surfaces of thick natural turbidites. On a smaller scale, similar structures occur also among sole markings (Pl. 44C). However, the mechanism through which such sole markings are produced is not clear, notably if transverse structures are devoid of internal lamination. In such situation, particularly obscure is their relationship to structures resulting from progressing liquefaction front and to transverse scales showing similar morphology. There is also a possibility of their independent and different origin cannot be eliminated (as a curent shear effect).

The impact of heavy slurries upon soft stratified substratum also leads to its erosion, whereby the ripped-off fragments of the substratum are lifted and incorporated into the moving slurry. The case in question was investigated experimentally in water tanks in which the layered bottom sediment was produced by successive introduction of diluted and variously coloured plaster-of-paris and clay suspensions (Dżułyński & Radomski, 1966). The bottom sediment, thus prepared, was then acted upon by dense t-current

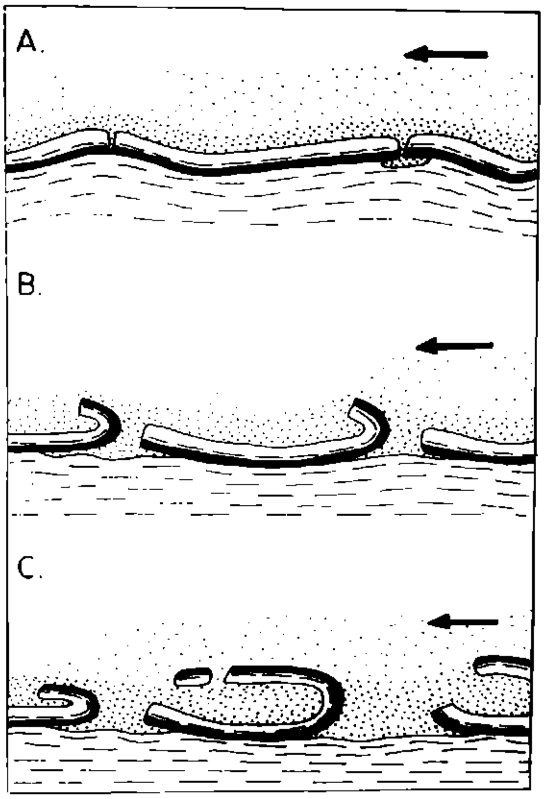

Fig. 33 Sketch showing formation of prolapse folds

made up of plaster-of-paris suspension and coarse sand. In proximal parts of the flume, the bottom layers, impoverished in clay, were broken into angular fragments. These fragments were incorporated into the moving suspension and scattered chaotically or concentrated at specific levels of the settling sediment. Such levels coresponded to drastic changes in density of the settling suspension (Pl. 46D). With very dense suspensions, the fragments were concentrated near or at the top of the deposited slurry. Further away from the discharge, the layers of the substratum were enriched in clay and thus prone to plastic deformation. The detached fragments of such layers were folded, contorted and overthrust in a highly complicated manner (Pl. 46B). The above mentioned experimental structures find their exact equivalents among natural structures (Pl. 46A, C).

A notable feature of the experiments under consideration is the appearance of "prolapse folds" (Wood & Smith, 1959) or "slump overfolds" (Crowell, 1957), i.e., the flat-lying recumbent folds with horizontal axes trending at right angle to the direction of flow. In experiments, such folds were seen to form when the passing suspension was bending the layers of the substratum into flow-transverse broad undulations whose crests broke up, producing transverse fissures. The fissures were immediately filled with the sinking suspension, which then penetrated laterally between the layers causing their further splitting (Fig. 33). The margins of the cleaved layers were bent upwards and those facing upcurrent were peeled off and transformed into flat recumbent folds (Pl. 47A). The overturned limbs of folds also broke into tabular fragments. Such slabs were sometimes observed to glide on top of the settling slurry (comp. e.g. Postma *et al.*, 1988) and, occasionally, preserved in it in upside-down position. This observation explains the occasional occurence of large overturned slabs on top of massive sandy turbidites and sediment flows in flysch.

Another notable feature of the experiments discussed was the appearance of the already mentioned longitudinal folds or fold-like deformations whose axes were parallel to the direction of flow (Pl. 47B, C,). In most instances, such deformations occur in the form of closely set isolated gutter-like bodies (Pl. 47D).

The stress distribution responsible for the formation of longitudinal folds, generated also longitudinal fractures in the subjacent layers. Such fractures were filled and widened by the sinking slurry, which then was injected laterally along suitable bedding planes (Pl. 49A, B, comp. also Hiscott, 1979) in much the same way as was the case with incipient prolapse folds or forcible detachment of clasts by sinking slurry from layered bottom sediment (comp. Pl. 1C). By this mechanism, large portions of the bottom layers may be lifted and entrained in the slurry. In the presence of intersecting transverse fractures, the ripped-off bodies acquire loaf-shaped forms (Pl. 49C). It is to be noted that the margins of such bodies are also warped upwards – a feature observed in some experimental and natural clastic wedges produced by downward injection of viscous and heavy materials into a layered substratum (Dżułyński, 1965a).

Longitudinal folds and fold-like disturbances are common in flysch rocks (Pl. 48A, B). Such disturbances are also observed in peaty solifluction tongues (Dżułyński & Pekala, 1980) and snow avalanches entering narrow pasages (Lajoie, 1972). Consequently, the orientation of fold axes in slump and sdiment flow deposits alone, should not be taken as a safe criterion of the strike of paleoslope (Hansen *et al.*, 1961; Murphy & Schlanger, 1962; Enos, 1969) showed that fold axes in natural subaerial and subaqueous slumps or sediment flows are not always normal to the direction of movement. Consequently, the orientation of axes alone, should not be taken as a safe criterion of the strike of paleoslope.

The existence of slump-like folds produced by impact of heavy suspensions on flat bottom surfaces illustrates the difficulties one may encounter in differentiating between slump generated and impact-induced slump-like disturbances in poorly exposed rocks. The chief criterion, indicative of the impact induced deformation, is the presence of exotic material in the matrix enclosing the contorted fragments, a quality which may be overlooked by an unpreposessing observer. Such material includes exotic rock fragments derived from outside the depositional basin which do not occur in the subjacent sediments. The slumps proper are generally devoid of such fragments and consist of material derived exclusively from the underlying strata.

It is to be noted that heavy slurry moving along resitant layers, which do not yield by plastic flow, produce in them open fissures which may be filled either from above or from below by by the subjacent soft sediment. Such features are common in turbidite sequences and confirmed by experiments (Pl. 48C).

CONCLUDING REMARKS

Structures to which appeal has been made in the preceding sections need not be summarized. As noted, one of the objectives of this paper was to indicate that genetically consanguineous sedimentary structures are sometimes divided by arbitrary classifications. Despite the wide spectrum of morphologies exhibited by erosional and deformational structures, many of these structures may be attributed to the same or similar formative processes (comp. Allen, 1984). A large number of such structures represent the "frozen" stages of one formative process. In geologic literature, such stages have been commonly assigned different names and regarded as genetically unrelated features.

Between erosion, deformation and deposition structures there are gradations which indicate overlaping and transitions between the formative processes. Thus, the change in morphology of prod markings to flute markings occurs consecutively without a clear separation between them. The differences in morphology of sole markings also reflect the differences in mechanical behavior of the sediment involved in their formation.

There is dispute about the role of bottom sediment in the formation of flute markings. This dispute has been concentrated on two sharply contrasted concepts; the "passive bed theory" and "defect theory" of erosional features (Allen, 1971a). According to the first concept, the pattern of turbulences within the flowing suspension controls the shape and distribution of sole markings independantly of bottom irregularities. The second concept implies that bottom irregularities determine the formation of such markings. The argument stems from discussion concerning the origin of flutes and scallops formed, on exposed rock surfaces, by water and air currents. According to Maxson (1940, p. 720), "the form of the flutes is determined by vortices formed by turbulent flow along large more or less flat surfaces and by disturbance of flow along irregularities of all types". Curl (1959) insisted that flow properties alone may control the shape and size of flutes in caves. According to other authors, the cave flutes are determined by surface irregularities (Colleman, 1947; Rudnicki, 1960). These latter ideas provided support for Allen's "defect theory" of flute markings preserved as moulds on bottom surfaces of turbidites.

The two ways of viewing the origin of flute markings are not mutually exclusive. The irregularities, preexistent in bottom sediments or inflicted by tools impinging upon the bottom are among the most important factors controling the locationand formation of flute markings and related erosional features on undersurfaces of current-deposited beds. On the other side, the vortices inherent to a turbulent flow and apparently independent of bottom irregularities, also may give rise to morphologically similar markings. It should be borne in mind, however, at any given part of the bottom, the pattern of streamlines, seemingly independent of bottom irregularities, may reflect earlier flow properties, and these might have been determined by earlier, undisclosed bottom defects. In addition, the unequal distribution of suspended load, the presence or absence of tools and the changing density of the flowing suspension, also are factors controling the development and distribution of scour markings. These factors cannot be assesed quantitatively from the structure of resulting sediments.

Parenthetically, it may be noted that scour and tool marks on bottom surfaces of turbidites find close analogues among features produced on glaciated rock surfaces by subglacial melt waters or moving glacier ice armored with hard rock fragments (comp. Chamberlin, 1885/86; Dahl, 1965). Also among tectonic marking or "tectoglyphs" (term after Teisseyre, 1921) on slickensided fault surfaces, there are analogues of groove and prod markings revealed by the soles of turbidites. Some of such tectoglyphs were given names coresponding to their sedimentary analogues, e.g., tectonic prod marks (Dżułyński & Kotlarczyk, 1965).

As indicated in the preceding sections, the interaction between the flow and the bottom depends, among other things, upon mechanical properties of the bottom sediment. For example, if the bottom sediment consists of incoherent sandy particles, the flow in double rolls results in entrainment and redistribution of grains. The grains are rearranged to form closely spaced ridges alligned parallel to the flow, as in the experiments on bed forms produced in sand (e.g. Crickmay, 1960; Karcz, 1970; Karcz & Kersey, 1980). If, however, the bottom sediment is plastic and shows a high degree of ductility, the same kind of currents may result in deformations with little or no entrainment of particles. Consequently, there is no contradiction between the "deformational" and "erosional" explanation of l-ridges. It is the pattern of motion that determines the form of structure. The genesis of spiral cross-currents responsible for the formation of l-ridges is disputable. This question, however, has more than one explanation.

Soft-sediment deformations may be contemporaneous with, or postdate the final deposition of particles which make up the resulting bed. In the first case, the particles settling upon the bottom may be still parts of the same flowing current. Consequently, the resulting structures are, with reason, called current structures. It is realized, however, that liquefaction of the already deposited sediment may set the particles in motion in much the same way as is the case with the settling suspension. The resulting post-depositional structures may be morphologically undistinguishable from syndepositional features.

Accordingly, it is often impossible to designate the deformation structures clearly in terms of their time of formation vis-a-vis the time of final deposition (Dzułynski & Walton, 1963).

Much controversy has arisen from a simple misconception that deformation structures alone may provide information on the direct cause of deformation. Hence, the abusive employment of genetic names. However, in disturbances generated by instability in density stratification, the work on deformation is exerted by potential energy stored in an unstable system. The "triggers" releasing this energy are not necessarily imprinted in the resulting structure and the conclusions concerning their nature are conjectural and may often be fraught with uncertainty. In such instances it is not warranted to specify explicitly the origin of structures in their denomination.

Morphologically and genetically identical deformations occur in various and, commonly, very different environments. Processes considered to be characteristic of a given environment have been often implicitly or explicitly identified with as being responsible for the formation of the structures discussed (Anketell *et al.*, 1970). For instance, the "involutions" or polygonal soils have been regarded as produced by the action of frost. These structures were considered as reliable indicators of cold climatic environments, because they, indeed, are very common in periglacial sediments. However, they also occur in other environments.

Many deformation structures in sediments are duplicated in igneous rocks. Thy and Wilson (1980) described and illustrated instructive examples of polygonal load casts from layered basic intrusions of the Fongen-Hyllingen complex in Norway. The present author observed similar features at the base of basaltic lavas laid down on rough, blocky and porous upper portions of the underlying lavas (Dżułyński, 1979). Needham (1978) called attention to large (up to 200 m in diameter) circular and polygonal load deformations produced by extrusion of basaltic lava onto water-saturated sand in Proterozoic rocks of Australia.

Bearing on the foregoing discussion are also parallel ridges with projecting excrescences described by Elwell *et al.* (1960) from interfaces separating different kinds of magma. The ridges were produced when the heavy magma was injected in the form of a sill upon not yet entirely solidified lighter magma. Elwell and others (1960) suceeded in producing experimentally a close analogue of the ridges in question by pouring the molten gum onto a layer of solid parafin. Similar structures were also reported by Nichols (1936) from a close-fitting contact between two Quaternary basaltic flows in which a massive lava was superimposed upon a vesicular lava (see also, Shrock, 1948, fig. 327).

The general rule in all the examples just cited is the presence of unstable density stratification in non-mobile and mobile ba systems. In this respect, the above mentioned igneous structures are analogous to some soft-rock deformations in layered sediments.

Longitudinal ridges occur also on walls of basaltic dykes cutting across horizontally disposed limestones (Baer, 1991). In such situation, the ridges are due to lateral pressure imposed on the walls of dykes by intruding lava. Similar ridges may occur within the igneous dykes by pressure of intruding magma on partly crystalized and thus more viscous margins of the dyke (Campbell, 1978).

Note of warning, however, is necessary concerning analogies based upon similarity in morphology of deformational structures. Nature has been prodigal in providing different mechanisms for bringing about morphologically similar structures. For example, the tadpoles struggling for air at the bottom of shallow pools give rise to polygonal nests whose moulds would be morphologically identical with polygonal load-casts. The formative agents are here different and the only feature in common is a tendency towards the appearance of surfaces of minimum potential energy, a property shared by both organic and inorganic structures. In this context, a passing reference may be made on ablation polygons in snow which are formed by convection cell-vortices of air on contact with the snow surface ce (e.g. Leighly, 1948; Richardson & Harper, 1957; Jahn & Kłapa, 1968). Also some of the polygonal taffonis might have originated in the same way. Consequently to asses the true nature of deformations in sedimentary rocks it is necessary to take into account the whole assemblage of environmental and related data.

The deformational structures resulting from instability in density stratification are independent of scale. The experiments designed to produce such structures are based on the assumption that within certain limits "Nature on the small scale and on the great acts always by the same laws and is ever analogous to herself" (Paul Frizi, 1762, quotation after Mavis *et al.*, 1935).

It is no coincidence that the previously described experiments meant for production of minor sedimentary deformations were, in principle, similar to those designed to produce tectonic and igneous features (e.g. Ramberg, 1971). Minor sedimentary deformations may be viewed as scaled down natural analogues of gigantic diapiric bodies. As indicated by Anketell *et al.* (1970), the investigation of the one may provide an insight into the others.

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REFERENCES

Aalto, K. R., 1995. Skin-friction lines preserved on flute casts, Franciscan Complex, Point St. George, Northern California. *J. Sediment. Res.*, A65: 127-128.

Ahnert, F., 1994. Modelling the development of non-periglacial sorted nets. *Catena*, 22: 43-63.

Aigner, T., 1985. *Storm depositional systems. Lecture Notes in Earth Sciences*, 3. Springer-Verlag, Berlin, 174 pp.

Allen, J. R. L., 1968. Flute marks and flow separation. *Nature*, 219: 602-604.

Allen, J. R. L., 1971a. Bedforms due to mass transfer in turbulent flows: a kaleidoscope of phenomena. *J. Fluid. Mech.*, 49: 40-63.

Allen, J. R. L., 1971b. Transverse erosional markings of mud and rock; their physical basis. *Sediment. Geol.*, 5: 167-185.

Allen, J. R. L., 1977. The possible mechanics of convolute lamination in graded sand beds. *J. Geol. Soc. London*, 134: 20-31.

Allen, J. R. L., 1984. *Sedimentary structures, their character and physical basis. Developments in Sedimentology*, 30, vol. 1+2. Elsevier, Amsterdam, 593+663 pp.

Anketell, J. M., Cegla J., & Dzulynski, S., 1969. Unconformable surfaces formed in the absence of current erosion. *Geol. Romana*, 8: 41-46.

Anketell, J. M., Cegła, J. & Dżułyński S., 1970. On the deformational structures in systems with reversed density gradient. *Rocz. Pol. Tow. Geol.*, 40: 3-29.

Anketell, J. M. & Dżułyński, S., 1968a. Patterns of density controlled convolutions involving statistically homogeneous and heterogeneous layers. *Rocz. Pol. Tow. Geol.*, 38: 401-409.

Anketell, J. M. & Dżułyński S., 1968b. Transverse deformational patterns in unstable sediments. *Rocz. Pol. Tow. Geol.*, 38: 411-416.

Arkin, Y. & Michaeli, L., 1986. The significance of shear strength in the deformation of laminated sediments in the Dead Sea area. *Israel J. Earth Sci.*, 35: 61-72.

Artyushkov, E. V., 1965. Convective deformations developed in feebly lithified sedimentary rocks. (In Russian only). *Izviest. Akad. Nauk. USSR, Ser. Geol.*, 12: 1-118.

Baer, G., 1991. Mechanism of dike propagation in layered rocks and in massive porous sedimentary layers. *J. Geophys. Res.*, B7: 911- 929.

Bartsch-Winkler, S. & Schmoll, H. R., 1984. Bedding types in Holocene Tidial sequences, Knik Arm, Upper Cook Inlet, Alasca. *J. Sediment. Petrol.*, 54: 1239-1250.

Benard, H., 1901. Les tourbillons cellulaires dans une nappe liquide transportant de la chaleur par convection en regime permanent. *Ann. Chimie Phys.* Ser. 7, 33: 62-144.

Best, J., 1992. On the entrainment of sediment and initiation of bed defects: Insight from recent developments within turbulent boundary layer research. *Sedimentology*, 30: 729-797.

Birkenmajer, K., 1958. Oriented flowage casts and marks in the Carpathian flysch and their relation to flute and groove casts. *Acta Geol. Polon.*, 8: 117-148.

Boer, De P. L., 1979. Convolute lamination in modern sands of the estuary of the Oosterschelde, the Netherlands, formed as the result of entrapped air. *Sedimentology*, 26: 283-294.

Bouma, A. H., 1962. *Sedimentology of some flysch deposits. A graphic approach to facies interpretation.* Elsevier, Amsterdam, 168 pp.

Brenchley, P. J. & Newall, G., 1977. The significance of contorted bedding in Upper Ordovician sediments of the Oslo region, Norway. *J. Sediment. Petrol.*, 47: 819-833.

Brodzikowski, K., 1981. Dilatancy and the cause of the deformational processes in unconsolidated sediments. *Ann. Soc. Geol. Polon.*, 51: 83-98.

Brodzikowski, K. & Haluszczak, A., 1987. Flame structures and associated structures in Quaternary glaciolacustrine and glaciodeltaic deposits: examples from central Poland. In: Jones, M. E. & Preston, R. M. F. (eds), *Deformations of sediments and sedimentary rocks. Geol. Soc. Spec. Publ.*, 29: 279-286.

Brodzikowski, K. & Van Loon, A. J., 1985. Inventory of deformational structures as a tool for unraveling the Quaternary geology of glaciated areas. *Boreas*, 14: 175-188.

Bucher, W. H., 1956. Role of gravity in orogenesis. *Geol. Soc. Am. Bull.*, 67: 1295-1318.

Bull, P. A., 1978. Surge mark formation and morphology. *Sedimentology*, 25: 877-886.

Bülow, W. V., 1964. Ein Taschen- und Tropfenboden von Grebs. *Geologie*, 13: 353-360.

Butrym, J., Cegła, J., Dżułyński, S. & Nakonieczny, S., 1964. New interpretation of "periglacial structures". *Folia Quatern.*, 17: 1-34.

Campbell, I. H., 1978. Some problems with the cumulus theory. *Lithos*, 11: 311-323.

Cegła, J., 1973. Tentative explanation of origin of structural grounds in SW Spitsbergen in the light of the hypothesis of systems with reversed density gradient. (In Polish, English summary). *Czasopismo Geograficzne*, 44, (2): 237-243.

Cegła, J. & Dżułyński, S., 1967. Experiments on feather fracture in sediments. *Rocz. Pol. Tow. Geol.*, 37: 489-497.

Cegła, J. & Dżułyński, S., 1970. Systems with reversed density gradient and their occurrence in periglacial zones. (In Polish, English summary). *Acta Univ. Wratisl.*, 124, *Studia Geogr.*, 13: 17-42.

Chamberlin, T. C., 1885/1886. The rock-scouring of the Great Ice Invasion. *U.S. Geol. Surv. Rep.*, 17: 155-248.

Chandra, K., 1938. Instability of fluids heated from below. *Proc. Roy. Soc.*, Ser. A, 164: 231-142.

Chang, Y. L., 1939. Laboratory investigations of flume traction and transportation. *Trans. Am. Soc. Civil Eng.*, 104: 1246-1283.

Clarke, J. M., 1917. Strand and undertow markings of Upper Devonian time as indications of prevailing climate. *New York State Museum Bull.*, 196: 199-238.

Colleman, J. C., 1947. An indicator of water flow in caves. *Geol. Mag.*, 82: 138-139.

Conybeare, C. E. B. & Crook K. A. W., 1968. Manual of sedimentary structures. *Dept. Nat. Devel., Bur. Miner. Res., Geol. Geoph. Comm. Austr. Bull.*, 102: 1-327.

Cowan, C. A. & James, N. F., 1992. Diastasis cracks: mechanically generated syneresis-like cracks in Upper Cambrian shallow water oolite and ribbon carbonates. *Sedimentology*, 39: 1101-1118.

Craig, G. Y. & Walton, E. K., 1962. Sedimentary structures and palaeocurrent directions from the Silurian rocks of Kirckudbrightshire. *Trans. Edinburgh Geol. Soc.*, 19: 100-119.

Crickmay, C. H., 1960. Lateral activity in a river of Northhwestern Canada. *J. Geol.*, 68: 377-395.

Crowell, J. C., 1955. Directional current structures from the Prealpine Flysch. *Geol. Soc. Am. Bull.*, 66: 1361-1384.

Crowell, J. C., 1957. Origin of pebbly mudstones. *Geol. Soc. Am. Bull.*, 68: 993-1010.

Curl, R. L., 1959. Flutes: geometry and propagation. *Nat. Speleol. Soc. Am. News*, 17: 121-160.

Czeppe, Z., 1961. Annual course of frost ground movements at Hornsund (Spitsbergen) 1957-1958. *Schedae Geogr., Univ. Jaga. Acta Scien. Liter.*, 42: 1-74.

Dahl, R., 1965. Plastically sculptured forms on rock surfaces in Northern Nordland, Norway. *Geogr. Ann.*, 47: 1-140.

Davenport, C. & Ringrose, P. S., 1987. Deformations of Scottish Quaternary sediment sequences by strong earthquake motion. In: Jones, M. E. & Preston, R. M. F. (eds), *Geol. Soc. Spec. Publ.*, 29: 299-314.

Davies, H. G., 1965. Convolute lamination and other structures from the Lower Coal Measures of Yorkshire. *Sedimentology*, 5: 305-325.

Dewers, F., 1934. Krtitische Betrachtungen zur Frage der Beschaffenheit und Enstehung der Taschenboden. *Abh. u. Vortrage Bremmer Wiss. Ges., Jb.*, 8/9: 306-330.

Dionne, J. C., 1971. Contorted structures in unconsolidated Quaternary deposits, Lake Saint-Jean and Saguenay regions, Quebeck. *Rev. Geogr. Montreal.*, 25: 5-33.

Dionne, J. C., 1973. Structures cylindriques vericales dans du Arthabasca, Quebeck. *Sediment. Geol.*, 9: 53-66.

Doe, T. W. & Dott, R. H. Jr., 1980. Genetic significance of deformed bedding - with examples from the Bavajo and Wever Sandstone of Utah, *J. Sediment. Petrol.*, 50: 793-812.

Dott, R. H. Jr. & Howard, J. K., 1962a. Convolute lamination in nongraded sequences. *J. Geol.*, 70: 114-121.

Dott, R. H. Jr. & Howard, J. K., 1962b. Convolute Lamination: A reply. *J. Geol.*, 71: 659-660.

Dunbar, C. O. & Rodgers, J., 1957. *Principles of Stratigraphy.* Willey, New York., 356 pp.

Dżułyński, S., 1963a. Directional structures in Flysch. *Studia Geol. Polon.*, 12: 1-136.

Dżułyński, S., 1963b. Polygonal structures in experiments and their bearing upon some periglacial phenomena. *Bull. Acad. Pol. Sci., Ser. Geol. Geogr.*, 11: 145-150.

Dżułyński, S., 1965a. New data on experimental production of sole markings. *J. Sediment. Petrol.*, 35: 196-212.

Dżułyński, S., 1965b. Experiments on clastic wedges. *Bull. Acad. Pol. Sci., Ser. Geol. Geogr.*, 13: 301-304.

Dżułyński, S., 1966. Sedimentary structures resulting from convection-like pattern of motion. *Rocz. Pol. Tow. Geol.*, 36: 3-21.

Dzulynski, S., 1979. Origin, geological history and environmental factors leading to development of structurally complex formations. *Int. Symp. The Geotechniques of Structurally Complex Formations*, vol. II. Associazione Geotechnica Italiana, Milano, pp.157-166.

Dżułyński, S., 1983. Geology of the Gurvan Turuu area. In: Breymeyer, A. & Klimek, K. (eds), Mongolian dry steppe geosystems: a case study of Gurvan Turuu area. *Geographical Studies, Spec. Issue*, 2, vol. III. Polish Academy of Sciences, Warszawa, pp. 16-20.

Dżułyński, S. & Kotlarczyk, J. 1962. On load-casted ripples. *Rocz. Pol. Tow. Geol.*, 30: 214-241.

Dżułyński, S. & Kotlarczyk, J., 1965. Tectoglyphs on slickensided surfaces. *Bull. Acad. Pol. Sci., Ser. Geol. Geogr.*, 13: 149-155.

Dzulynski, S. & Pekala, K., 1980. Bog cirques and solifluction valleys in granitic rocks. *Z. Geomorph.*, 24: 219-235.

Dżułyński, S. & Radomski, A., 1955. Origin of groove casts in the light of turbidity current hypothesis. *Acta Geol. Polon.*, 5: 47-56.

Dżułyński, S. & Radomski, A., 1956. Clastic dikes in the Carpathian flysch. *Rocz. Pol. Tow. Geol.*, 26: 47-56.

Dżułyński, S. & Radomski, A., 1966. Experiments on bedding disturbances produced by impact of heavy suspensions upon horizontal sedimentary layers. *Bull. Acad. Pol. Sci., Ser. Geol. Geogr.*, 14: 227-230.

Dżułyński, S. & Sanders, J., 1962a. On some current markings in Flysch. *Rocz. Pol. Tow. Geol.*, 30: 144-145.

Dżułyński, S. & Sanders, J., 1962b. Current marks and firm bottom mud. *Trans. Conneticut Acad. Arts. Sci.*, 42: 57-96.

Dzulynski, S. & Simpson, F., 1966a. Experiments on interfacial current markings. *Geol. Romana*, 5: 197-214.

Dżułyński, S. & Simpson F., 1966b. Influence of bottom irregularities and transported tools upon experimental sole markings. *Rocz. Pol. Tow. Geol.*, 36: 284-294.

Dżułyński, S. & Ślączka, A., 1958. Directional structures and sedimentation of the Krosno beds. *Rocz. Pol. Tow. Geol.*, 28: 205-260.

Dżułyński, S. & Ślączka, A., 1959. An example of large-scale bottom erosion in the flysch basin. *Rocz. Pol. Tow. Geol.*, 29: 327-333.

Dżułyński, A. & Ślączka, A., 1960. Sole markings produced by fish bones acting as tools. *Rocz. Pol. Tow. Geol.*, 30: 249-255.

Dżułyński, S. & Ślączka, A., 1965. On ripple-load convolution. *Bull. Acad. Pol. Sci., Ser. Geol. Geogr.*, 13: 135-139.

Dzulynski, S. & Smith, A. J., 1963. Convolute lamination its origin, preservation and directional significance. *J. Sediment. Petrol.*, 33: 616-627.

Dżułyński, S. & Smith, A. J., 1964. Flysch facies. *Rocz. Pol. Tow. Geol.*, 34: 239-266.

Dzulynski, S. & Walton, E. K., 1963. Experimental Production of sole markings. *Trans. Edinbourgh Geol. Soc.*, 19: 279-305.

Dżułyński, S. & Walton E. K., 1965. *Sedimentary features of flysch and graywackes. Developments in Sedimentology*, 7. Elsevier, Amsterdam, 274 pp.

Dzulynski, S., Ksiazkiewicz, M. & Kuenen, Ph. H., 1959. Turbidites in flysch of the Polish Carpathians. *Geol. Soc. Am. Bull.*, 70: 1089-1118.

Dżułyński, S., Shideler, G. L. & Ślączka A., 1972. Impact induced dendritic ridges in soft sediments. *Rocz. Pol. Tow. Geol.*, 42: 285-288.

Eardley, A. J. & White, M., 1947. *Flysch and Molasse. Geol. Soc. Am. Bull.*, 55: 976-999.

Einsele, G., 1963. "Convolute bedding" und ahnliche Sedimentstrukturen im rheinischen Oberdevon und anderen Ablagerungen. *N. Jb. Geol. Pal ont. Abh.*, 116: 162-198.

Einsele, G., Overbeck R., Schwarz, H. U. & Unsold, G., 1974. Mass physical properties, sliding and erodibility of experimentally deposited and differently consolidated clayey muds. *Sedimentology*, 21, 339-372.

Elton, C. S., 1927. The nature and origin of soil-polygons in Spitsbergen. *Quart. J. Geol. Soc.*, 83: 163-192.

Elwell, R. W. D., Skelhorn, R. R. & Drysdall. A. R., 1960. Inclined granitic pipes in the diorites of Guernsey. *Geol. Mag.*, 97: 89-105.

Emery, K. O., 1950. Contorted Pleistocene strata at Newport Beach, California. *J. Sediment. Petrol.*, 20: 111-115.

Emery, K. O., 1964. Turbidites -Precambrian to present. *Woodshole Oceanogr. Inst.Contr.*, 1423: 485-495.

Enos, P., 1969. Anatomy of a flysch. *J. Sediment. Petrol.*, 39: 680-723.

Eyles, N. & Clark, B. M., 1986. Gravity induced soft-sediment deformations in glaciomarine sequences of the Upper Proterozoic Port Askaig Formation, Scottland. *Sedimentology*, 32: 789-814.

Fairbridge, R. W., 1958. What is a consanguineous association. *J. Geol.*, 66: 319-323.

Flood, R. D., 1981. Distribution, morphology, and origin of sedimentary furrows in cohesive sediment, Southampton Water. *Sedimentology*, 28: 511-529.

Flood, R. D., 1983. Classification of sedimentary furrows and a model for furrow initiation and evolution. *Geol. Soc. Am. Bull.*, 94: 630-639.

Folk, R. L., 1976. Rollers and ripples in sand, streams and sky: rhythmic alteration of transverse and longitudinal vortices in three orders. *Sedimentology*, 23: 649-660.

Folk, R. L., 1977. Longitudinal ridges with tunning-fork junctions in the laminated interval of flysch beds: evidence for low-order helicoidal flow in turbidites. *Sediment. Geol.*, 19: 1-6.

Gilbert, G. K., 1914. The transportation of debris by running water. *U.S. Geol. Surv. Prof. Pap.*, 86: 1-263.

Graham, A., 1934. Shear patterns in unstable layer of air. *Phil. Trans. Roy. Soc.*, Ser. A: 232, 285-296.

Grass, A. J., 1971. Structural features of turbulent flow over smooth and rough boundaries. *J. Fluid. Mech.*, 50: 233-255.

Grass A. J., Stuart R. J. & Mansour-Tehrani, M., 1991. Vortical structures and coherent motion in turbulent flow over smooth and rough boundaries. *Phil. Trans. Roy. Soc. London*, Ser. A 336, no. 1640: 35-65.

Gripp, K., 1927. Beitrage zur Geologie von Spitsbergen. *Naturwiss. Ver. Hamburg. Abh.*, 21: 1-38.

Gripp, K., 1952. Zwei Beitrage zur Frage der Periglazialen Vorgange. *Meyniana*, 1: 112-118.

Gripp, K. & Simon, W. G., 1933. Experimente zum Brodelboden Problem. *Centralbl. Min., Abh.*, Abt. B: 433-440.

Halicki, B., 1960. Sur la differante genese des deformations des sediments dans un milieux hydroplastique. *Biul. Perygl.*, 7: 55-57.

Hall, J., 1843. *Geology of New York, 4.* Survey of the Fourth Geological District, Albany N.Y., 525 pp.

Hamberg, A., 1915. Zur Kentniss der Vorgange im Erdboden beim Gefrieren und Auftauen sowie Bemerkungen ueber die erste Kristallisation des Eises in Wasser. *Geol. Foren. Stockholm, Forh.*, 37: 583-619.

Hansen, E., Potter, S. C., Hall, B. A. & Hills, A., 1961. Decolment structures in glacial-lake sediments. *Geol. Soc. Am. Bull.*, 72: 1415- 1418.

Havley, J. E. & Hart, R. C., 1934. Cylindrical structures in Sandstones. *Geol. Soc. Am. Bull.*, 45: 1017-1034.

Hiscott, R., 1979. Clastic sills and dikes associated with deep-water sandstones. Tourelle formation. Ordovician, Quebeck. *J. Sediment. Petrol.*, 49: 1-10.

Holland, C. H., 1959. On convolute bedding in the Lower Ludlovian rocks of North-East Radnorshire. *Geol. Mag.*, 95: 230-236.

Holland, C. H., 1960. Load-cast terminology and origin of convolute lamination. *Geol. Soc. Am. Bull.*, 71: 633-634.

Hsü, K. J., 1970. The meaning of the word Flysch: a short historical research. In: Lajoie, J. (ed.), *Flysch sedimentology in North America. Geol. Ass. Can. Spec. Papers*, 7, pp. 1-11.

Hubert, J. F., Scott K. M. & Walton E. K., 1966. Internal groove markings in Silurian flysch sandstones. *J. Sediment. Petrol.*, 36: 237-241.

Hunt, C. B. & Washburn, A. L., 1960. Patterned ground. In: Hunt, C. B., Robinson. T. W., Bowles, W. A. & Washbrun, A. L. (eds), *Hydrologic Basin Death Valley California. U.S. Geol. Survey Prof. Pap.*, 494 B: 104-134.

Jahn, A. & Czerwiński, J., 1965. The role of impulses in the process of periglacial soil structure. *Acta Univ. Vratisl., 44, Studia Geograf.*, 7: 1-24.

Jahn, A. & Klapa, M., 1968. On the origin of ablation hollows (polygons) on snow. *J. Glaciology*, 7: 299-312.

Jan Rolfe. W. D., 1963. A fine air-heave structures from the Old Red Sandstone. *Geol. Mag.*, 97: 133-136.

Jerzykiewicz. T. & Norris D. K., 1994. Stratigraphy and syntectonic sedimentation of the Campanian "Belly River" clastic wedge in the southern Canadian Cordillera. *Cretaceous Res.*, 15: 367-399.

Johnson, A. M., 1970. *Physical processes in geology.* Freeman, Copper & Company, San Francisco, 577 pp.

Kaiser, K., 1965. Ein Beitrag zur Frage der Solifluktiongrenze in den Gebirgen Vorderasiens. *Z. Geomorph.*, 9: 461-474.

Karcz, I., 1968. Fluviatile obstacle marks from the Wadis of the Negev (South Israel). *J. Sediment. Petrol.*, 38: 1000-1012.

Karcz, I., 1970. Possible significance of transition flow in interpretation of origin of some natural bed forms. *J. Geoph. Research*, 75: 2869-2873.

Karcz, J., 1972 Sedimentary structures formed by flash floods in Southern Israel. *Sediment. Geol.*, 7: 161-182.

Karcz, I. & Kersey, D., 1980 Experimental study of free-surface instability and bedforms in shallow flows. *Sediment. Geol.*, 27: 263-300.

Keighley, D. E. & Pickerill. R. K., 1994. Flute-like marks and associated structures from the Carboniferous Port Hood formation of Eastern Canada: Evidence of secondary origin in association with sediment intrusion. *J. Sediment. Research*, A44: 253-263.

Kelling, G. & Walton, E. K., 1957. Load cast structures, their relationship to upper-surface structures and their mode of origin. *Geol. Mag.*, 94: 481-490.

Kindle, E. M., 1917. Deformations of unconsolidated beds in Nova Scotia and southern Idaho. *Geol. Soc. Am. Bull.*, 28: 323-334.

Klaczyńska, K., 1960. The flow folds and flow loads in the Quaternary sediments. (In Polish, English abstract). In: Zbiór Prac i komunikatów Treści Geologicznej. Instytut Geologiczny, Muzeum Ziemi, Warszawa, pp. 53-57.

Klysz, P., 1975. Plastic deformation in fossil depression at Smolina near Turek. *Badania Fizjograficzne nad Polską Zachodnią*, Ser. A, 27: 139-160.

Knapp, R. T., 1943. Density currents -their meaning and their effect on the turbulence structures of the associated flow. *University of Iowa Studies in Engineering Bull.*, 27: 289-306.

Kolar, V., 1956. Spiral motion in fluids. *Rozpravy Ceskoslov. Akad. Ved.*, 66: 1-118.

Kopstein, F. P. W. H., 1954. *Gradeded bedding of the Harlech Dome.* Thesis. State Univ. Groningen, Groningen, 97 pp.

Krejci-Graf, K., 1932. Definition der Begriefe - Marken, Spuren, Fahrten, Bauten, Hieroglyphen und Fukoiden. *Seckenbergiana Lethea*, 14: 19-39.

Krzywiec, P., 1994 Sedymentacja, tektonika i geodynamika centralnej części zapadliska przedkarpackiego - wnioski z interpretacji danych sejsmicznych. (In Polish only). In: *I Krajowa konferencja naukowo-techniczna, Część II - Zbiór referatów*, Instytytut Górnictwa Naftowego i Gazownictwa, pp. 157-171

Książkiewicz, M., 1958. Stratigraphy of the Magura series in the Średni Beskid (Carpathians). *Biul. Inst. Geol.*, 135: 43-96.

Książkiewicz, M., 1968. Observations on joints in the Flysch Carpathians. (In Polish, English summary). *Rocz. Pol. Tow. Geol.*, 38: 335-384.

Kuenen, Ph. H., 1953. Significant features of graded bedding. *Bull. Am. Assoc. Petrol. Geol.*, 37: 1044-1066.

Kuenen, Ph. H., 1957. Sole markings in graded grywacke beds. *J. Geol.*, 65: 231-258.

Kuenen, Ph. H., 1958. Experiments in Geology. *Trans. Geol. Soc. Glasgow*, 23: 1-28.

Kuenen, Ph. H., 1965. Value of experiments in geology. *Geologie en Mijnbouw*, 44: 22-36.

Kuenen. Ph. H., 1968. Settling convection and grain-size analysis. *J. Sediment. Petrol.*, 38: 817-831.

Kuenen, Ph. H. & Menard, H. W., 1952. Turbidity currents, graded and non-graded sediments. *J. Sediment. Petrol.*, 22: 83-96.

Kuenen, Ph. H. & Migliorini, C. I., 1950. Turbidity currents as a cause of graded bedding. *J. Geol.*, 55: 91-127.

Kuenen, Ph. H. & Prentice I. E., 1957. Flow markings and load casts. *Geol. Mag.*, 94: 173-174.

Laird, M. G., 1970. Vertical sheet structures - A new indicator of sedimentary fabrics. *J. Sediment. Petrol.*, 40: 429-434.

Lajoie, J., 1972. Slump fold axis orientation: indication of paleoslope, *J. Sediment. Petrol.*, 42: 584-586.

Lanteaume. M., Beadoin, B., & Canpredon, R., 1967. *Figures sedimentaires du Flysch "Gres D'Anot", Sinclinal de Peira-Cava.* Centre National de la Recherche Scientifique, Paris, 99 pp.

Lea, P. D., 1990. Pleistocene periglacial eolian deposits in southwestern Alasca: Sedimentary facies and depositional processes. *J. Sediment. Petrol.*, 60: 582-591.

Leighly, J., 1948. Cuspate surfaces of melting ice and firn. *Geograph. Rev.*, 38: 301-306.

Lonsdale, P., 1950. Manganese-nodule bedforms and thermohaline density flows in deep-sea valley on Carnegie Ridge, Panama Basin. *J. Sediment. Petrol.*, 50: 133-148.

Low, A. R., 1925. Instability of viscous fluid motion. *Nature*, 115: 299-300.

Lowe, D. R., 1975. Water escape structures in coarse-grained sediments. *Sedimentology*, 22: 157-204.

Lowe, D. R. & LoPiccolo R. D., 1974. The characteristics and origin of Dish and Pillar structures. *J. Sediment. Petrol.*, 44: 484-501.

Maas, H. & Muller, E., 1954. Bodenkundliche und geologische Hinweise zur Elster-Eiszeit bei Haltern (Westf.). *Geol. Jb.*, 69: 407-410.

Macar, P. & Autun. F., 1949-1950. Pseudonodules et glissement sous-aquatique dans l'Emsien inferieur de Oesling (Grand Douche de Luxemburg). *Ann. Soc. Geol. Belge*, 73: 121-150.

Macdonald, D. I. M., 1986. Proximal and distal sedimentological variations in linear turbidite troughs.; implictions for the fan model. *Sedimentology*, 33: 243-259.

Matthes, G. H., 1947. Macroturbulence in natural stream flow. *Trans. Am. Geoph. Union.*, 28: 255-261.

Mavis, F. T., Chitty Ho, & Yun-Cheng Tu, 1935. The transportation of detritus by flowing water. *Univ. of Iowa Studies in Engineering, New Series*, 294: 52 pp.

Maxson, J. H., 1940. Fluting and faceting of rock fragments. *J. Geol.*, 48: 717-751.

McBride, E. F., 1962. Flysch and associated beds of the Martinsbourgh Formation (Ordovician), Central Appalachians. *J. Sediment. Petrol.*, 32: 39-91.

McBride, E. F. & Yaekel, L. S., 1963. Relationship between parting lineation and rock fabric. *J. Sediment. Petrol.*, 33: 779-782.

McCallien, W. J., Ruxton, B. P. & Walton, B. P., 1964. Mantle rock tectonics: A study in Tropical Weathering at Accra, Ghana. *Overseas Geol. Min. Resources*, 9: 267-294.

McKee, E. D. & Goldberg, M., 1969. Experiments on formation of contorted structures in mud. *Geol. Soc. Am. Bull.*, 80: 231-244.

McKee, E. D., Reynolds, M. A. & Baker, Jr., C. H., 1962. Laboratory studies on deformation in unconsolidated sediment. *U.S. Geol. Surv. Prof. Pap.*, 45 D: 151-155.

Mead, W. J., 1925. The geologic role of dilatancy. *J. Geol.*, 33: 685-699.

Mellen, J., 1956. Pre-Cambrian sedimentation in the northeast part of Cohutta Mountain Quadrangle - Georgia. *Georgia Mineral Newsletter*, 9: 46-60.

Menard, H. W., 1950. Sediment movement in relation to current velocity. *J. Sediment. Petrol.*, 20: 148-160.

Mial, A. D., 1984. Flysch and Molasse: the elusive models. *Ann. Soc. Geol. Polon.*, 54: 281-291.

Mills, P. C., 1983. Genesis and diagnostic value of soft-sediment deformation structures - A review. *Sediment. Geol.*, 35: 83-104.

Morgenstern, N. R., 1967. Submarine slumping and the initiation of turbidity currents. In: Richards, A. E. (ed.), *Marine Geotechnique*. Univ. of Chicago Press, Chicago, pp. 189-220.

Mortensen, H., 1932. ber die physikalischen Moeglichkeit der "Brodel"- Hypothese: *Centralbl. Mineralog.*, Abt. B: 417-422.

Murphy, M. A. & Schlanger, S. D., 1962. Sedimentary structures in Ihas and Sao Sebastian Formation (Cretaceous), Reconcavo Basin. *Am. Ass. Petr. Bull.*, 46: 457-477.

Myrrow, P. M., 1992a. Pot and gutter casts from the Chapel Islands formation, Southeast Newfunland. *J. Sediment. Petrol.*, 62: 992-1007.

Myrrow, P. M., 1992b. Bypass-zone tempestite facies. Model and proximality trends for ancient muddy shorelines. *J. Sediment. Petrol.*, 62: 99-115.

Needham, R. S., 1978. Giant-scale hydroplastic deformation structures formed by loading of basalt onto water-saturated sand, Middle Proterozoic, Northern Territory, Australia. *Sedimentology*, 25: 285-295.

Nelson, C. H., Normark, W. R., Bouma, A. H. & Carlson, P. R., 1978. Thin-bedded turbidites in modern submarine canyons, fans and trenches. In: Stanley, D. J. & Keeling, G. (eds), *Sedimentation in submarine canyons fans and trenches*. Dowden, Hutchinson and Ross, Stroudsburg, Pensylvalnia, pp. 177-189.

Nettleton, L. L. & Elkins, T. A. 1947. Geologic models made from granular materials. *Trans. Am. Geoph. Union.*, 28: 451-465.

Nichols, R.L., 1936. Flow-units in basalt. *J. Geol.*, 44: 617-630.

Nichols, R. J., Sparks, S. J. & Wilson, J. N., 1994. Experimental studies of the fluidization of layered sediments and the formation of fluid escape structures. *Sedimentology*, 41: 233-253.

Norrman, J. O., 1964. Lake Vattern, investigations on shore and bottom morphology. *Geograf. Ann.*, 44: 1-138.

Nougier, J., 1964. Observationsperiglaciares dans l'Archipel de Kuerguelen. *Z. Geomorph.*, 8: 143-149.

Otvos Jr., E. G., 1965. Types of rhomboid beach surface patterns. *Am. J. Sci.*, 263: 271-276.

Owen, J. S., 1908. Experiments on the transporting power of sea currents. *Geogr. J.*, 31: 415-420.

Owen, G., 1987. Deformation process in unconsolidated sand. In: Jones, M. E. & Preston, R. M. F. (eds), *Deformations of sediments and sedimentary rocks. Geol. Soc. Spec. Publ.*, 29, pp. 11-24.

Panin, N., 1962. Neskolko sledov mehaniceskogo i organiceskogo Proishozdenia w miocenovih otlozeniach zony sliyania rek. (In Russian only). *Revue de Geologie et Stratigraphie (Bucuresti), Tirage A*, 6: 263-271.

Paola, Ch., Gust, G. & J. B. Southard, 1986. Skin friction behind isolated hemispheres and the formation of obstacle marks. *Sedimentology*, 33: 279-293.

Parkash, B., 1970. Downcurrent change in sedimentary structures in Ordovician turbidite graywackes. *J. Sediment. Petrol.*, 40: 572-590.

Parker, J. M., 1942. Regional systematic jointing in slightly deformed sedimentary rocks. *Geol. Soc. Am. Bull.*, 53: 381-408.

Pavoni, N., 1959. Rollmarken von Fischwirbeln aud dem Oligozanen Flyschschiefern von Engi-Matt (Kt. Glarus). *Ecloge Geol. Helv.*, 52: 941-949.

Peabody, F. E., 1947. Current crescent in the Triassic Moenkopi formation. *J. Sediment. Petrol.*, 17: 73-76

Pękala, K., 1964. Load structures in the polygenetic terrace of the San river at the mouth of the Wisznia river. (In Polish, English summary). *Ann. Univ. M. Curie-Skłodowska (Lublin), Sect. B.*, 19: 177-190.

Pettijohn, F. J. & Potter, P. E., 1964. *Atlas and Glossary of Primary Sedimentary Structures*. Springer-Verlag, Berlin, 370 pp.

Postma, G., Nemec, W. & Kleinspeha, K. L., 1988. Large floating clasts in turbidites; a mechanism for their emplacement. *Sediment. Geol.*, 58: 47-61.

Prentice, J. E., 1956. The interpretation of flow-markings and load casts. *Geol. Mag.*, 93: 393-400.

Ramberg, H., 1971. Dynamic models simulating rift valleys and continental drift. *Lithos*, 4, 259-276.

Rayleigh Lord, 1916. On convection currents in a horizontal layer. *Phil. Mag.*, Ser. 6, 192: 529-546.

Rettger, R. E., 1935. Experiments in soft rock deformations. *Bull. Am. Assoc. Petr. Geol.*, 19: 271-292.

Reynolds, O., 1885. On the dilatancy of media composed of rigid particles in contact. *Phil. Mag.*, Ser. 5, 20: 469-481.

Ricci Lucchi, F. 1969. Considerazioni sulla formazione di alcune inpronte de corrente. *Gior. di Geologia, Ann. del Museo di Bologna*, Ser. 2, 36: 363-415.

Richards, A. F. & Hamilton, E. L., 1967. Investigations of deep-sea sediment cores., III. In: A. F. Richards (ed.), *Consolidation. Marine Geotechnique*. Univ. of Urbana Press, Urbana, pp. 93-117.

Richardson, P. D., 1968. The generation of scour marks near obstacles. *J. Sediment. Petrol.*, 38: 965-970.

Richardson, W. E. & Harper, R. D. M., 1957. Ablation polygons on snow - furher observations and theories. *J. Glaciol.*, 3: 25-27.

Richmond, G. M., 1949. Stone nets, stone stripes and soil stripes in the Wind River, Wyoming. *J. Geol.*, 57: 143-153.

Rizzini, A. & De Rosa, E., 1969. Strutture sedimentarie in depositi fluvio-lacustri periglaciali. *Riv. Ital. Paleont.*, 75: 419- 451.

Rodgers, J., 1954. Terminology of limestones and related rocks; an interim report. *J. Sediment. Petrol.*, 24: 225-234.

Rohdenburg, H. & Mayer, B., 1969. Zur Deutung Pleistozaner Periglacial Formen. *Göttinger Bodenkundliche Berichte*, 7:

49-70.

Romanovsky, V. & Cailleux, S., 1942 Sols polygonaux et fentes de dessication. *Soc. Geol. Franc.*, 12: 321-327.

Rozanski, G., 1943. Stone-centered polygons. *J. Geol.*, 51: 330-341.

Rudnicki, J., 1960. Experimental work on flutes dvelopment. (In Polish, English summary). *Speleologia*, 2: 17-30, Warszawa.

Rücklin, H., 1938. Stromungsmarken im Unteren Muschelkalk des Saarlandes. *Senckenbergiana*, 20: 94-114.

Salomon, W., 1926, Die Rehbokischen Wasserwazen und ihre Bedeutung f_r die Erosion und Akkumulation. *Geol. Rund.*, 17: 418- 427.

Sanders, J. E., 1960. Origin of convoluted laminae. *Geol. Mag.*, 97: 409-421.

Sanders, J. E., 1965. Primary sedimentary structures formed by turbidity currents an related resedimentation mechanism. In: Middleton, G. V., (ed.), *Primary sedimentary structures and their hydrodynamic interpretation. Soc. Econ. Paleontol. Mineral. Spec. Publ.*, 12: 192-217.

Seilacher, A., 1963. Umlagerung und Rolltransport von Cephalopoden Gehausen. *N. Jb. Geol. Paleont. Mitt.*, 11: 593-615.

Seilacher, A., 1993. Fossillagerstätte im Muschelkalk. In: Hagdorn, H. & Seilacher, A. (eds), *Muschelkalk. Schontaler Symposium 1991*. Goldschneck-Verlag, Stuttgart, pp. 215-222.

Selley, R. C., 1964. The penecontemporaneous deformation of heavy mineral bands in the Torridonian sandstone of Northwest Scotland. In: van Straaten, L. M. J. U. (ed.), *Deltaic and shallow marine deposits. Developments in Sedimentology*, 1. Elsevier, Amsterdam, pp. 362-367.

Selley, R. C., 1988. *Applied Sedimentology*. Academic Press, London, 140 pp.

Selley, R. C., Shearman, D. J., Sutton J. & Watson, J., 1963. Some underwater disturbances in the Toridonian of Skye and Raasay. *Geol. Mag.*, 100: 224-243.

Sharp, R. P., 1942. Periglacial involutions in northeastern Illinois *J. Geol.*, 50: 274-301.

Shrock, R. R., 1948. *Sequence in layered rocks*. McGraw-Hill Book Company, New York, 507 pp.

Signiorini, R., 1936. Determinazione del senso di sedimentazione degli strati nelle formazioni arenacee del' Appenino settentrionale. *Bol. Soc. Geol. Ital.*, 55: 259-267.

Smith, B., 1916. Ball or pillow-form structures in sandstones. *Geol. Mag.*, 3: 234-240.

Sonnenfeld, P. & Hudec, P. P., 1985. Origin of clay films in rock salt. *Sediment. Geol.*, 44: 113-120.

Sorensen, T., 1935. Bodenformen und Planzendecke in Nordostgroenland, *Medd. om Grenland*, 93, (4): 69 pp.

Sparrow, E. M. & Husar, R. B., 1969. Longitudinal vortices in natural convection flow on inclined plates. *J. Fluid. Mech.*, 37: 251- -255.

Stauffer, P. H., 1967. Grain-flow dposits and their implications, Santa Ynez Mountains, California. *J. Sediment. Petrol.*, 37: 487-508.

Steusloff, U., 1950. Periglazialer Tropfen und Taschenboden. *Geol. Jb.*, 66: 305-312.

Steward, H. B., 1956. Contorted sediments in modern coastal lagoon explained by laboratory experiments. *Bull. Am. Assoc. Petrol. Geol.*, 40: 153-179.

Steward, A. D., 1963. On certain slump structures in the Torridonian sandstones of Applecross. *Geol. Mag.*, 100: 205-218.

Steward, H. B., 1956. Contorted sediments in modern coastal lagoons explained by laboratoty experiments. *Bull. Am. Assoc. Petrol. Geol.*, 40: 153-179.

Studer, B., 1827. Notice geognostique sur quelques parties de la chaine de Stockhorn et sur la houille du Siementhal, canton Berne. *Ann. Sci. Nat.*, 11: 1-20. Paris.

Sullwold, H. H., 1959. Nomenclature of load deformations in turbidites. *Geol. Soc. Am. Bull.*, 70: 1247-1248.

Sutton, J. & Watson, J., 1960. Sedimentary structures in the Epidotic grits. *Geol. Mag.*, 97: 106-132.

Sutton, R. G. & Lewis, T. L., 1966. Regional patterns of cross-laminae and convolutions in a single bed. *J. Sediment. Petrol.*, 36: 226-229.

Tanaka, K., 1970. Sedimentation of the Cretaceous Flysch sequence in the Ikushunbetsu area, Hokkaido, Japan. *Geol. Surv. Japan Report*, 236: 1-102.

Teisseyre, W., 1921. La tectonique compare des Subkarpathes. (In Polish, French summary). *Kosmos*: 242-474, Lwów.

Ten Haaf, E., 1956. Significance of convolute lamination. *Geol. en Mijnbouw*, 18: 188-194.

Ten Haaf, E., 1959. *Graded beds of the Northern Appenines*. Thesis. State Univ. of Groningen, Groningen: 102 pp.

Terzaghi, K., 1956. Varieties of submarine slope failures. *Proc. Texas Conf. Soil Mech. Found. Eng.*, 8: 41-52.

Thompson,. J. J. & Newal, H. F., 1885. On the formation of vortex rings by drops falling into liquids and some allied phenomena. *Proc. Roy. Soc. London*, 39: 417-437.

Thy, P. & Wilson, J.R., 1980. Primary igneous load-cast deformation structures in the Fongen-Hyllingen layered basic intrusion, Trondheim Region, Norway. *Geol. Mag.*, 117: 363-371.

Turner, F. J. & Weiss, L. E., 1963. *Structural Analysis of Metamorphic Tectonites*. McGraw Hill, New York, 545 pp.

Vanoni, V. A., 1946. Transportation of suspended sediment by water. *Trans. Amer. Soc. Civ. Eng.*, 111: 67-102.

Visher, G. S. & Cunnigham, R. D., 1981. Convolute lamination - A theoretical analysis: Example of a Pensylvanian sandstone. *Sediment. Geol.*, 28: 175-188.

Walker, R. G., 1966. Deep channels in-turbidite bearing formations. *Bull. Am. Assoc. Petrol. Geol.*, 50: 1899-1917.

Walker, R. G., 1980. Modern and ancient submarine fans: reply. *Bull. Am. Ass. Petrol. Geol.*, 62: 932-966.

Washburn, A. L., 1956. Classification of patterned ground and review of suggested origin. *Geol. Soc. Am. Bull.*, 67: 823-666.

Washburn, A. L., 1969. Weathering, frost action, and patterned ground in the Mesters Vig District, Northeast Greenland. *Meddelelser om Grřnland*, 176: 304 pp.

Wenthworth, C. M. Jr., 1967. Dish structures, a primary sedimentary structure. *Bull. Am. Assoc. Petr. Geol.*, 51: 485 (Abstract).

Whitaker, J. H. McD., 1973 . "Gutter casts", a new name for scour-and-fill structures with examples from the Llandoverian of Ringerike and Malmoya, Southern Norway. *Norsk. Geol. Tidsscrift*, 55: 403-417.

Williams, H. S., 1881. Channel fillings in Upper Devonian shales. *Am. J. Sci.*, 21: 318-320.

Williams, E., 1960. Intrastratal flow and convolute folding. *Geol. Mag.*, 97: 208-214.

Wilson, I. G., 1972. Aeolian bedforms -their development and origin. *Sedimentology*, 19: 172-210.

Wood, A. & Smith A. J., 1959. The sedimentation and sedimentary history of the Aberystwyth Grits (upper Llandloverian). *Quart. J. Geol. Soc.*, 114: 169-195.

Wooding, R. A., 1969. Growth of fingers at an unstable diffusing interface in a porous medium of Hell-Shaw cell. *J. Fluid. Mech.*, 39, 477-495.:

Woodworth, J. B., 1897. On the fracture system of joints, with remarks on certain great fractures. *Boston Soc. Nat. History Proc.*, 27: 163-184.

Wright, Ch. A., 1936. Experimental study of the scour of a sandy river by clear and muddy water. *J. Research Nat. Bureau Standarts, Washington D.C.*, 2: 193-206.

Wunderlich, F., 1967. Die Enstehung von "convolute bedding" an Plattenrändern. *Senck. Leth.*, 48: 345-349.

Streszczenie

Rozważania nad pochodzeniem struktur erozyjnych i deformacyjnych w ławicach osadowych

Stanisław Dżułyński

Rozważania, które są przedmiotem tej pracy, dotyczą struktur sedymentacyjnych w obrębie i na powierzchniach ławic skał osadowych, a częściowo również w pokrywach glebowych. Rozważania te są próbą synoptycznego ujęcia wyników badań doświadczalnych i terenowych, prowadzonych przed laty przez autora pracy i jego współpracowników. Wyniki tych badań, są rozproszone w różnych publikacjach, a struktury doświadczalne i naturalne były tylko wyrywkowo zestawiane ze sobą dla porównania. Pełniejszemu zestawieniu służy niniejsze opracowanie, a czas który upłynął od zakończenia badań pozwala na krytyczne spojrzenie na ich rezultaty.

Prawie wszystkie procesy, w następstwie których tworzyły się określone struktury, nakładały się na siebie i przechodziły jedne w drugie. Między strukturami, które w klasyfikacjach są często rozdzielone i traktowane odrębnie, istnieją formy przejściowe. Na spągu ławic, przejścia takie istnieją między śladami niesionych przedmiotów (zadziory uderzeniowe), jamkami wirowymi i strukturami odkształceniowymi. W zakresie hieroglifów prądowych, forma i pochodzenie tych struktur zależy, między innymi, od właściwości mechanicznych osadu dennego. W niespoistych osadach piaszczystych, w których poszczególne ziarna mają dużą swobodę ruchu, przepływ powoduje przemieszczanie ziarn. Powstają wówczas struktury erozyjno-akumulacyjne. W ciągliwych osadach ilastych, taki sam przepływ wywołuje głównie deformacje plastyczne. Na przykład, przepływy spiralne na podłożu piaszczystym powodują powstanie podłużnych grzbiecików rozdzielonych bruzdkami erozyjnymi, natomiast na podłożu ilastym w wyniku takich samych przepływów powstają grzbieciki i bruzdy deformacyjne. Biorąc pod uwagę wyłącznie morfologię, dyskusja czy są to struktury erozyjno-akumulacyjne czy odkształceniowe, staje się bezprzedmiotowa.

Szczególny nacisk położono w pracy na mniej lub bardziej regularne deformacje w dwudzielnych układach niestatecznych, w których nałożone na siebie warstwy są statystycznie jednorodne i jednakowej miąższości oraz odkształcają się plastycznie.

Najczęściej, niestateczność powstaje w następstwie nałożenia warstwy cięższej na lżejszą. Tego rodzaju układy są układami spustowymi i zawierają pewien zasób energii potencjalnej, która w określonych warunkach może być przetworzona w pracę potrzebną do deformacji. Formy odkształceń będą tu analogiczne do tych, które powstają w podgrzewanych płynach gdzie niestateczność gęstościowa jest wywołana różnicami temperatury. Podobieństwo morfologiczne jest w tym przypadku uwarunkowane podobieństwem procesów formotwórczych. W procesach tych istotną rolę odgrywa: (1) lepkośc kinematyczna, która jest miarą ruchliwości ośrodków podlegających odkształceniom płynnym oraz, (2) brak lub obecność ruchu poziomego między warstwami o różnej gęstości. Przy braku horyzontalnego ruchu między warstwami, układ niestateczny nazywamy układem horyzontalnie nieruchomym. Jeśli natomiast ruchom pionowym towarzyszy przemieszczenie poziome, niestateczne układy nazywami układami poziomo ruchomymi.

W układach poziomo nieruchomych, proces deformacyjny upodabnia się do komórkowych zawirowań konwekcyjnych. Materiał o mniejszej lepkości kinematycznej wciska się w ośrodek o większej lepkości pod postacią słupowych "diapirów", które przy gęstym rozmieszczeniu stykają się wzajemnie ścianami. W początkowych stadiach rozwojowych, "diapiry" przybierają formy wielobocznych wypukłości lub wklęsłości.

W podwodnych środowiskach morskich lub jeziornych opadająca na dno zawiesina piaszczysta posiada stosunkowo małą lepkość. Przy braku poziomego przemieszczenia zawiesina taka grzęźnie w bardziej spoistym mule pod postacią poligonalnych lub owalnych pogrązów. W tym przypadku jest to deformacja syndepozycyjna.

Analogiczne struktury powstają również jako struktury postdepozycyjne. Bodźcem wyzwalającym deformację może tu być n.p. wstrząs sejsmiczny który spowoduje upłynnienie lub uplastycznienie osadów. Zarówno syndepozycyjne jak i postdepozycyjne struktury są morfologicznie do siebie podobne lub takie same.

W środowiskach kontynentalnych zaburzenie niestabilnej równowagi może nastąpić przez nawodnienie warstwy ilastej podścielającej ławicę żwirową lub piaszczystą. W strefie półsuchych klimatów ma to miejsce podczas gwałtownej ulewy kiedy wezbrane potoki naniosą żwir lub piasek na wyschnięty muł. Deformacja rozpocznie się z chwilą, gdy wsiąkająca w muł woda spowoduje jego rozmiękczenie. Dochodząc do powierzchni naniesionej ławicy, wyciskany osad ilasty o stosunkowo małej lepkości kinematycznej będzie rozpychał na boki gruboziarnisty ma-teriał dając początek glebom poligonalnym. W wilgotnych środowiskach "peryglacjalnych" dwudzielność i niestateczność po-kryw glebowych jest wynikiem mrozowego wymarzania lub osa- dzenia przez roztopowe wody warstwy gruzowej na zamarzniętym drobnoziarnistym lub ilastym podłożu. Bodźcem, który w takich warunkach może spowodować deformację, będzie odwilż. W tym rzypadku również może dojść do utworzenia się gleb poligonalnych, które same w sobie nie są jednak wskaźnikiem chłodnego klimatu i procesów mrozowych. Nim dojdzie do przebicia warstwy gruzowej, zdeformowana powierzchnia która oddziela warstwy o różnej gęstości i lepkości kinematycznej jest odwróconym odbiciem zwierciadlanym pogrązów na spągu osadów prądowych. Biorąc pod uwagę procesy formotwórcze, pogrązy na spągu piaskowców i gleby poligonalne są strukturami pokrewnymi. Analogia taka obejmuje również śródławicowe konwolucje powstałe wyłącznie w następstwie pionowych przemieszczeń grawitacyjnych.

W niestatecznych układach ruchomych miejsce zawirowań komórkowych zajmują zawirowania spiralne. Jeśli przemieszczenie poziome jest stosunkowe nieznaczne to syndepozycyjne deformacje na spągu piaskowców mają formy łuskowe. Jeżeli natomiast ruch poziomy przeważa to deformacje mają formę podłużnych grzbietów.

Analogiczne odkształcenia pojawiają się także w obrębie ławic jako struktury syndepozycyjne w stosunku do ławicy jako całości, to znaczy przed jej ostatecznym uformowaniem. Do tego rodzaju struktur należą podłużne konwolucje w postaci śródławicowych garbów o kierunku równoległym do kierunku przepływu. Podobne struktury tworzą się również w pokrywach glebowych na skłonach stoków (gleby pasowe) i w niezastygłych magmach. W tym ostatnim przypadku ma to miejsce na styku dwu magm, różniących się gęstością i lepkością kinematyczną, z których cięższa intruduje na lżejszą. Przy braku ruchu między magmami, deformacje na ich styku są analogiczne do wielobocznych pogrązów na spągu ławic skał osadowych.

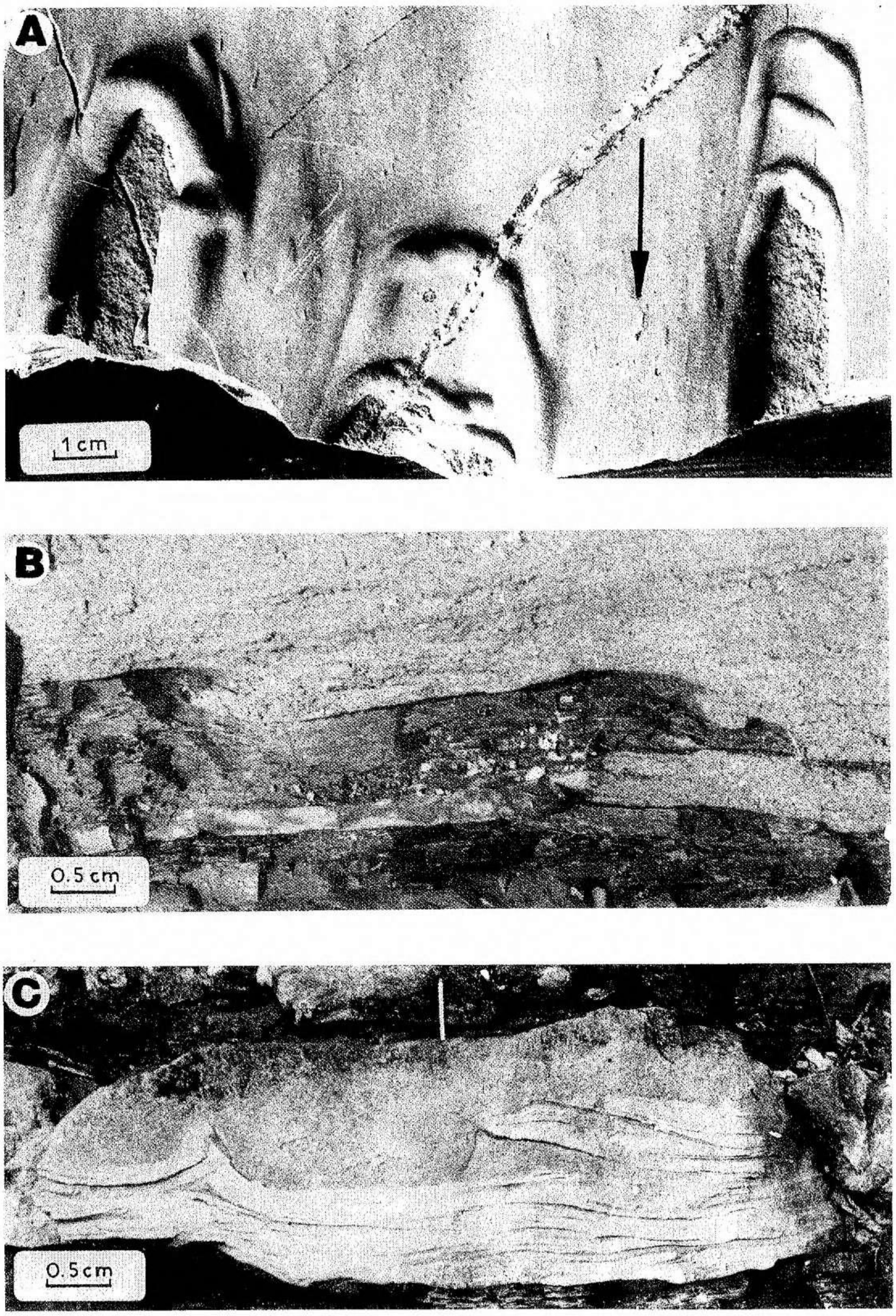


Plate 1 (**A**) Moulds of crescentic scour marks produced at upcurrent side of resistant patches of lithified shaly intercalation. Successive crescents indicate intermittent retreat of undercut edges of resistant patches. Oligocene Krosno Beds, Wernejowka. Comp. Fig. 1. (**B**) Undercutting of cohesive bottom mud by strong t-current (sheet flow). Direction of flow is away from observer. Oligocene Magura Beds, Ropica Górna. (**C**) Incipient uplift of semi-consolidated bottom laminae by downsinking heavy suspension. Detail of amalgamated bed. Direction of flow away from observer. Oligocene Magura Beds, Ropica Górna. Comp. Fig. 35

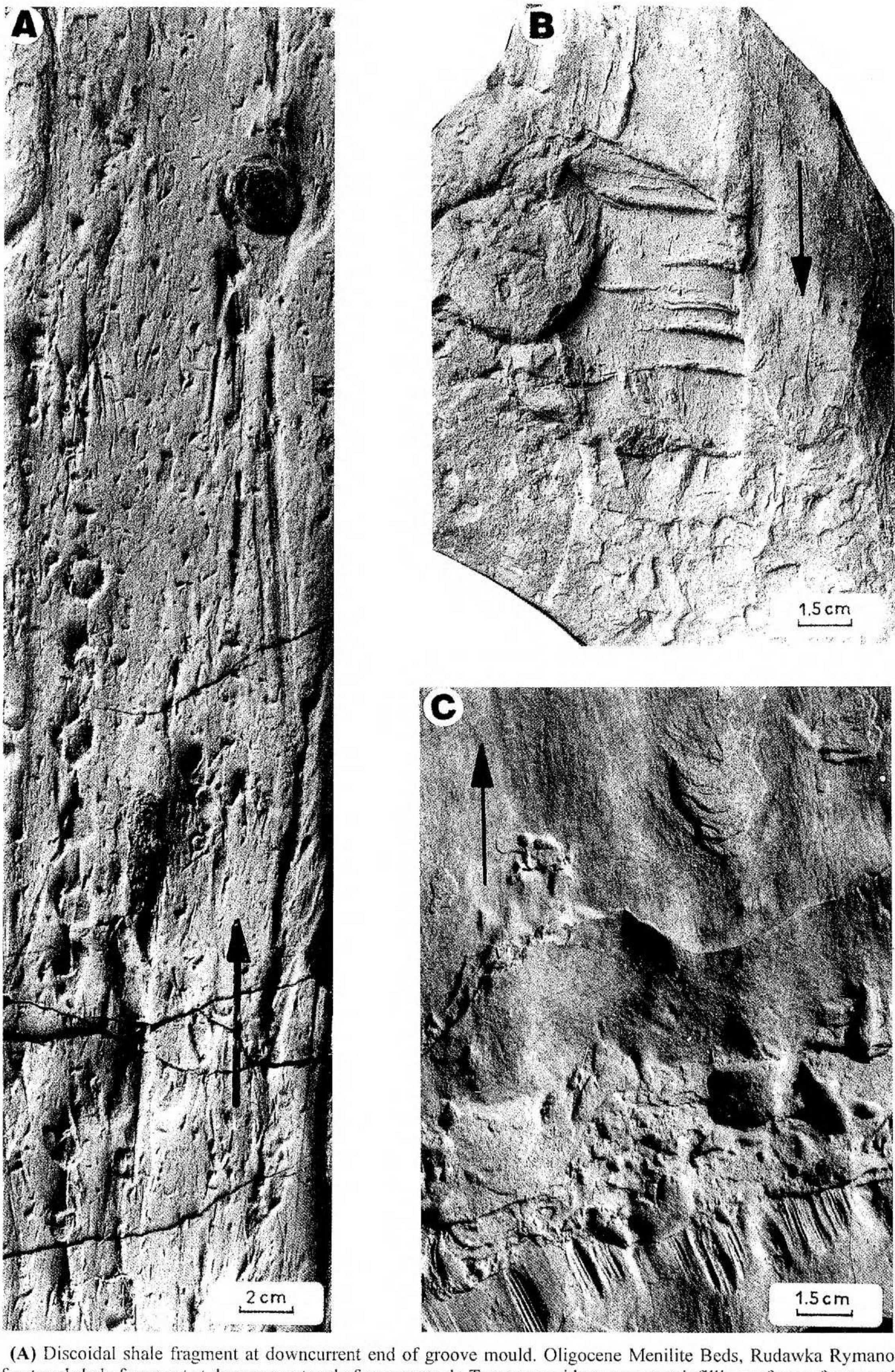


Plate 2 (**A**) Discoidal shale fragment at downcurrent end of groove mould. Oligocene Menilite Beds, Rudawka Rymanowska. (**B**) Mould of fractured shale fragment at downcurrent end of groove mark. Transverse ridges represent infillings of open fractures by settling sand. Oligocene Podhale Flysch, Witów. (**C**) Mould of brush mark produced by impact of large shale fragment with moulds of smaller pieces of shale detached from margins of impinging fragment on collision with bottom. Oligocene Krosno Beds, Besko

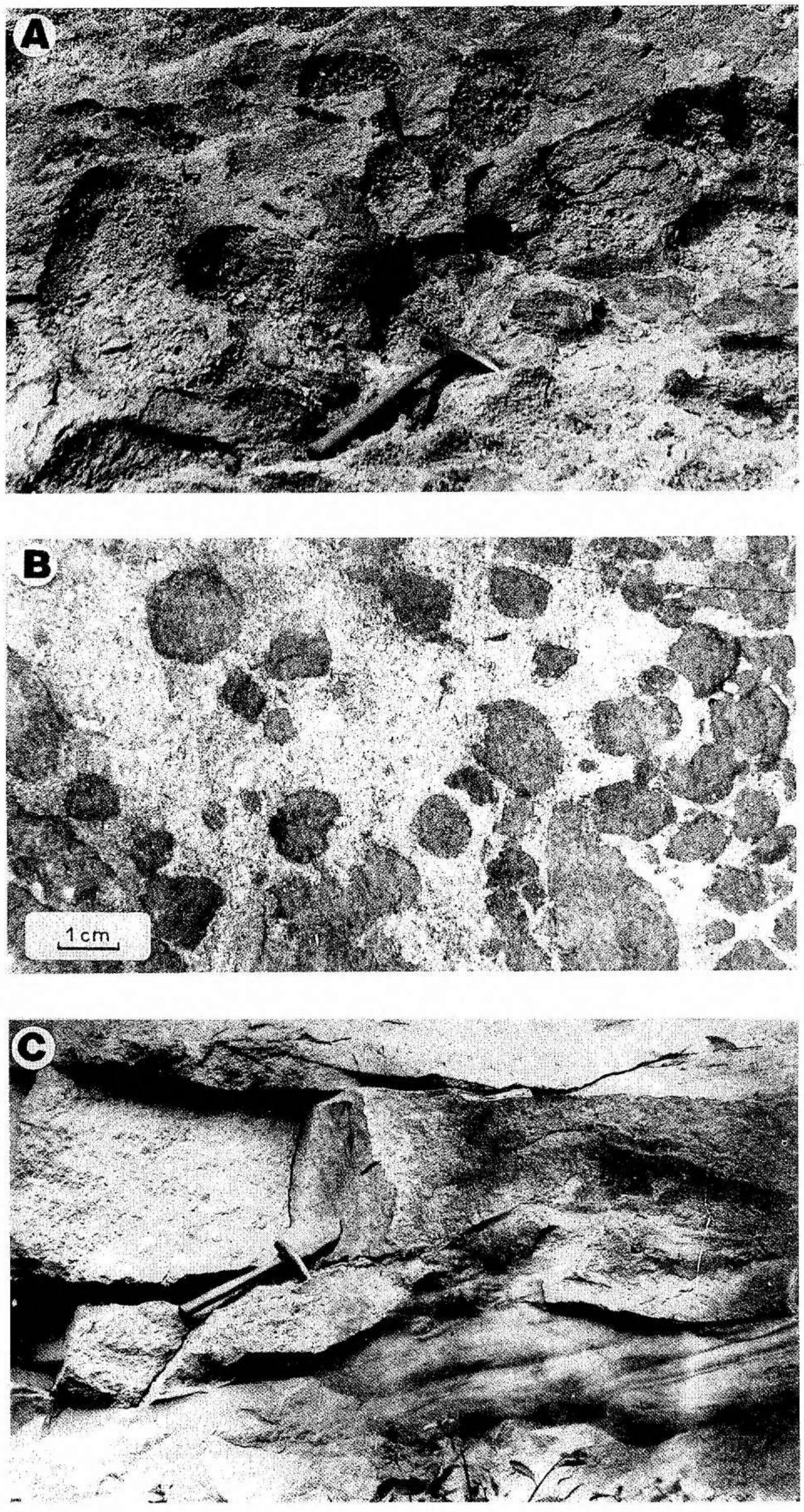


Plate 3 (**A**) Armored mud balls in fluxoturbidite. Eocene Ciężkowice Beds, Wola Komborska. (**B**) Discoidal shale fragments on intrastratal parting surface of sandstone. Oligocene Menilite Beds, Rudawka Rymanowska. (**C**) Scratches and grooves produced by soft mass of sand gliding upon sandy substratum. Detail of amalgamated fluxoturbite bed. Upper Cretaceous Istebna beds, Rożnów

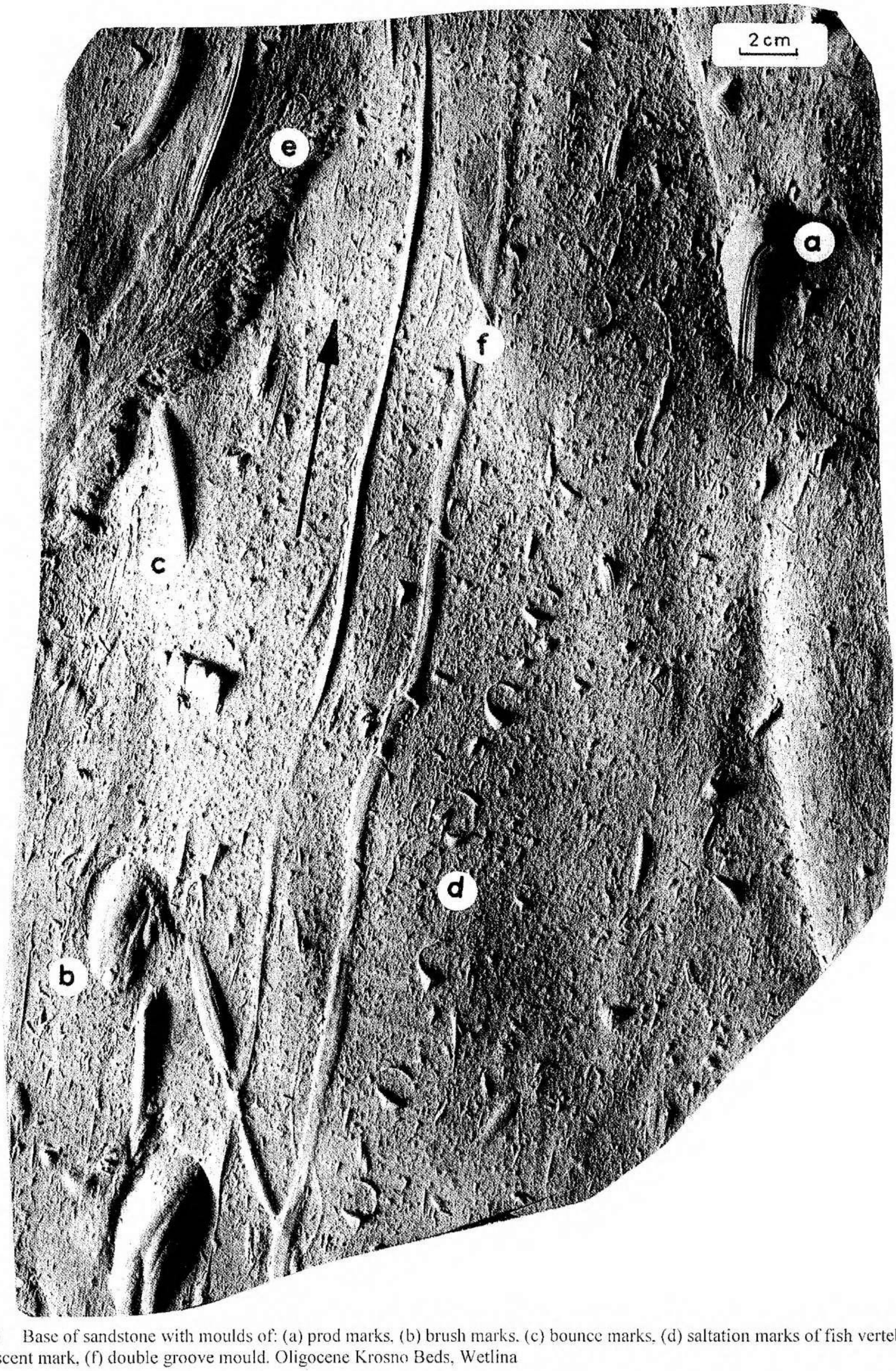


Plate 4 Base of sandstone with moulds of: (a) prod marks, (b) brush marks, (c) bounce marks, (d) saltation marks of fish vertebrae, (e) frondescent mark, (f) double groove mould. Oligocene Krosno Beds, Wetlina

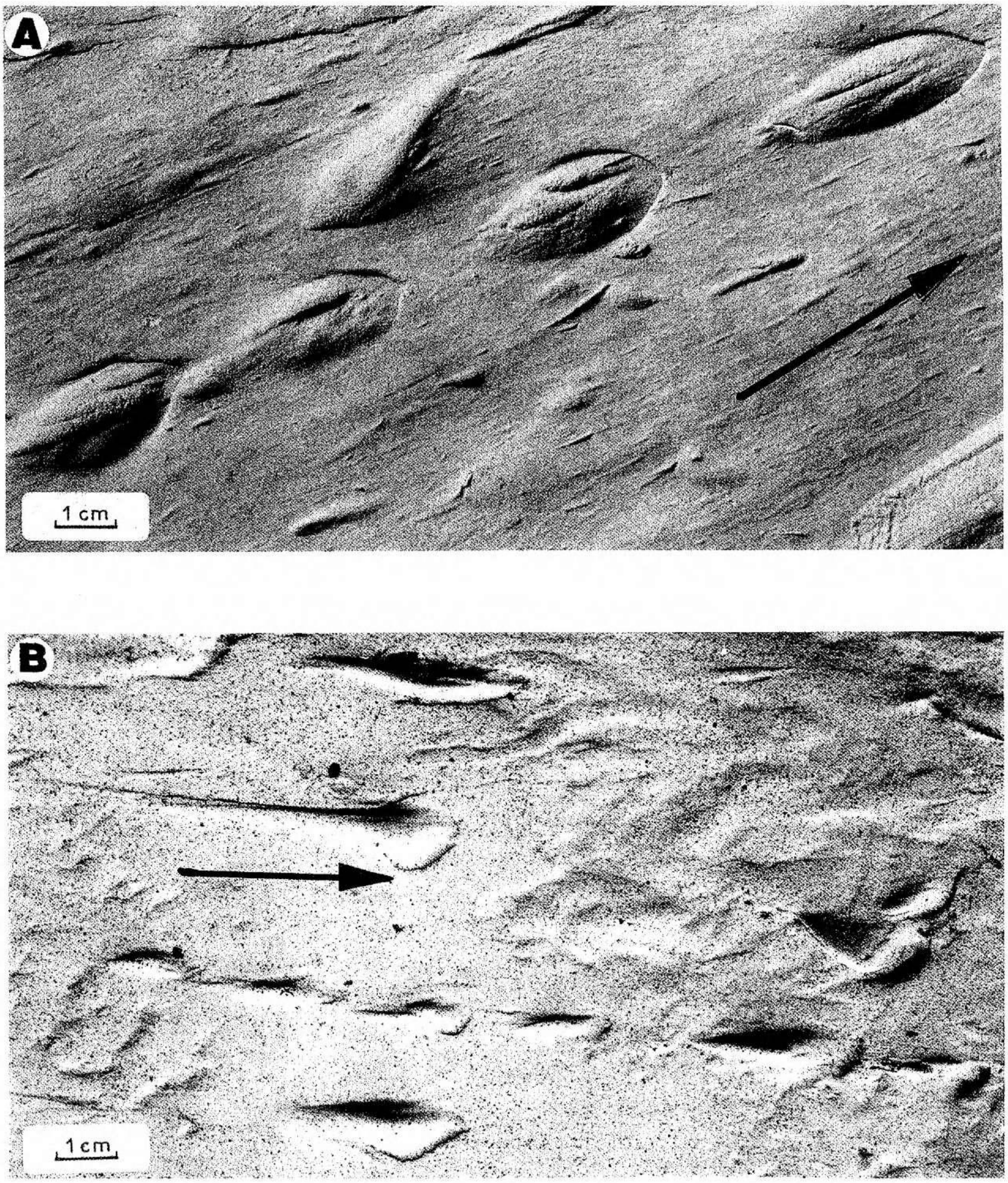


Plate 5 (**A**) Skip casts. Oligocene Krosno Beds, Kąty. (**B**) Experimental skip casts. Base of artificial turbidite

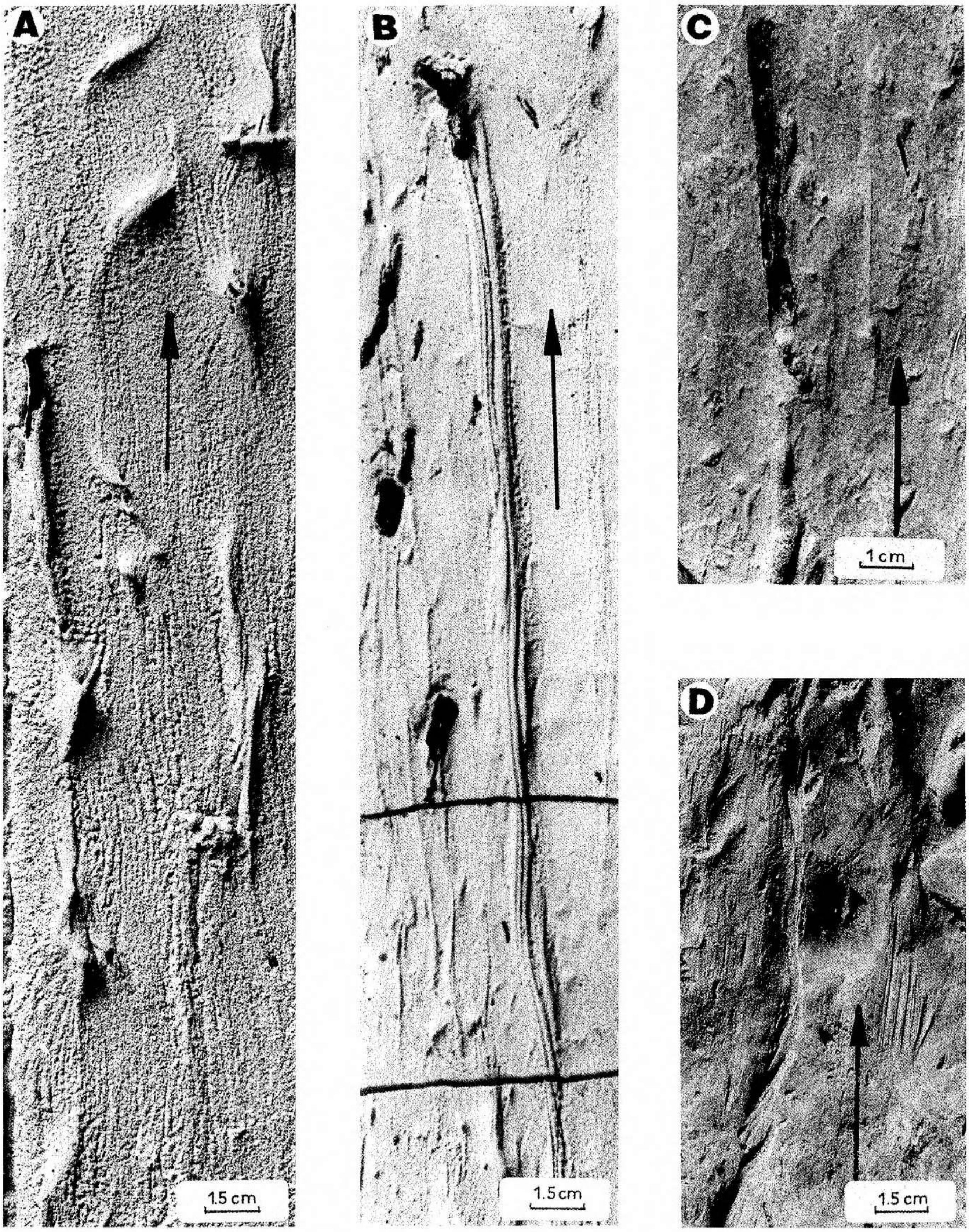


Plate 6 **(A)** Mould of groove produced by dragged lignite fragment, prod markings and longitudinal ridges. Base of artificial turbidite. **(B)** Twisted groove mould produced by dragged lignite. Base of artificial turbidite (after Dżułyński, 1965). **(C)** Piece of wood at downcurrent end of groove mould. Oligocene Menilite Beds, Rudawka Rymanowska. **(D)** Twisted groove mould . Oligocene Menilite Beds, Rudawka Rymanowska

Plate 7 **(B)** Drag Mark. Upper Mancos Shale, Cretaceous, Black Mesa, California, USA. **(B)** Experimental drag marks. Base of artificial turbidite (after Dzulynski & Simpson, 1966)

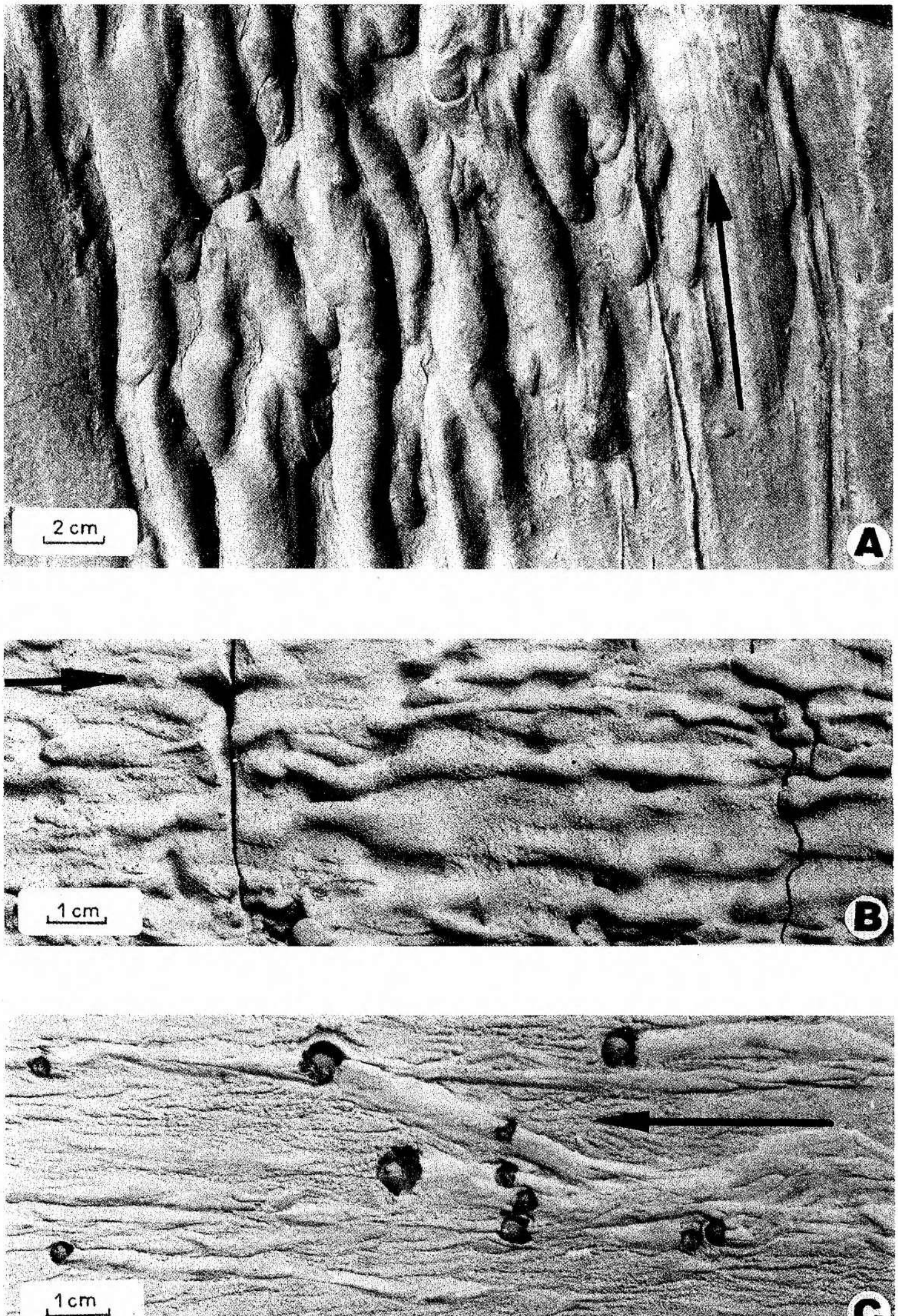


Plate 8 (**A**) Rounded groove moulds. Oligocene Krosno Beds, Rudawka Rymanowska. (**B**) Moulds of rounded grooves produced by rolling clay spheres upon soft substrate. Base of artificial turbidite. (**C**) Groove moulds produced by clay spheres rolling upon thin layer of soft bottom mud underlain by more cohesive clay. Base of artificial turbidite

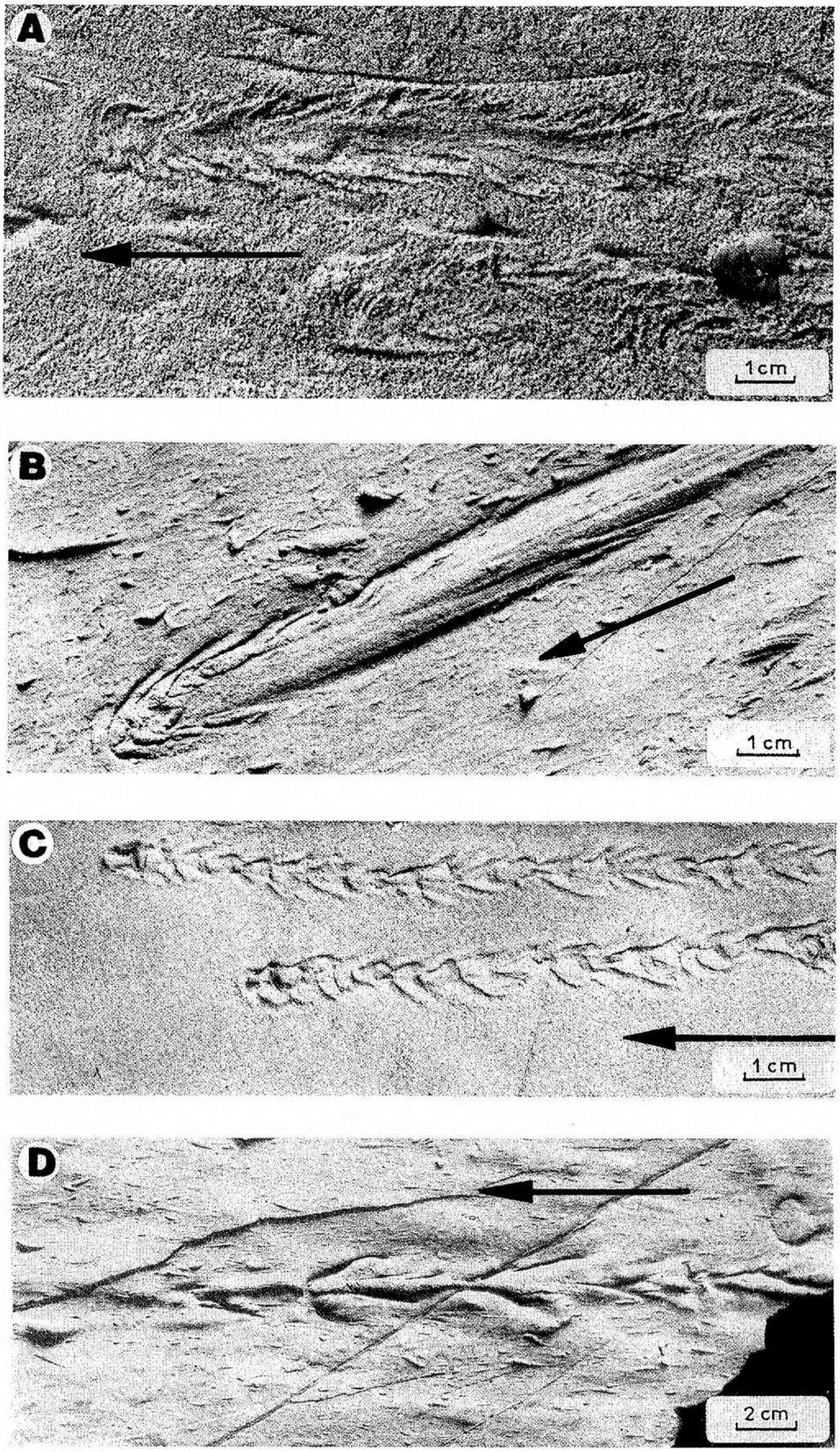


Plate 9 (**A**) Doubly ruffled groove. Base of artificial turbidite (after Dzulynski & Simpson, 1966). (**B**) Partly double ruffled groove. Oligocene Krosno Beds, Rudawka Rymanowska. (**C**) Chevron mark produced by skiming fish vertebrae. Base of artificial turbidite. (**D**) Chevron mark produced by skiming fish vertebra. Oligocene Krosno Beds, Kąty

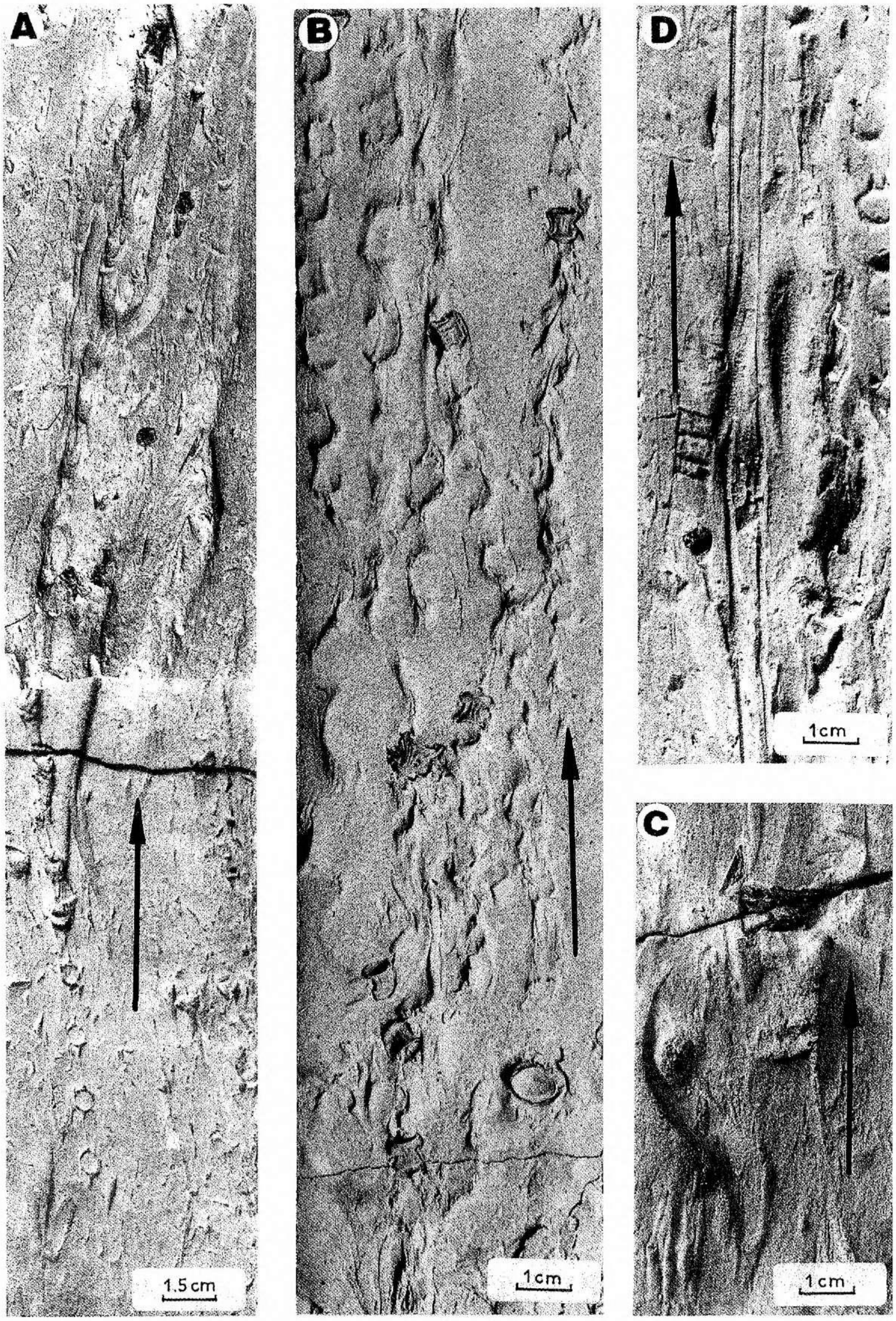


Plate 10 **(A)** Markings produced by somersaulting and saltating fish vertebra. Note fish vertebra at downcurrent end of markings. Oligocene Menilite Beds, Rudawka Rymanowska. **(B)** Markings produced by somersaulting and saltating fish vertebrae. Base of artificial turbidite (after Dżułynski, 1965). **(C)**, **(D)** markings produced by rolling fish vertebrae. Note fish vertebra preserved at downcurrent end of marking C. Oligocene Menilite Beds, Rudawka Rymanowska

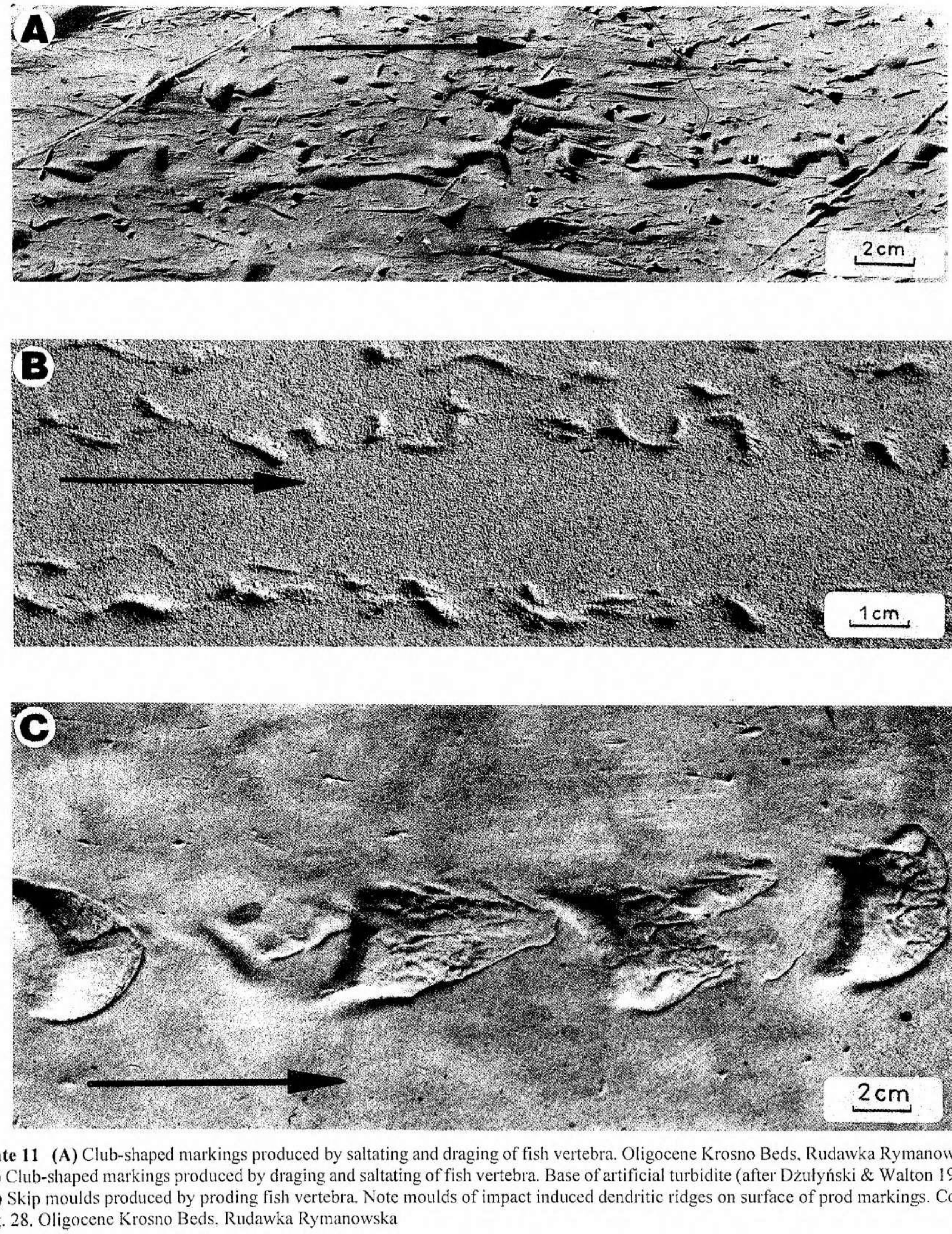


Plate 11 **(A)** Club-shaped markings produced by saltating and draging of fish vertebra. Oligocene Krosno Beds. Rudawka Rymanowska. **(B)** Club-shaped markings produced by draging and saltating of fish vertebra. Base of artificial turbidite (after Dżułyński & Walton 1965). **(C)** Skip moulds produced by proding fish vertebra. Note moulds of impact induced dendritic ridges on surface of prod markings. Comp. Fig. 28. Oligocene Krosno Beds. Rudawka Rymanowska

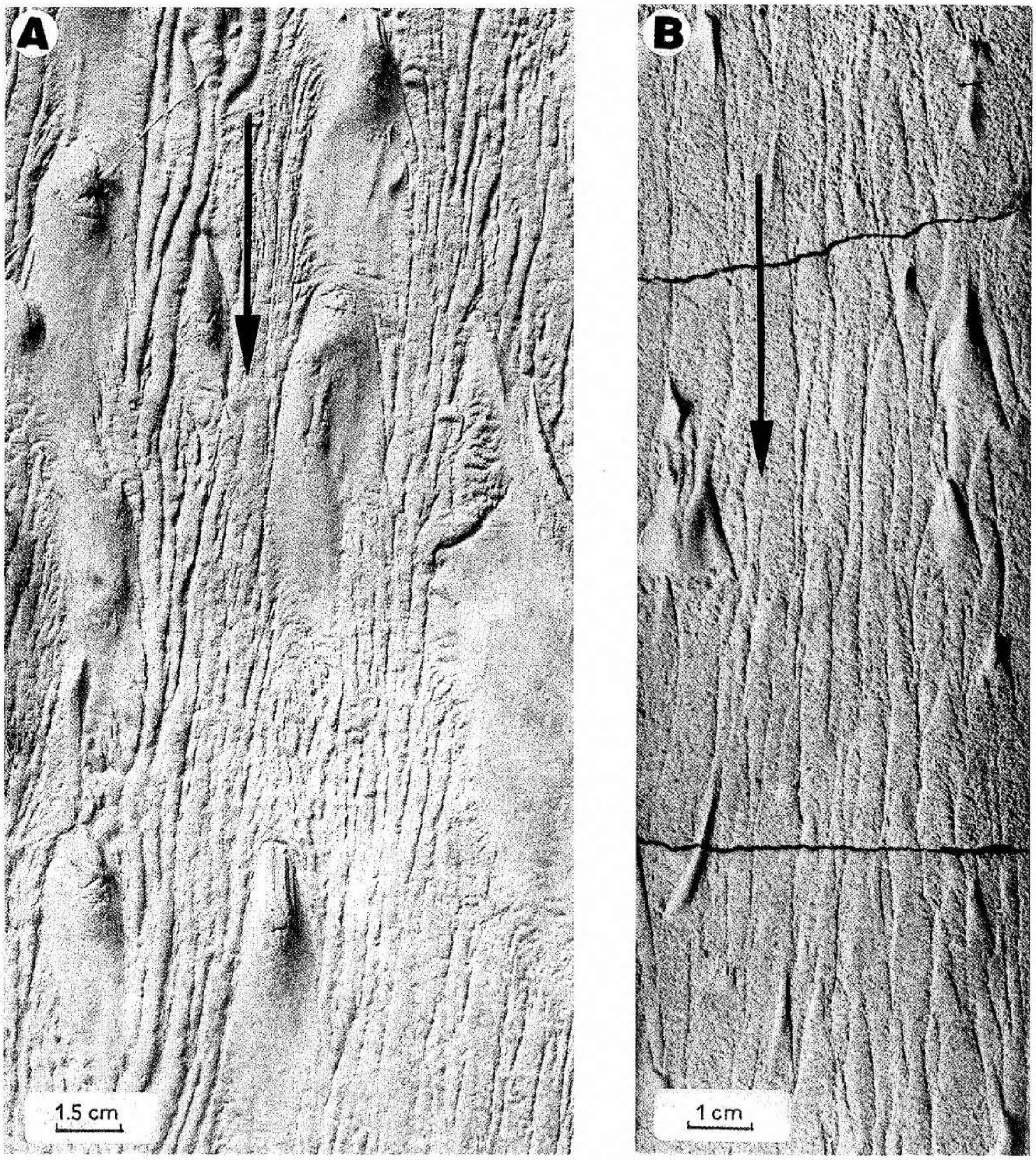


Plate 12 **(A)** Flute moulds at downcurrent ends of prod marks and moulds of l-ridges. Oligocene Krosno Beds, Rudawka Rymanowska. **(B)** Flute moulds at downcurrent ends of prod marks and moulds of l-ridges. Base of artificial turbidite

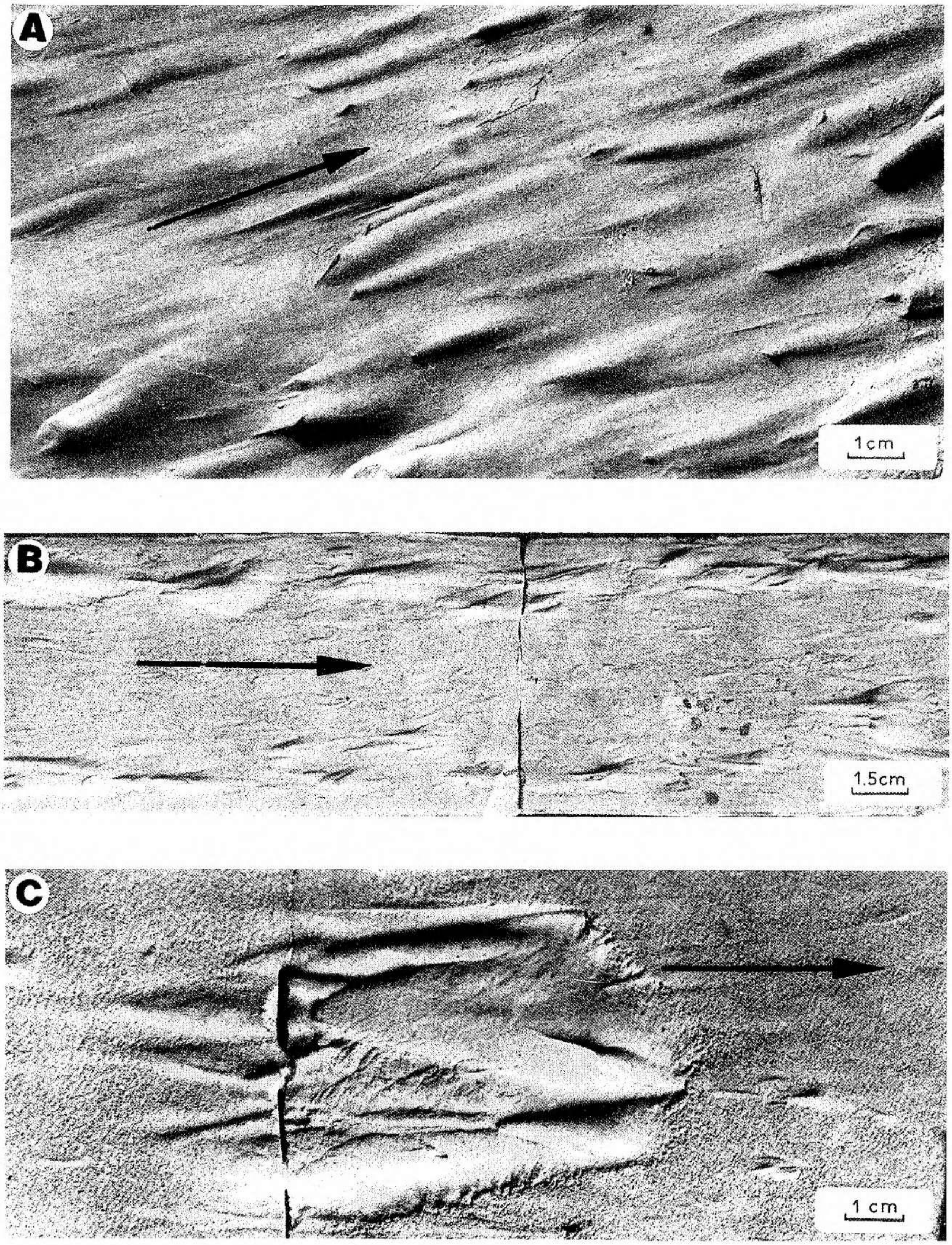


Plate 13 **(A)** Flute markings at downcurrent ends of prod markings. Oligocene Krosno Beds, Rudawka Rymanowska. **(B)** Flute moulds at downcurrent ends of prod markings. Base of artificial turbidite (after Dzulynski & Simpson, 1966). **(C)** Relatively large flute moulds at downcurrent ends of prod marks. Note prod mark with smaller flute superimposed upon large flute. Base of artificial turbidite (after Dzulynski & Simpson, 1966)

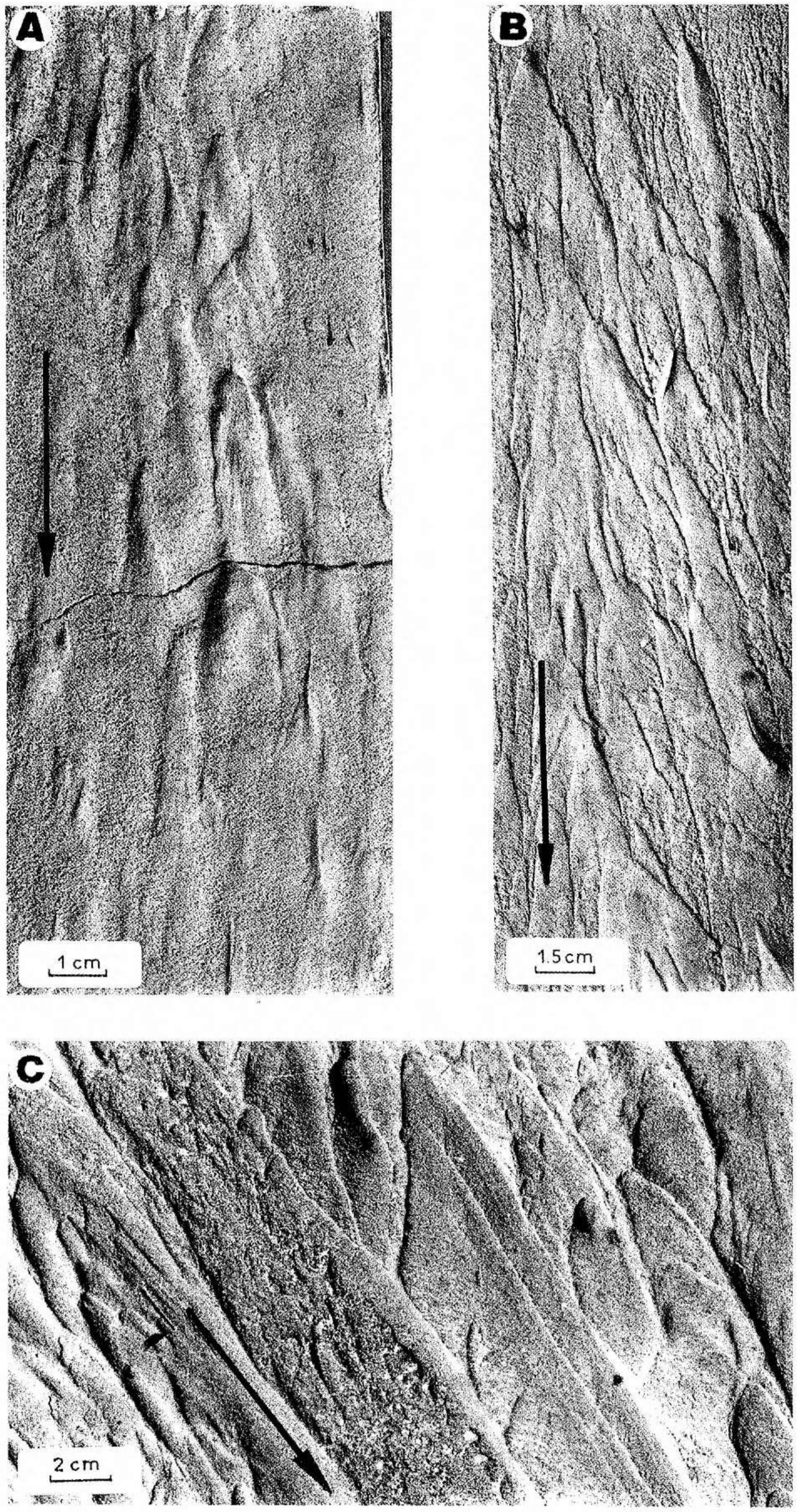
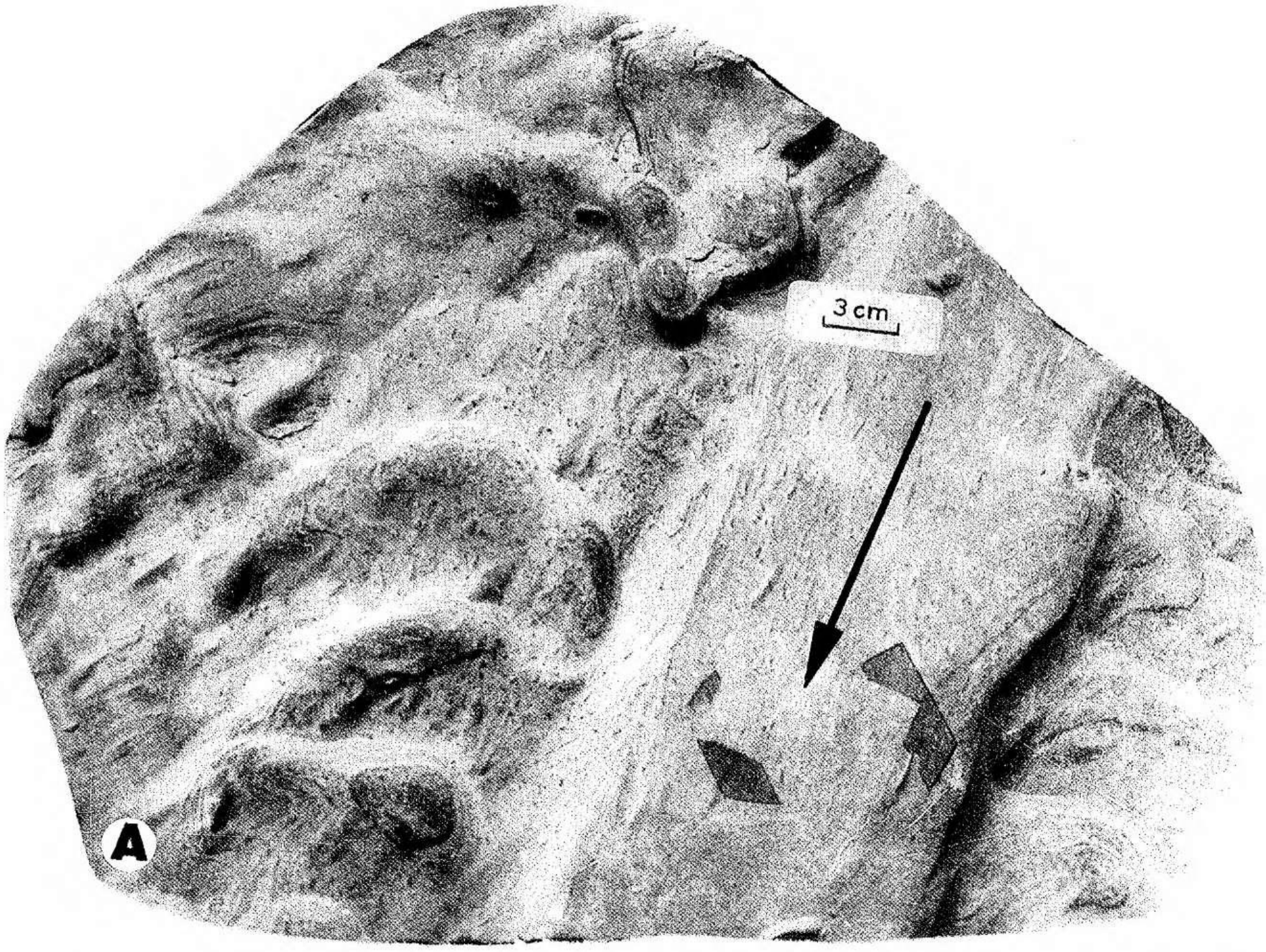


Plate 14 (**A**) Bulbous and sharp-pointed flute moulds. Base of artificial turbidite (after Dzulynski & Simpson, 1966). (**B**) Flat-bottomed flute moulds. Base of artificial turbidite (after Dzulynski & Simpson 1966). (**C**) Bulbous (left) and flat (right) flute moulds. Oligocene Krosno Beds, Rudawka Rymanowska



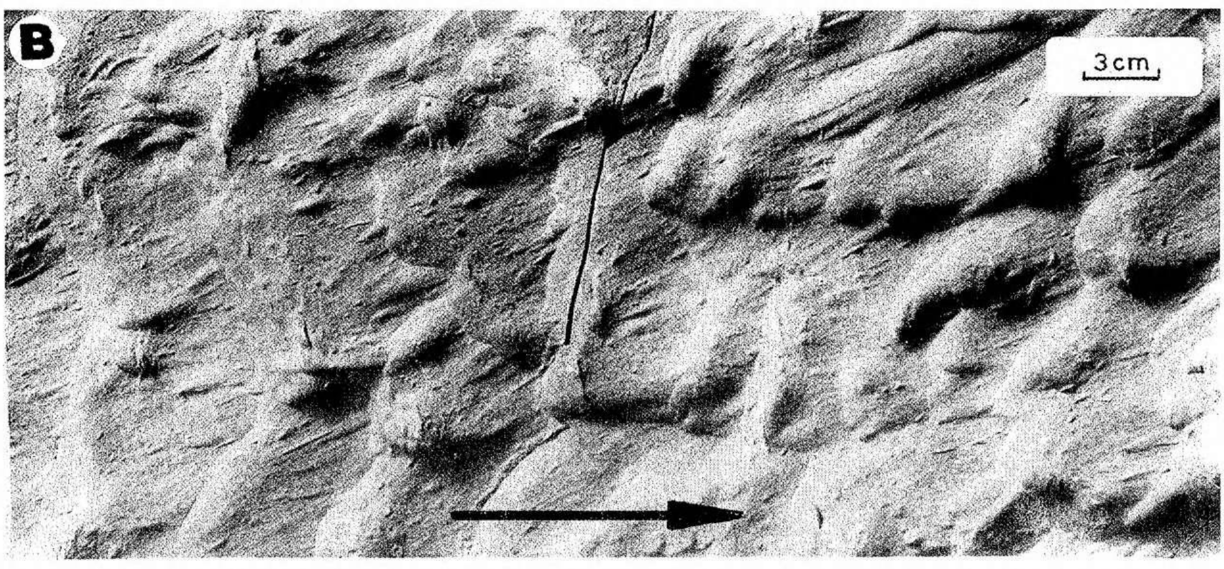
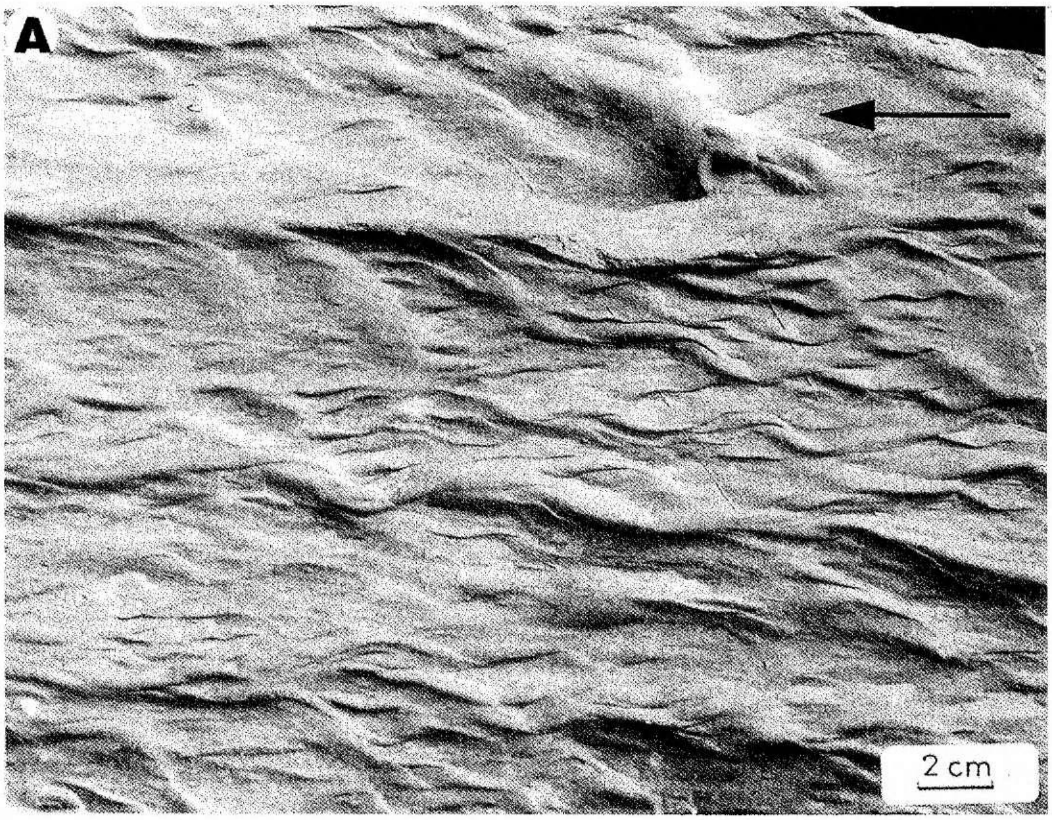


Plate 15 **(A)** Irregular transverse scour marks developed upon horizontally laminated substrate. Note parallel contour lines on walls of moulds recording horizontal silty laminae in sub-bottom sediment. Oligocene Krosno beds, Rudawka Rymanowska. **(B)** Moulds of transverse scour markings passing into elongated flute moulds. Oligocene Krosno Beds, Wernejówka



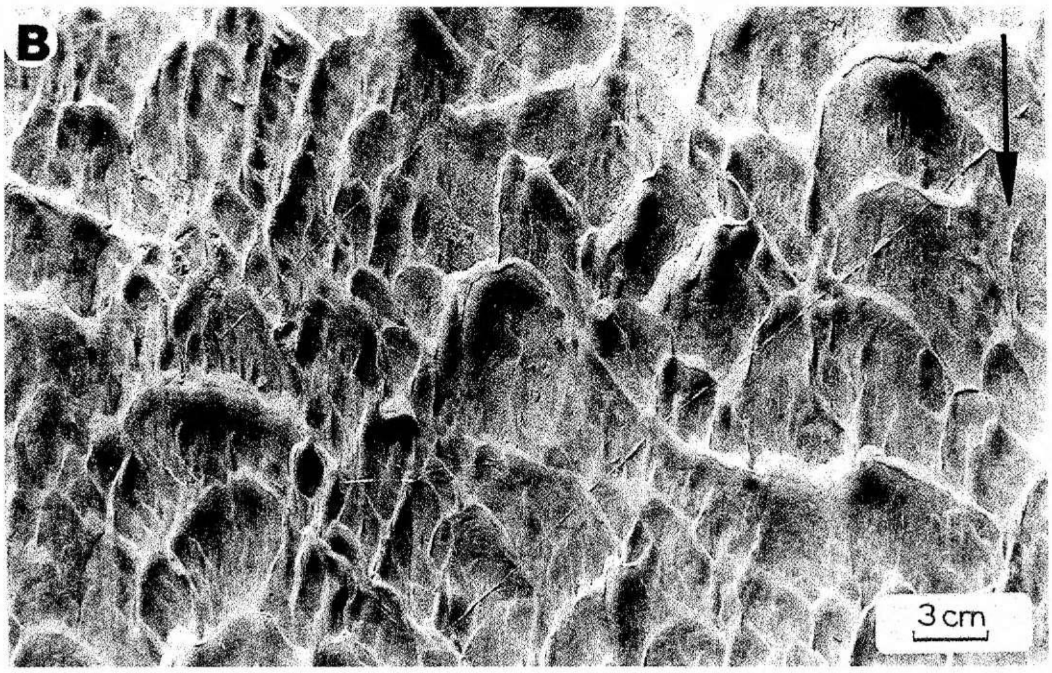


Plate 16 (**A**) Slightly meandering moulds of rill markings. Oligocene Krosno Beds, Wernejówka. (**B**) Rill markings arranged in diagonal pattern. Oligocene Krosno Beds, Wernejówka

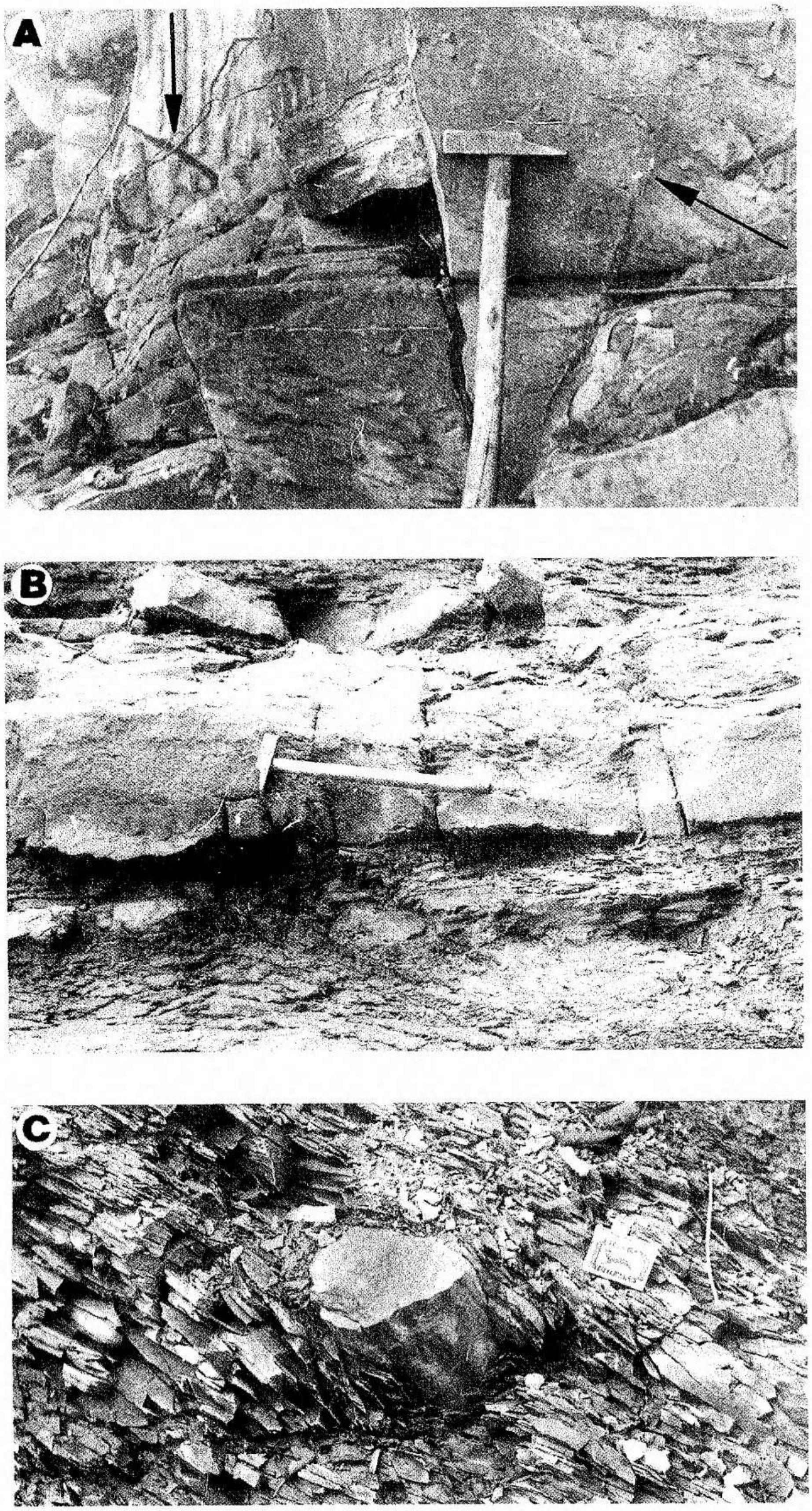


Plate 17 **(A)** Changing directions of palecurrent markings in succesive flysch beds. Oligocene Krosno Beds, Wernejówka. **(B)** Amalgamated turbidit bed. Remnants of light coloured lower layer, deposited by current coming from NNW, overlain by coarse grey layer laid down by current coming from SE. Oligocene Krosno Beds, Sieniawa (after Dżułyński & Ślączka, 1958). **(C)** Isolated erosion channel (gutter mark) in shales. Direction of flow away from observer. Oligocene Magura Beds, Dragaszów

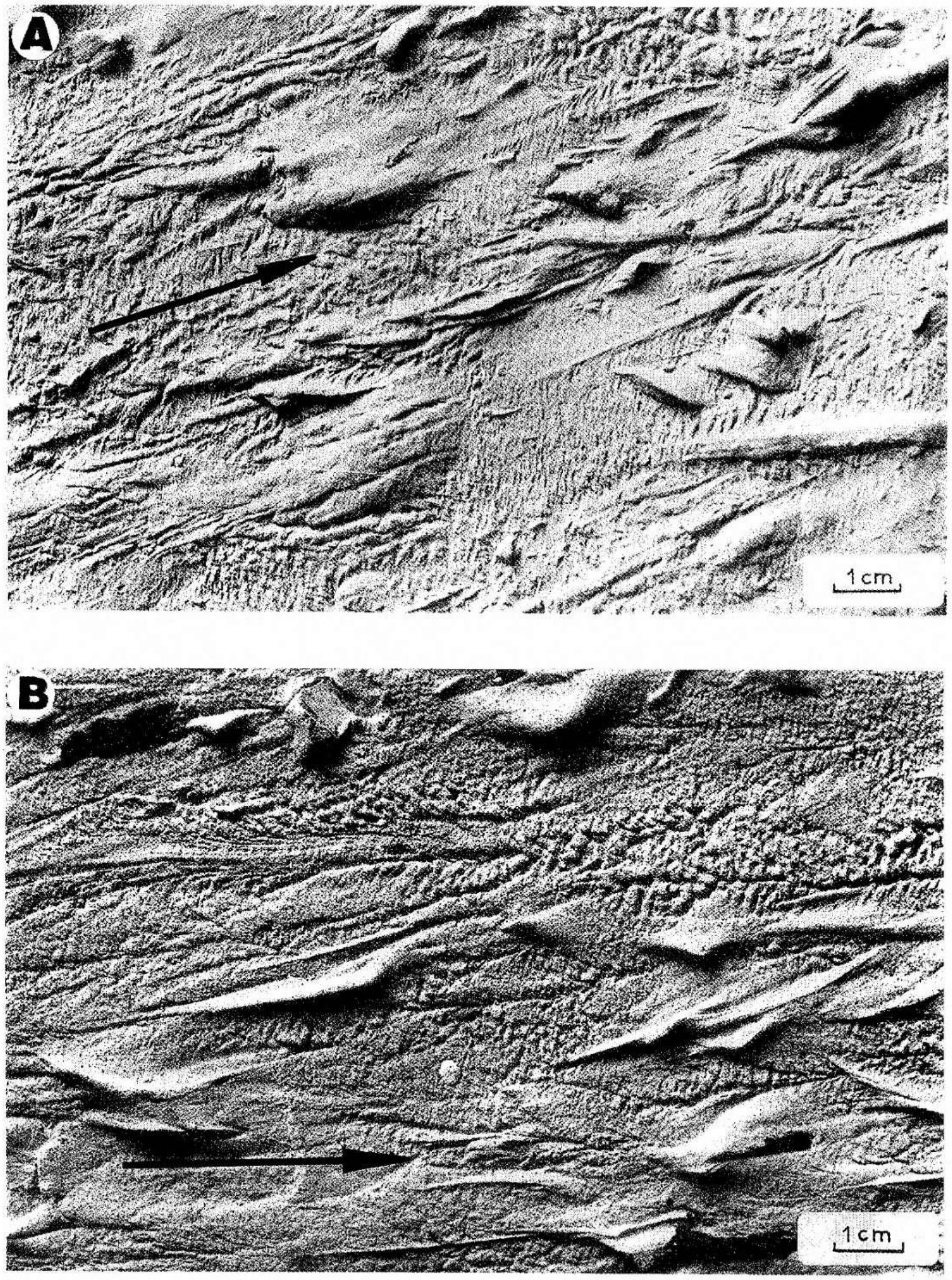


Plate 18 (**A**) Assemblage of scour, prod and deformational markings. Oligocene Krosno Beds, Rudawka Rymanowska. (**B**) Analogous assemblage of markings as in (**A**). Base of artificial turbidite

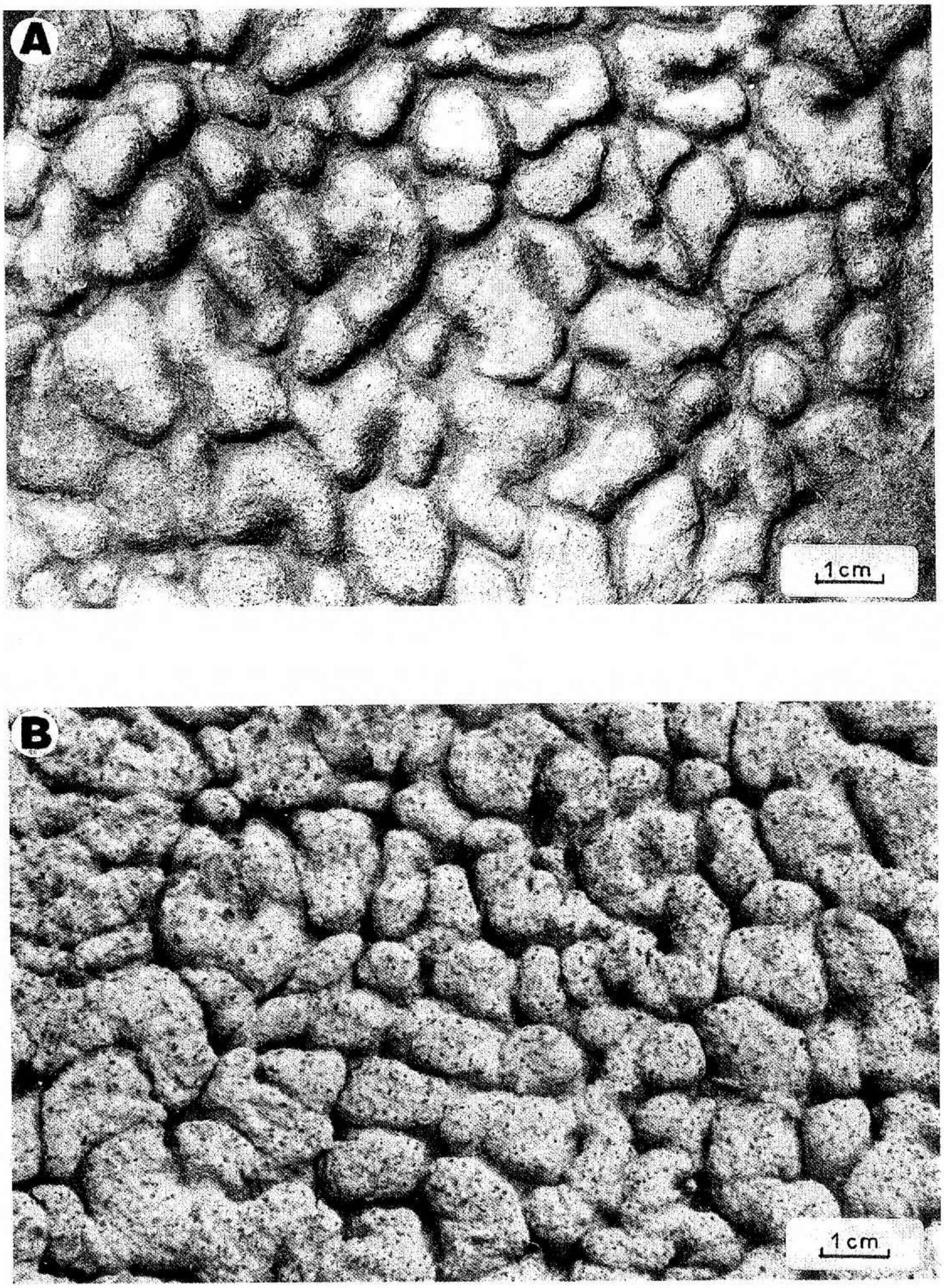


Plate 19 **(A)** Polygonal load casts. Oligocene Krosno Beds, Rudawka Rymanowska. **(B)** Polygonal load casts. Base of artificial turbidite (after Dzulynski & Simpson, 1966)

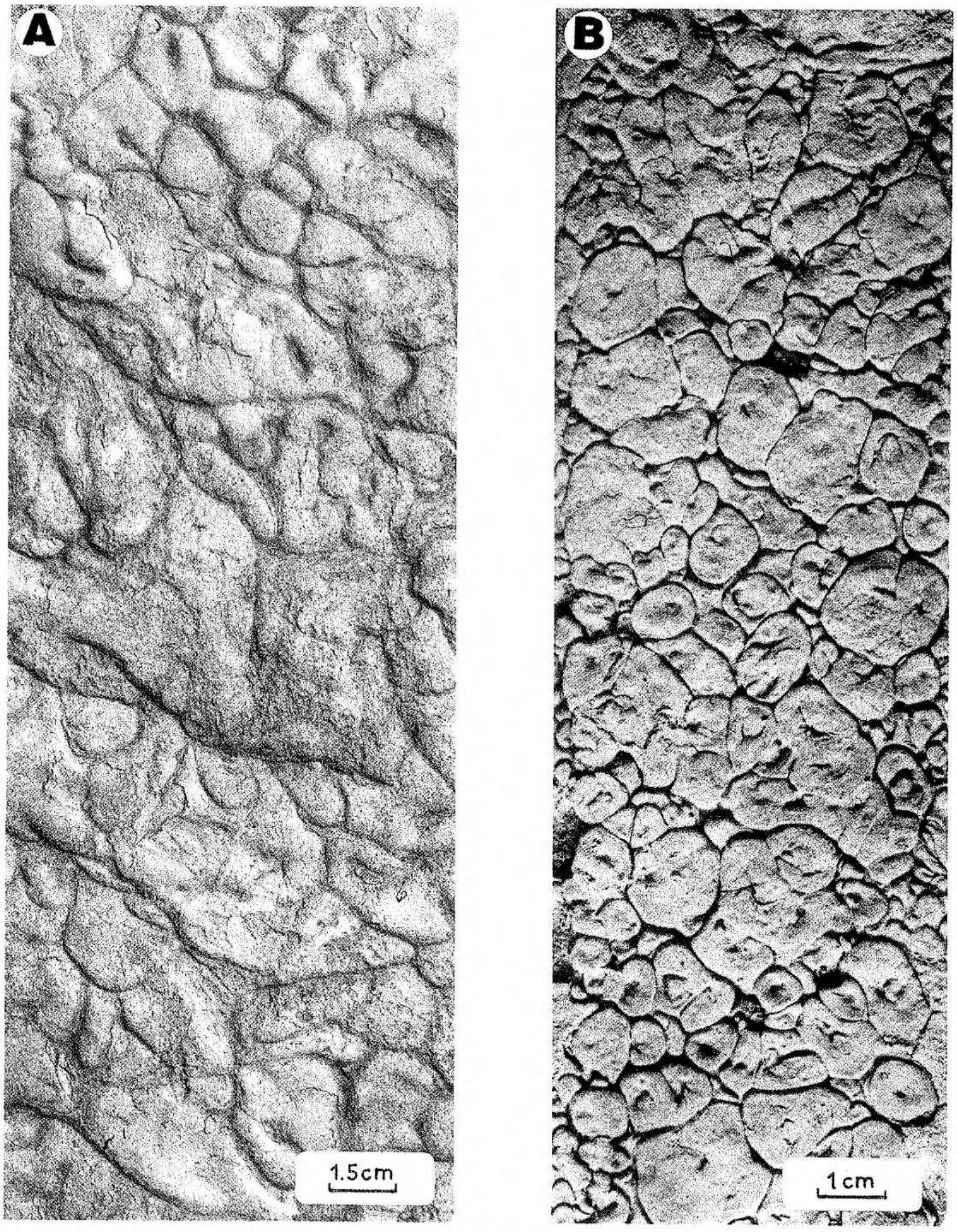


Plate 20 **(A)** Rounded load-casts. Note moulds of point injections of clay in centres of some load casts. Eocene Magura Beds, Koninki. **(B)** Rounded load casts. Base of artificial turbidite

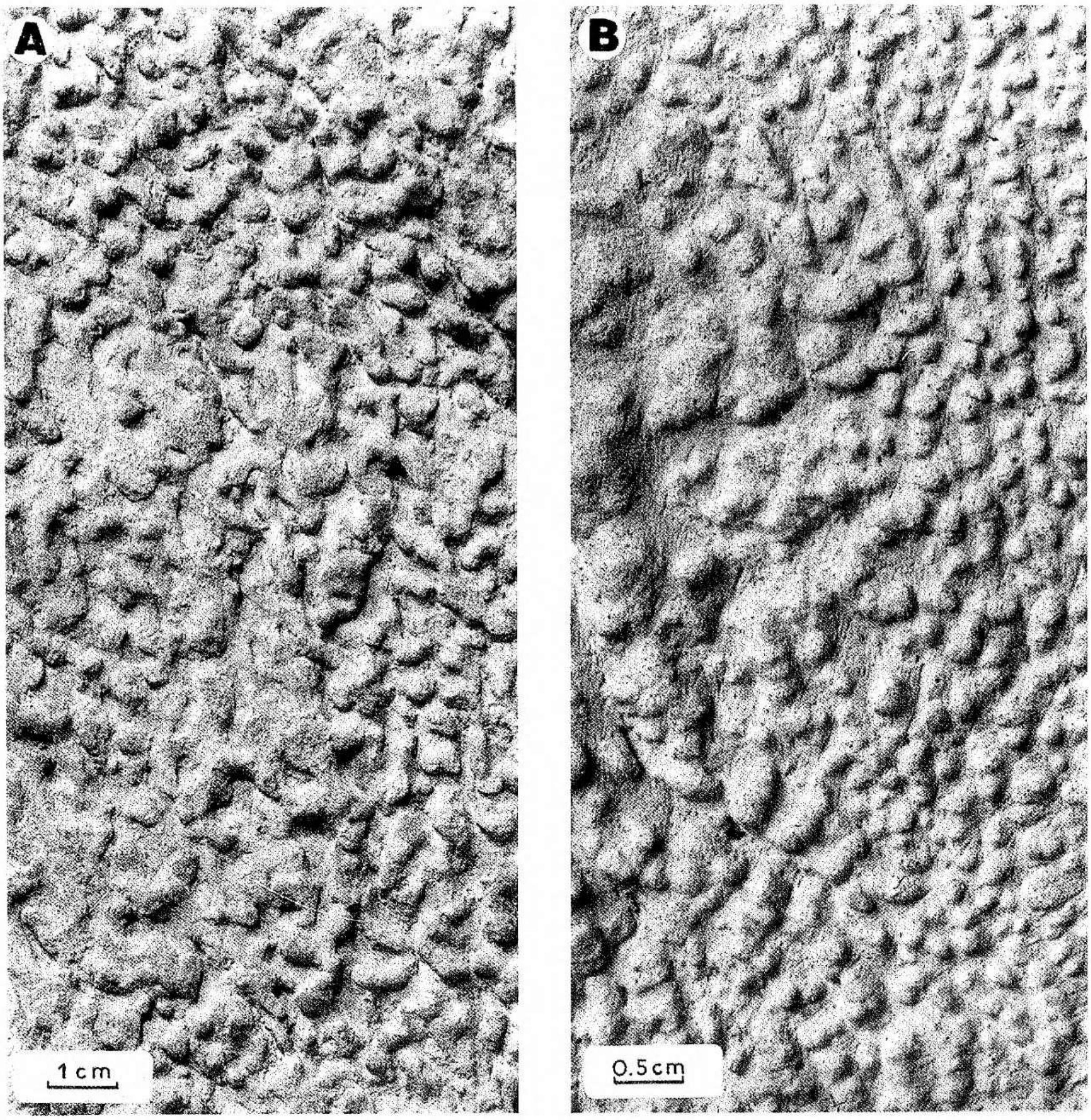


Plate 21 (**A**) Minor pendulous load deformations. Oligocene Krosno Beds, Rudawka Rymanowska. (**B**) Minor pendulous load deformations. Base of artificial turbidite

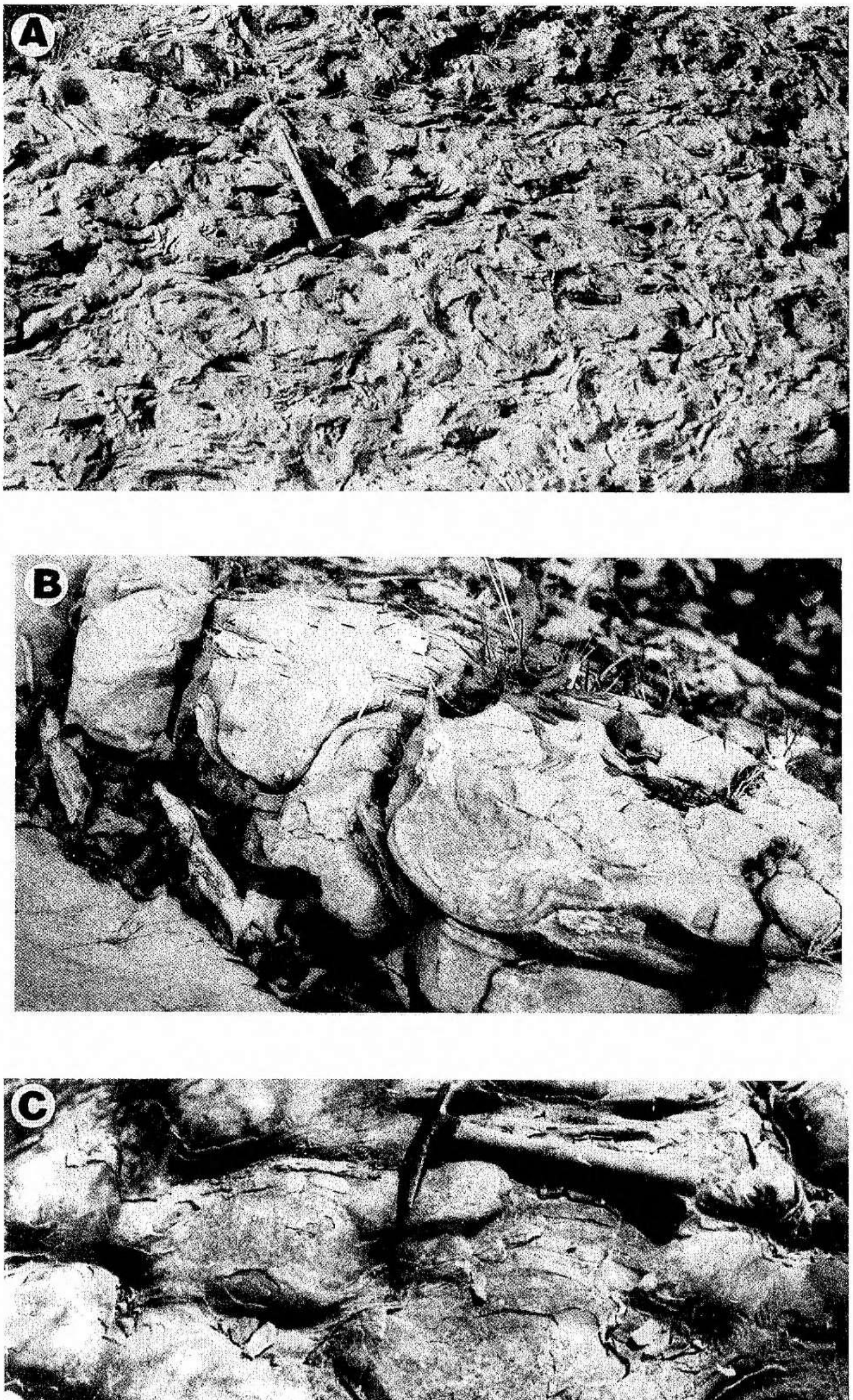


Plate 22 (**A**) Casque-shaped convolutions. Top view. Cretaceous Inoceramian Beds, Koninki. (**B**) Casque-shaped convolutions. Side view. Cretaceous Inoceramian Beds, Ropica Gorna. (**C**) Casque-shaped convolutions. Base of upper member of dichotomous convoluted bed in plan view. Oligocene Magura Beds, Ropica Górna

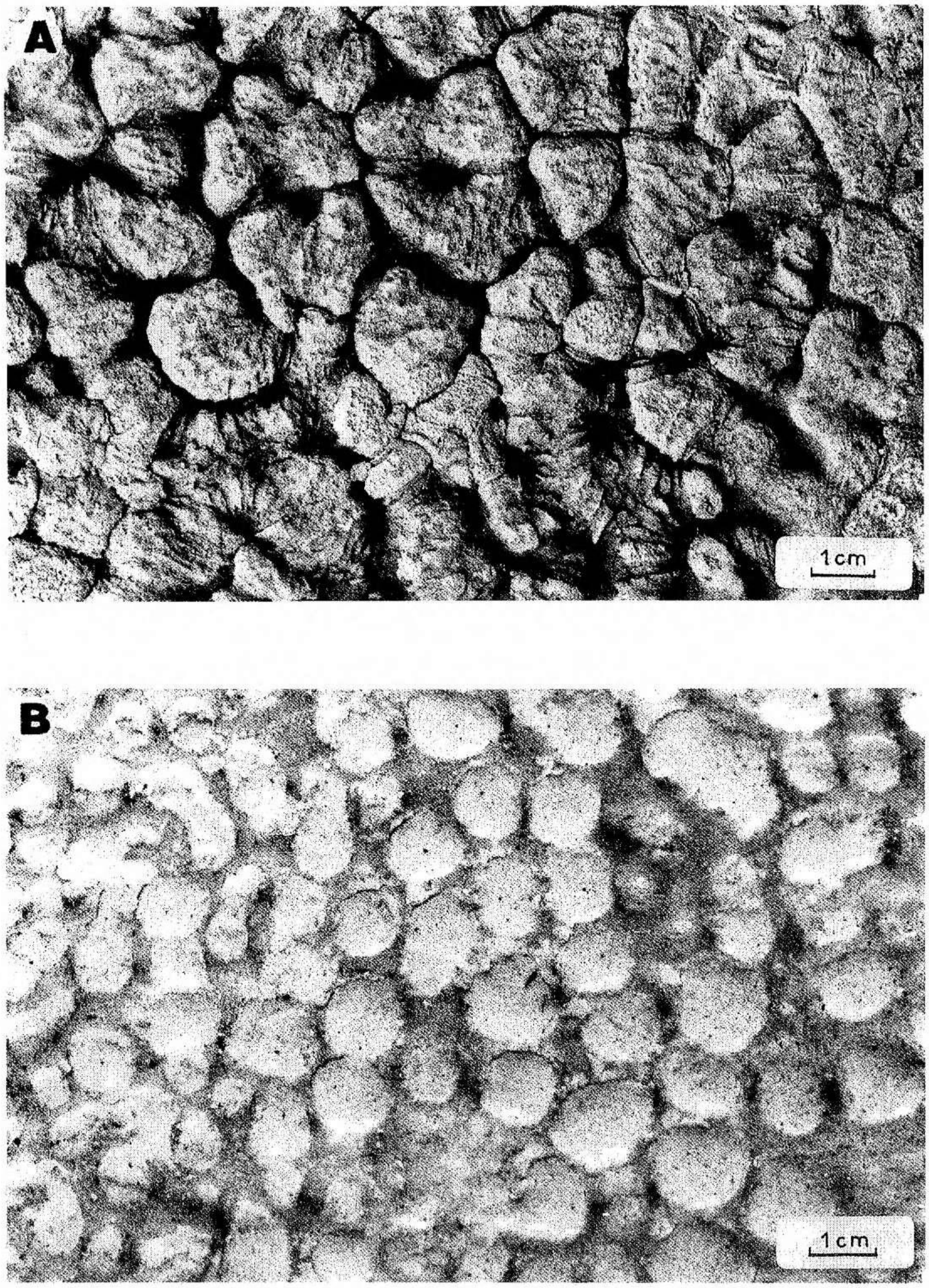


Plate 23 **(A)** Surface of rising columns of lower member (a) before they reached top surface of non mobile (ba) system. Experimental structure. **(B)** Experimental sorted polygons and sorted circles on top surface of non mobile (ba) system (after Cegła & Dżułyński, 1970)

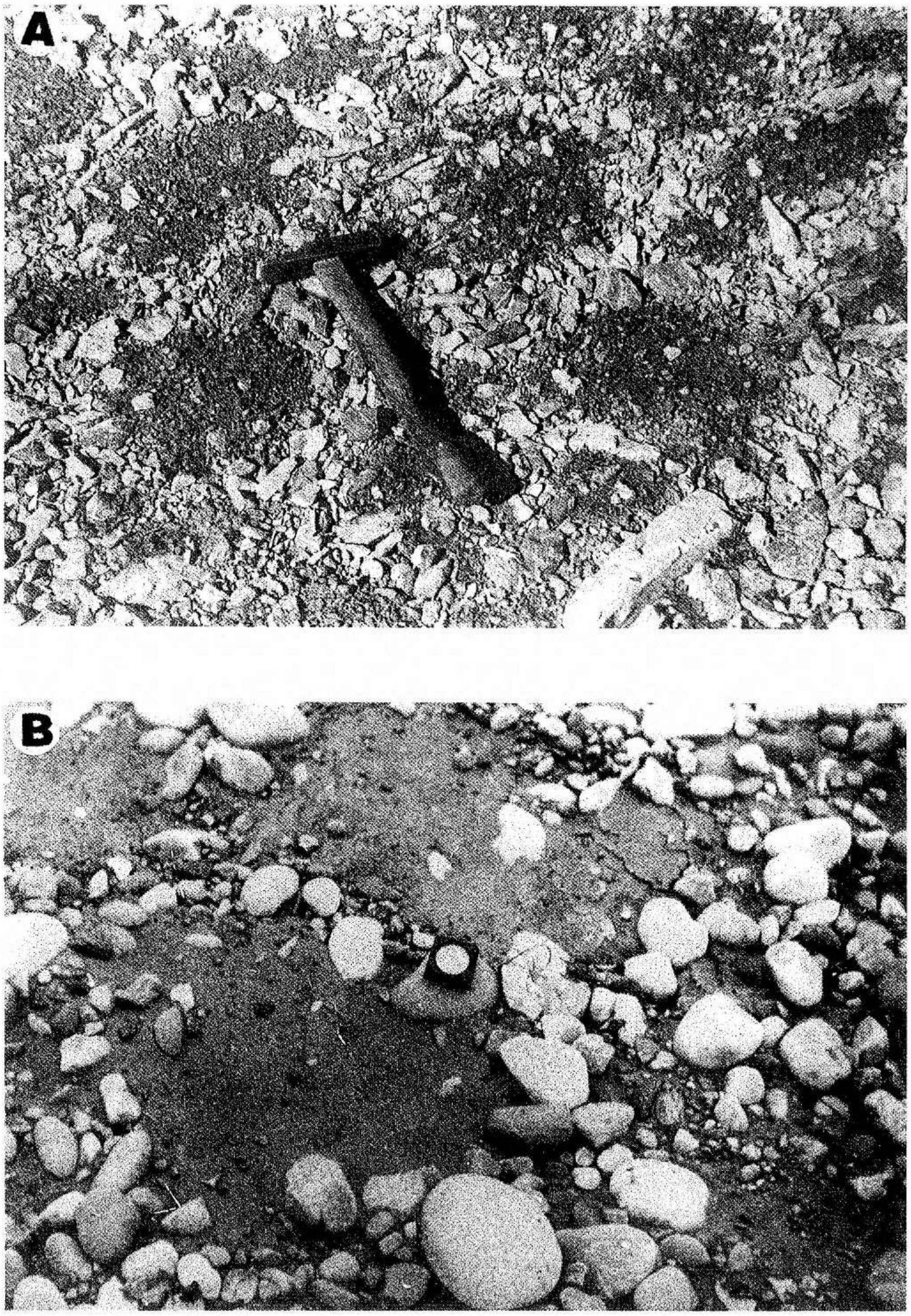


Plate 24 **(A)** Sorted polygons in periglacial environment. Engadin, Alps, Switzerland. **(B)** Sorted polygons in subtropical environment. Sivaliks, India (after R. Soja, in preparation)

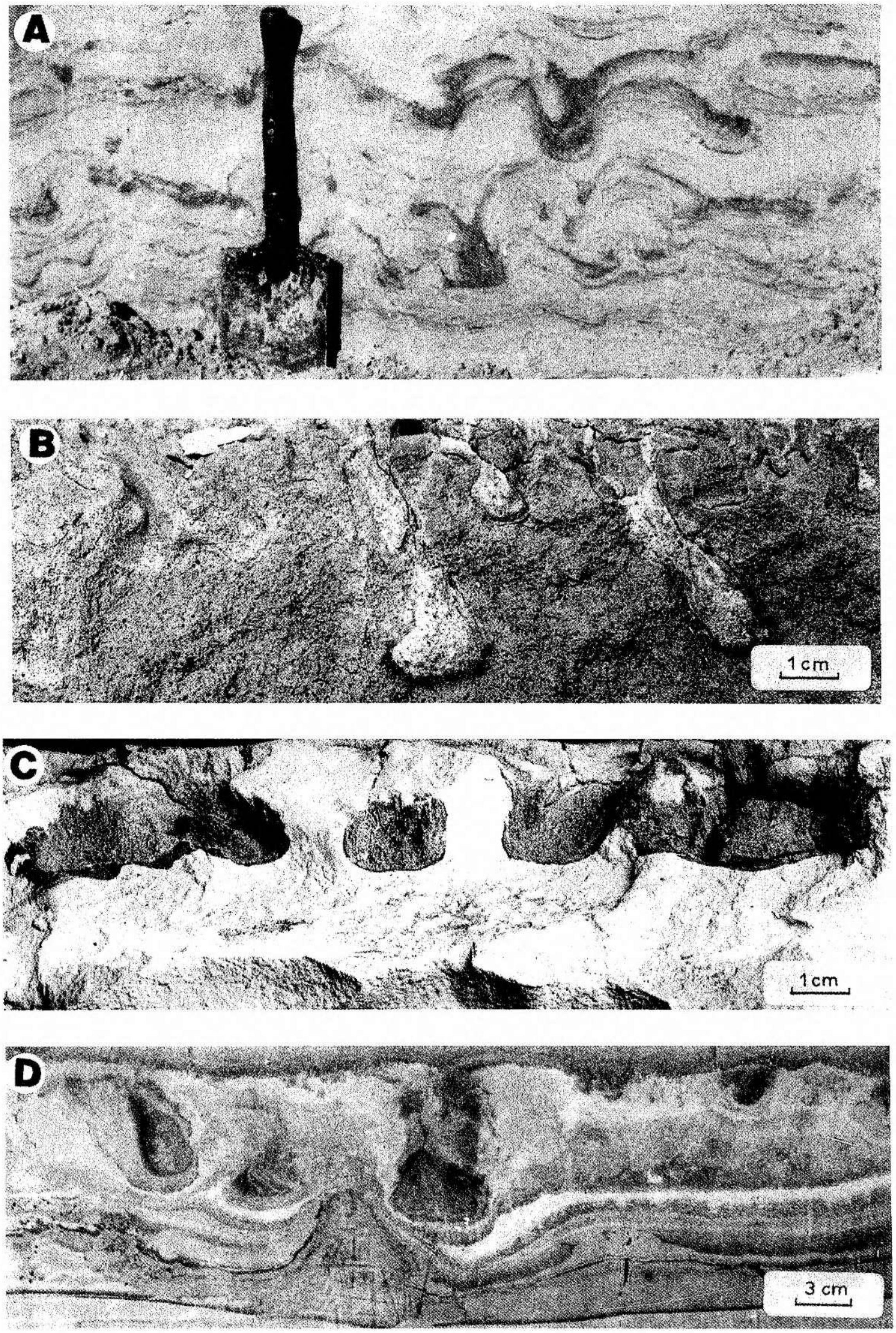


Plate 25 (**A**) Drop structures. Pleistocene fluvio-glacial sediments, Leżajsk, Carpathian foreland. (**B**) Experimental drop structures descending from tripple points of sorted polygons (after Cegła & Dżułyński, 1970). (**C**) Experimental drop structures descending from centres of polygonal compartments. (**D**) Irregular load pockets and unconformable truncation surface produced at sediment-water interface, in absence of current erosion. Experimental structure

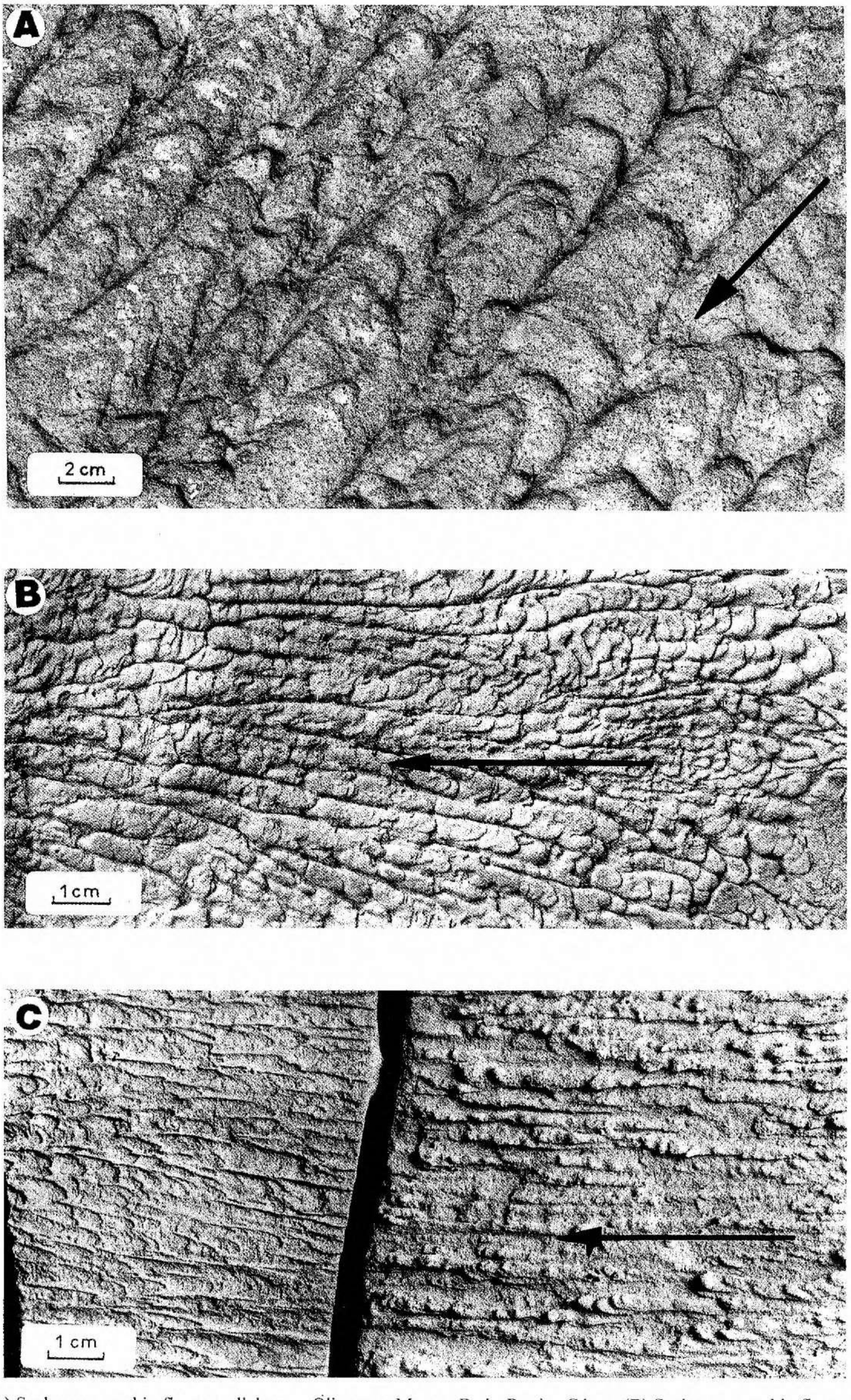


Plate 26 **(A)** Scales arranged in flow-parallel rows. Oligocene Magura Beds, Ropica Górna. **(B)** Scales arranged in flow-parallel rows. Base of artificial turbidite (after Dzulynski & Simpson, 1966). **(C)** Positives of experimental scales arranged in flow-parallel rows. Explanation in text (after Dzulynski & Simpson, 1966)

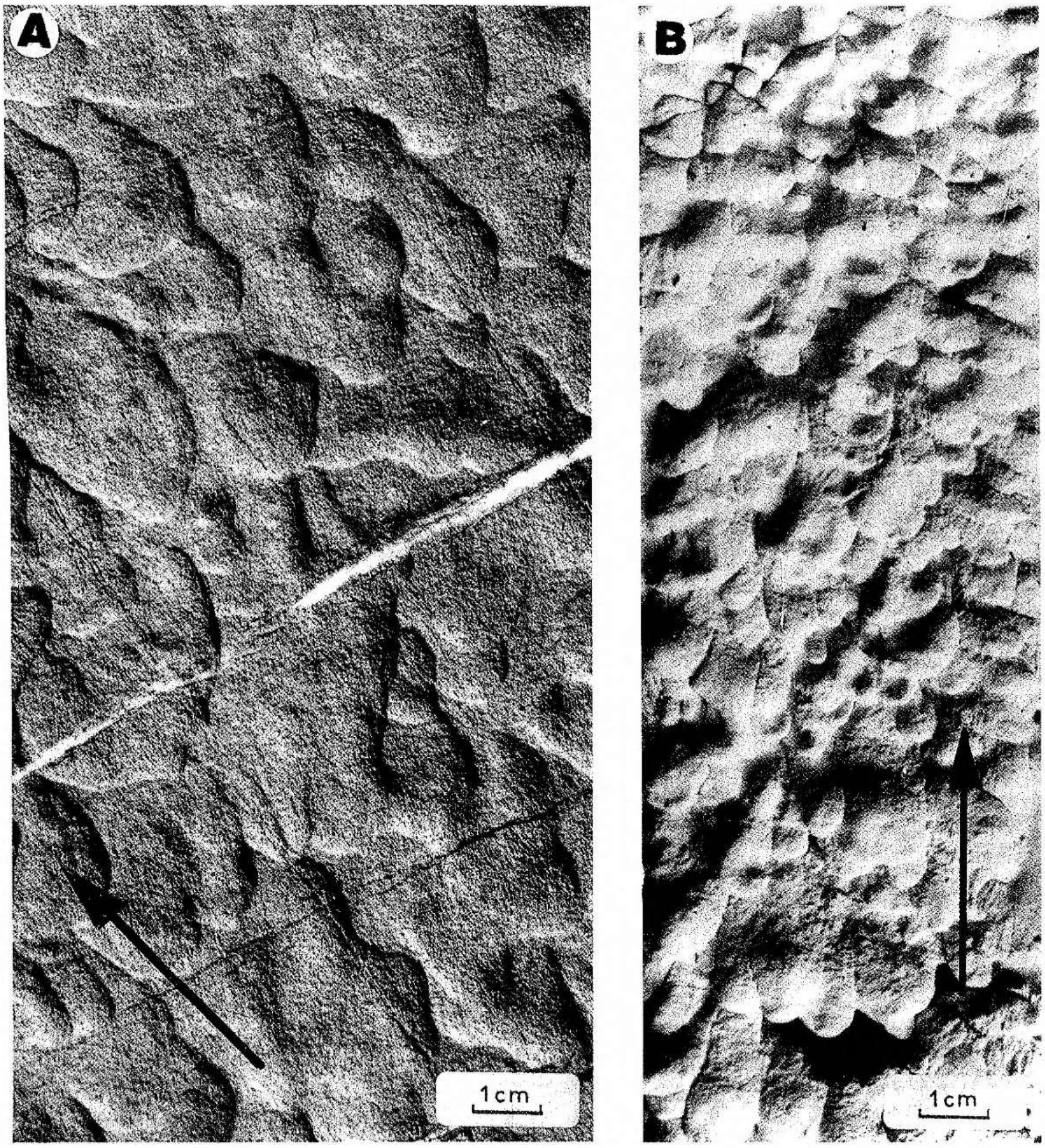


Plate 27 (A) Scales arranged in en echelon pattern. Oligocene Krosno Beds, Rudawka Rymanowka. (B) Scales arranged in en echelon pattern. Base of artificial turbidite (after Dżułyński & Walton 1965)

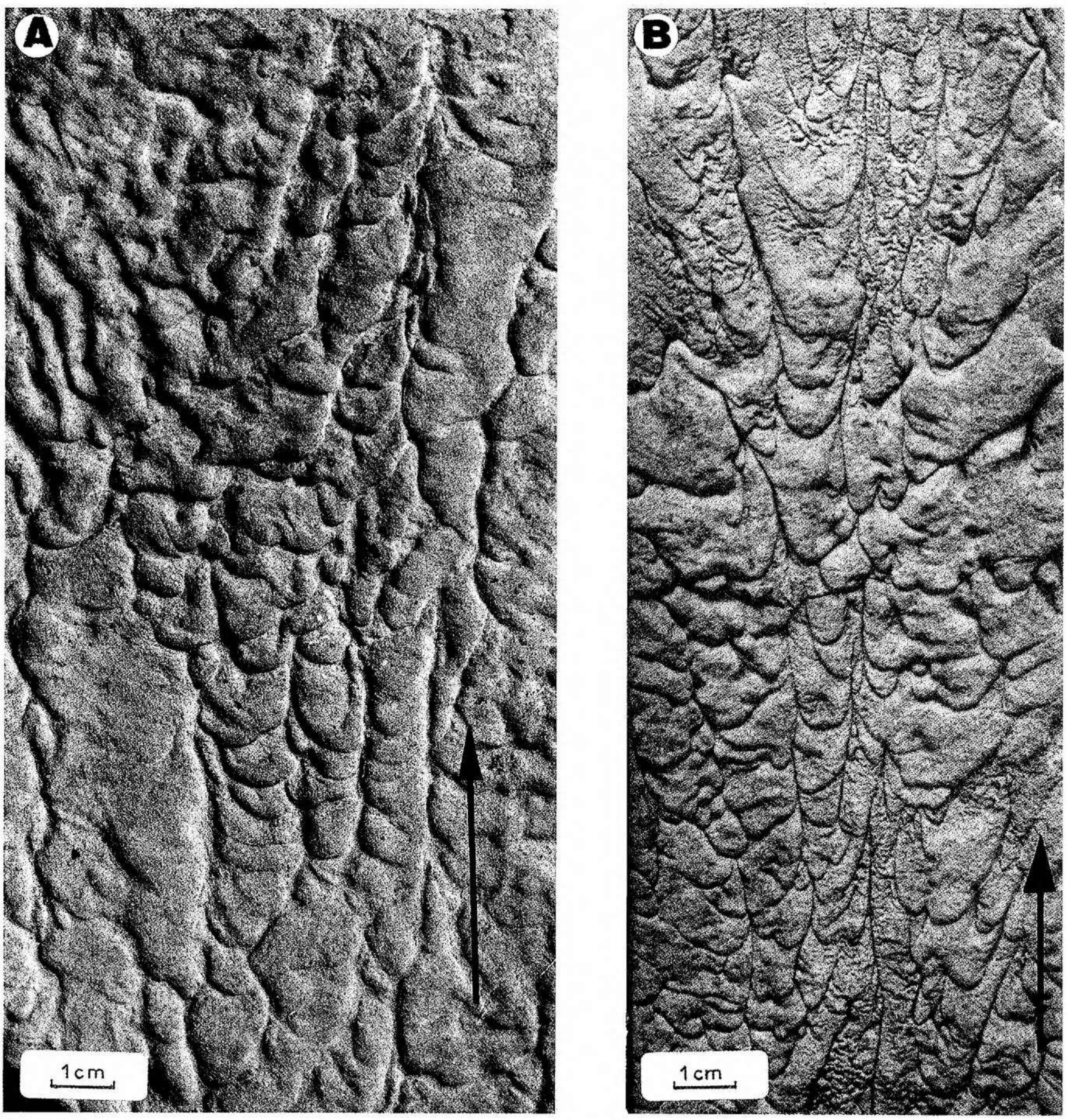
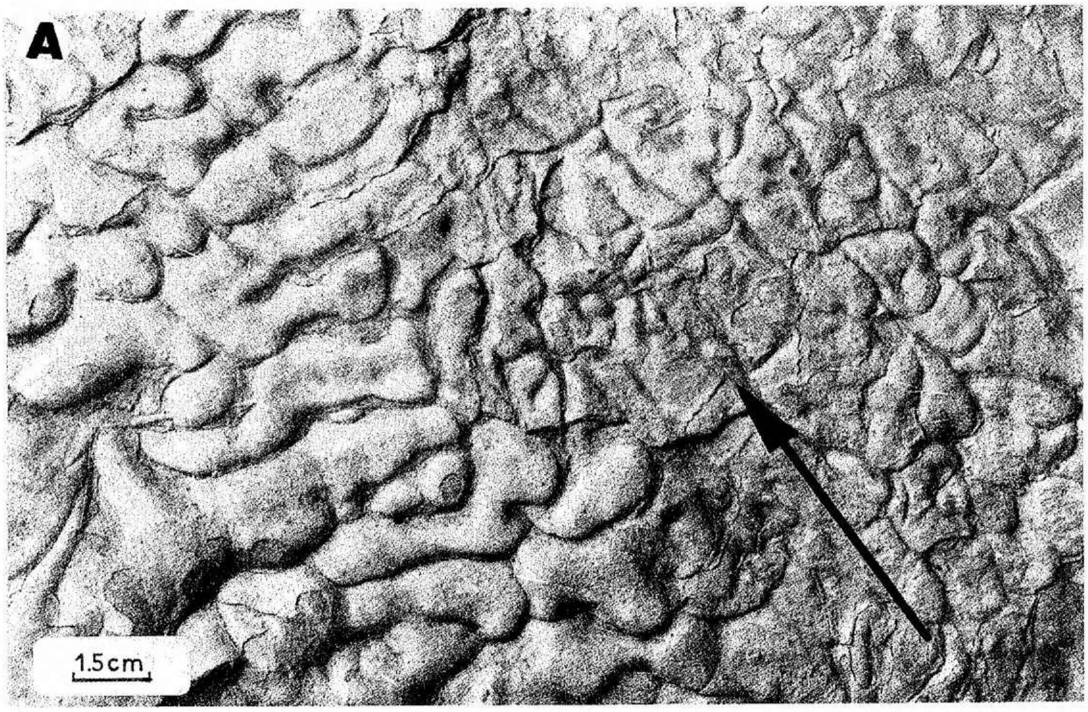


Plate 28 (**A**) Mixed assemblage of scales arranged in rows and in en echelon pattern. Oligocene Krosno Beds, Rudawka Rymanowska. (**B**) Mixed assemblage of scales. Base of artificial turbidite



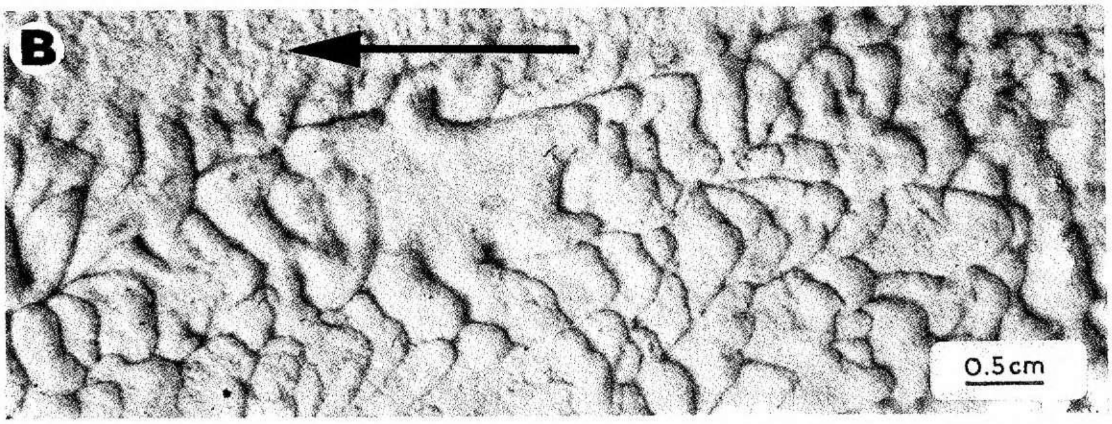


Plate 29 **(A)** Transverse scales and polygonal structures. Eocene Magura beds, Koninki. **(B)** Transverse scales. Base of artificial turbidite (after Dżułyński & Simpson, 1966a)

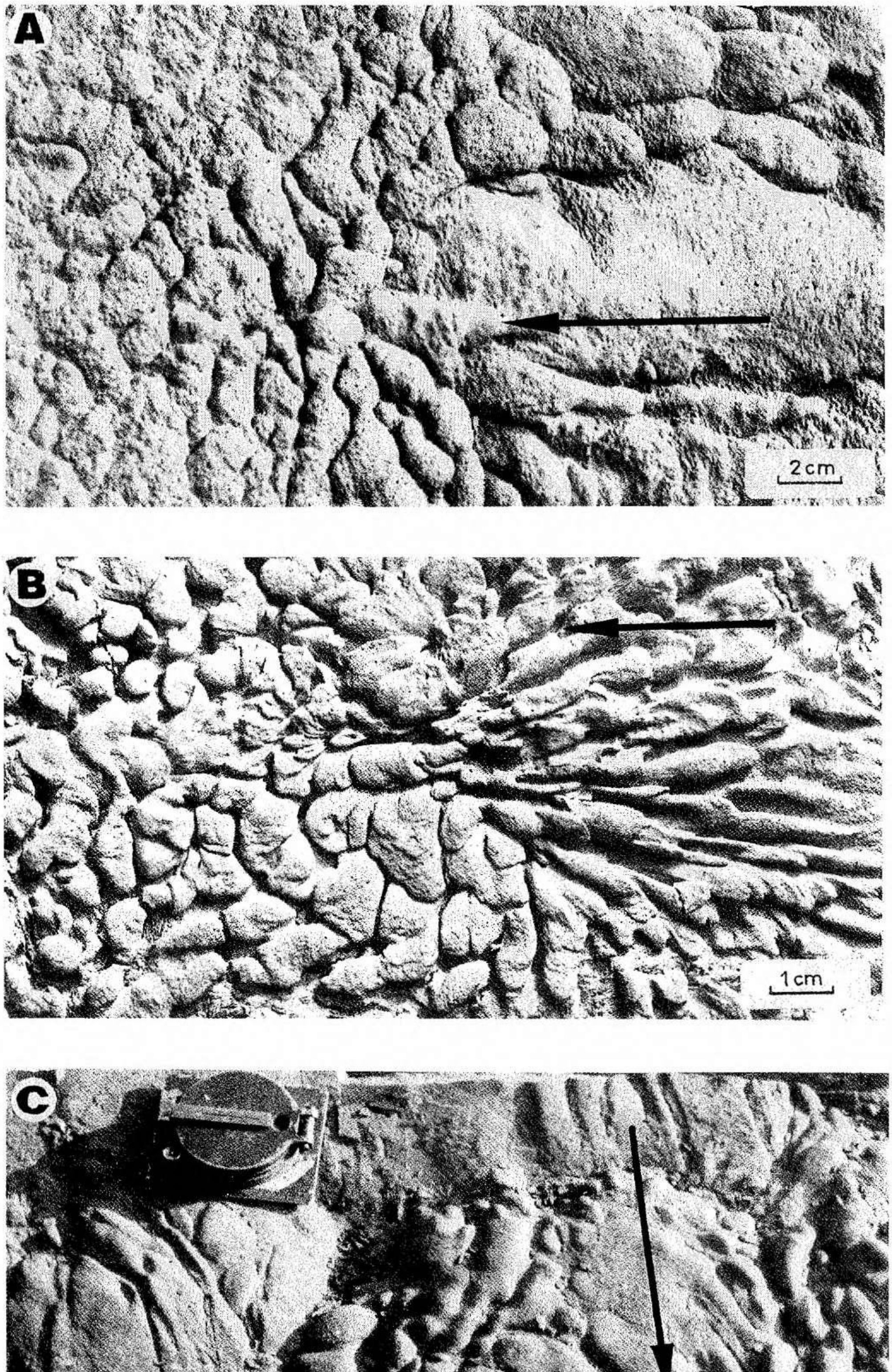
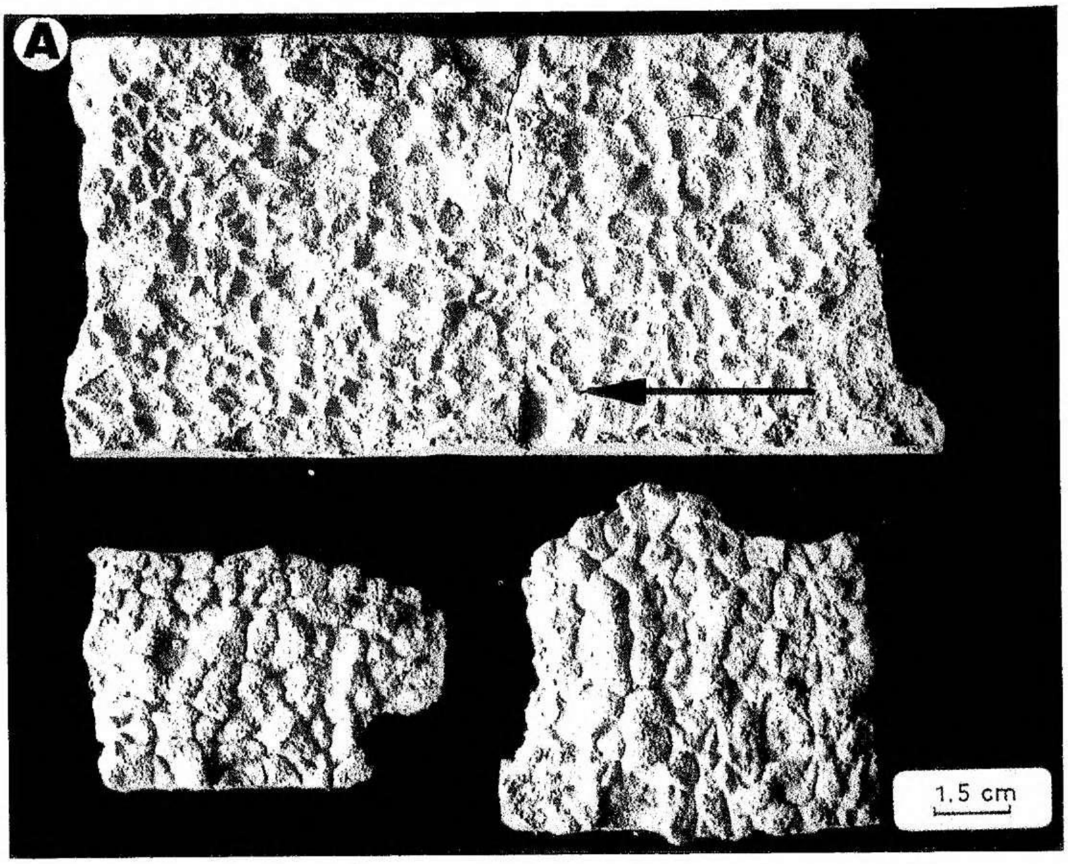


Plate 30 (**A**) Scaly structures passing into transverse structures. Oligocene Krosno Beds, Wolkowyja. (**B**) Scaly structures passing into transverse and roughly polygonal structures. Base of experimental turbidite (after Dżułyński, 1965). (**C**) Scaly structures and/or flute moulds passing into pillow-like structures. Oligocene Krosno beds, Tylawa



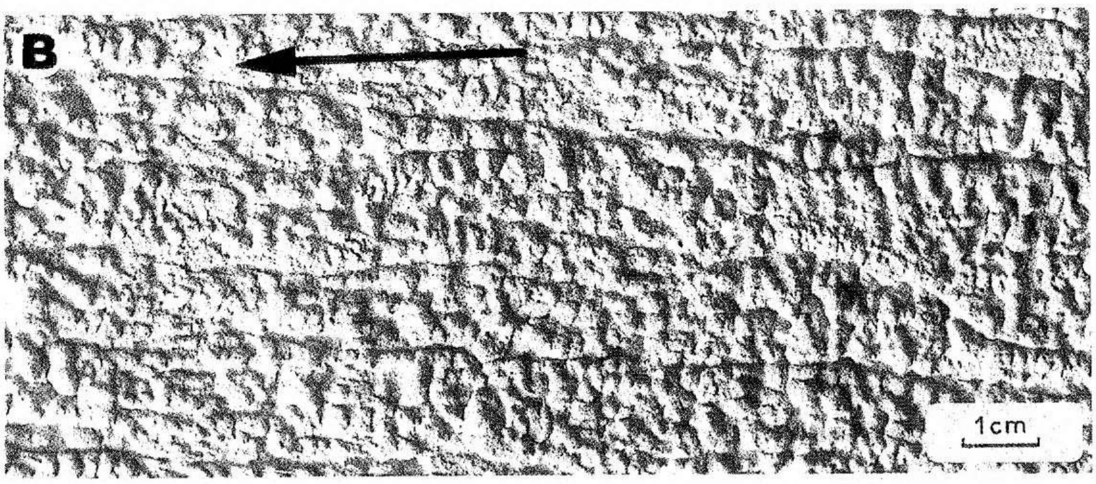


Plate 31 **(A)** Experimental transverse and polygonal scaly structure. Upper picture shows "positive" and lower "negative" aspects of same structure. **(B)** Transverse structures between moulds of longitudinal ridges. Base of artificial turbidite

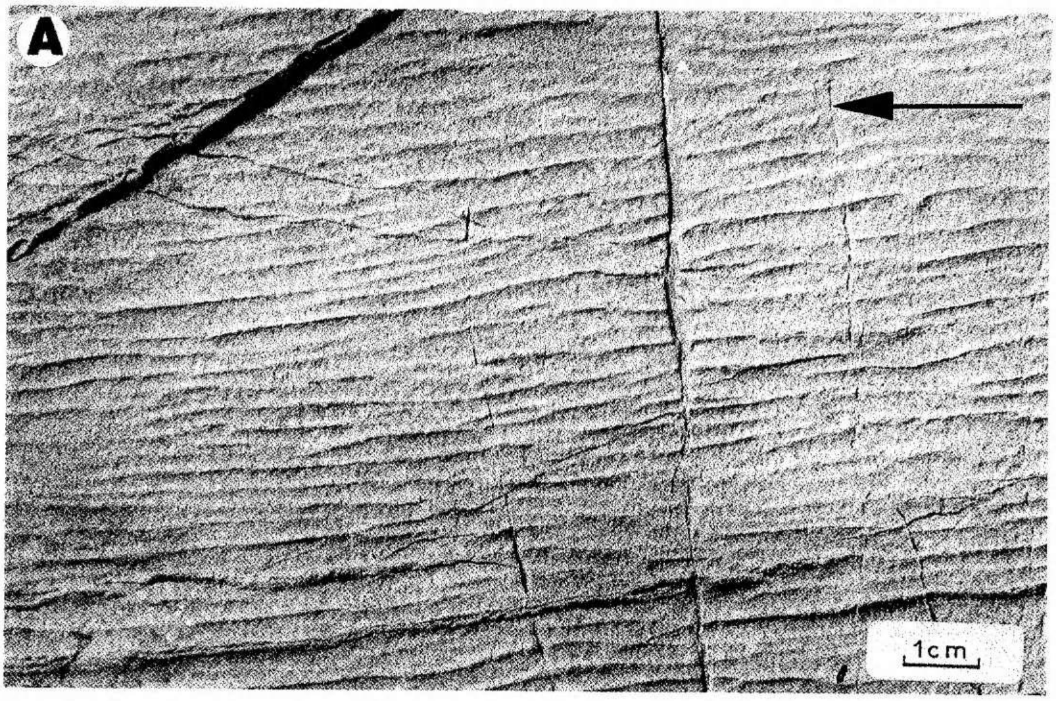


Plate 32 **(A)** Closely spaced moulds of l-ridges. Oligocene Krosno Beds, Rudawka Rymanowska. **(B)** Closely spaced moulds of l-ridges. Base of artificial turbidite (after Dżułyński, 1965)

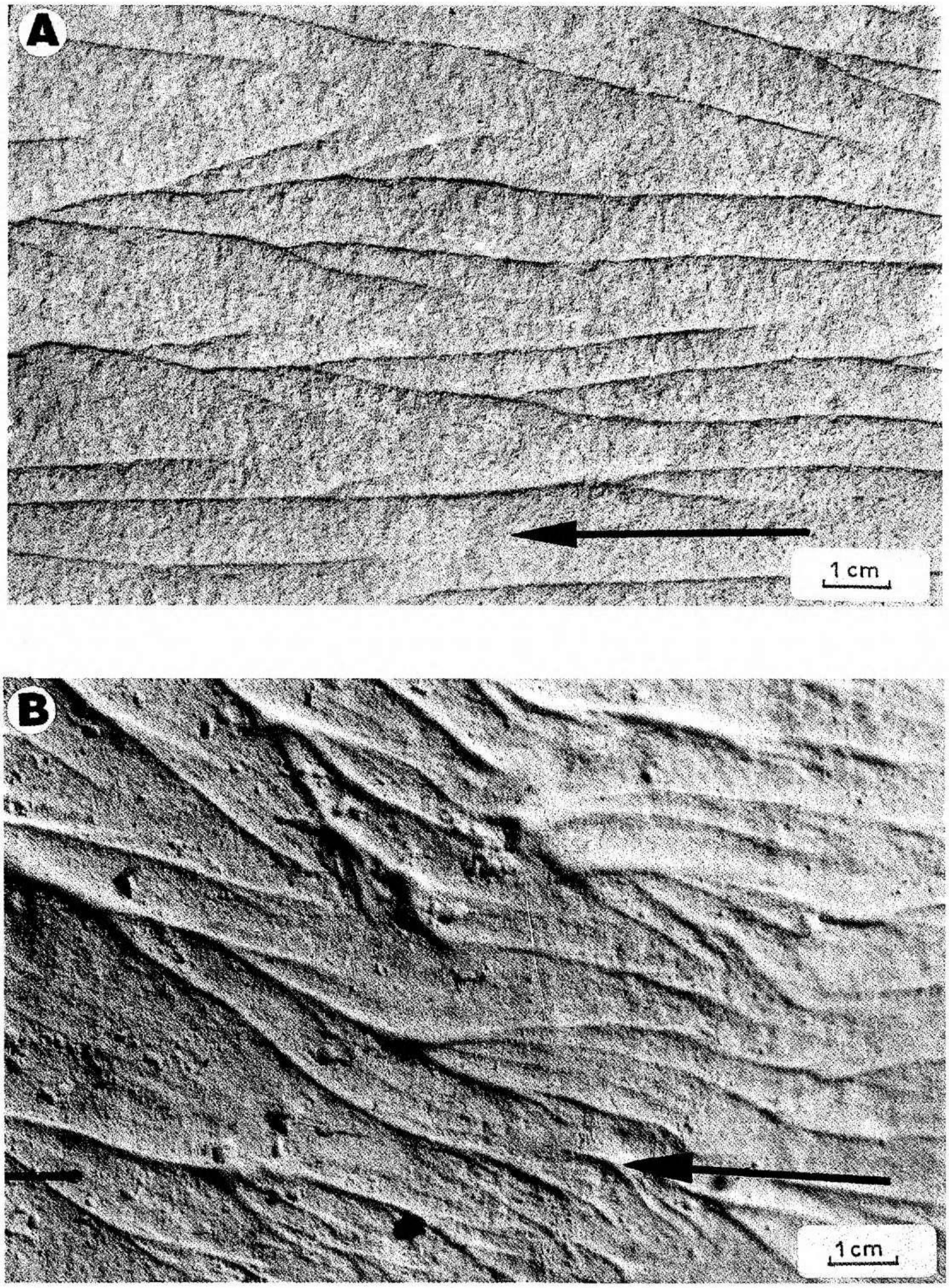


Plate 33 **(A)** Moulds of dendritic ridges. Oligocene Krosno beds, Mokre. **(B)** Experimentally produced dendritic ridges. Explanation in text

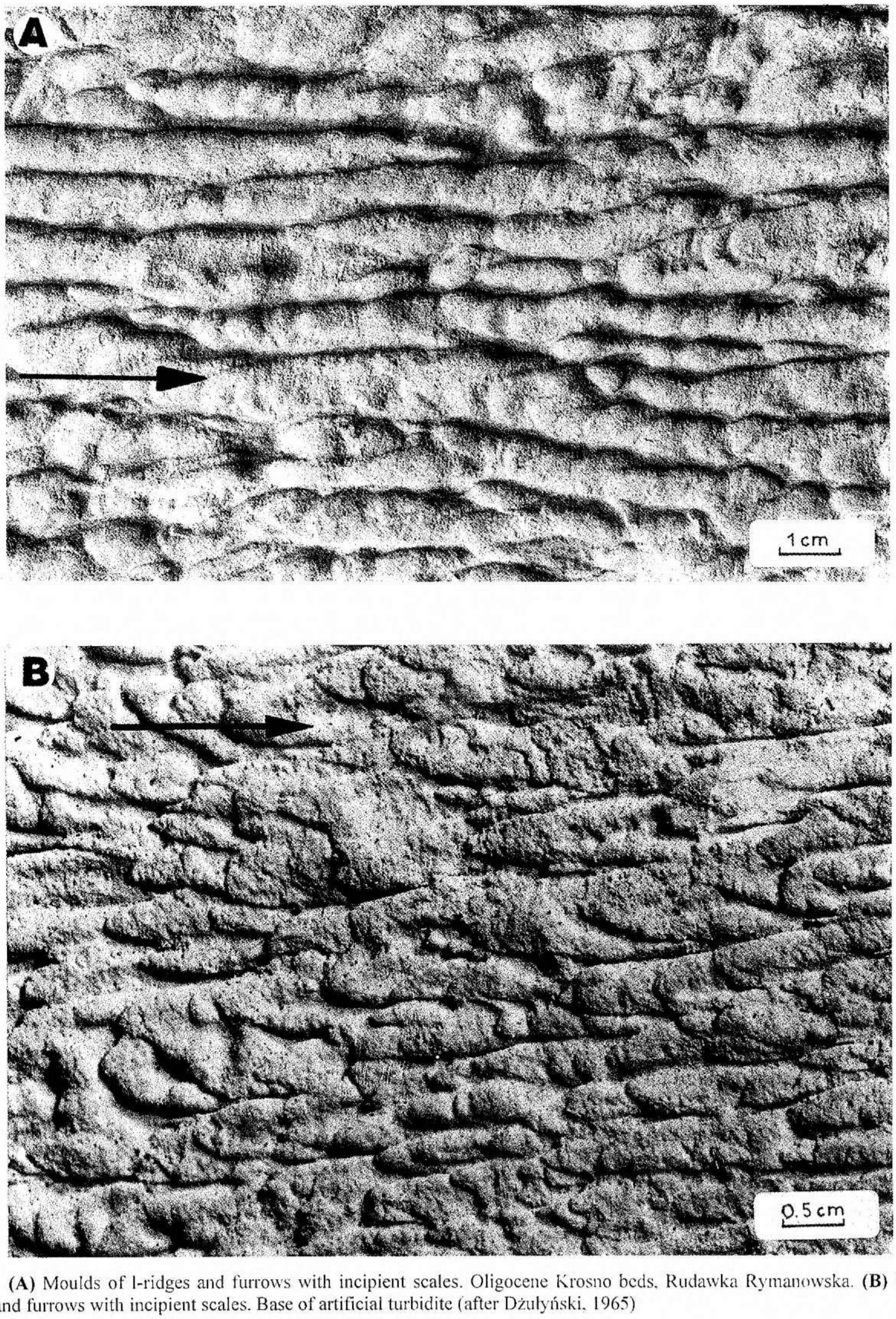


Plate 34 **(A)** Moulds of l-ridges and furrows with incipient scales. Oligocene Krosno beds, Rudawka Rymanowska. **(B)** Moulds of l-ridges and furrows with incipient scales. Base of artificial turbidite (after Dżułyński, 1965)

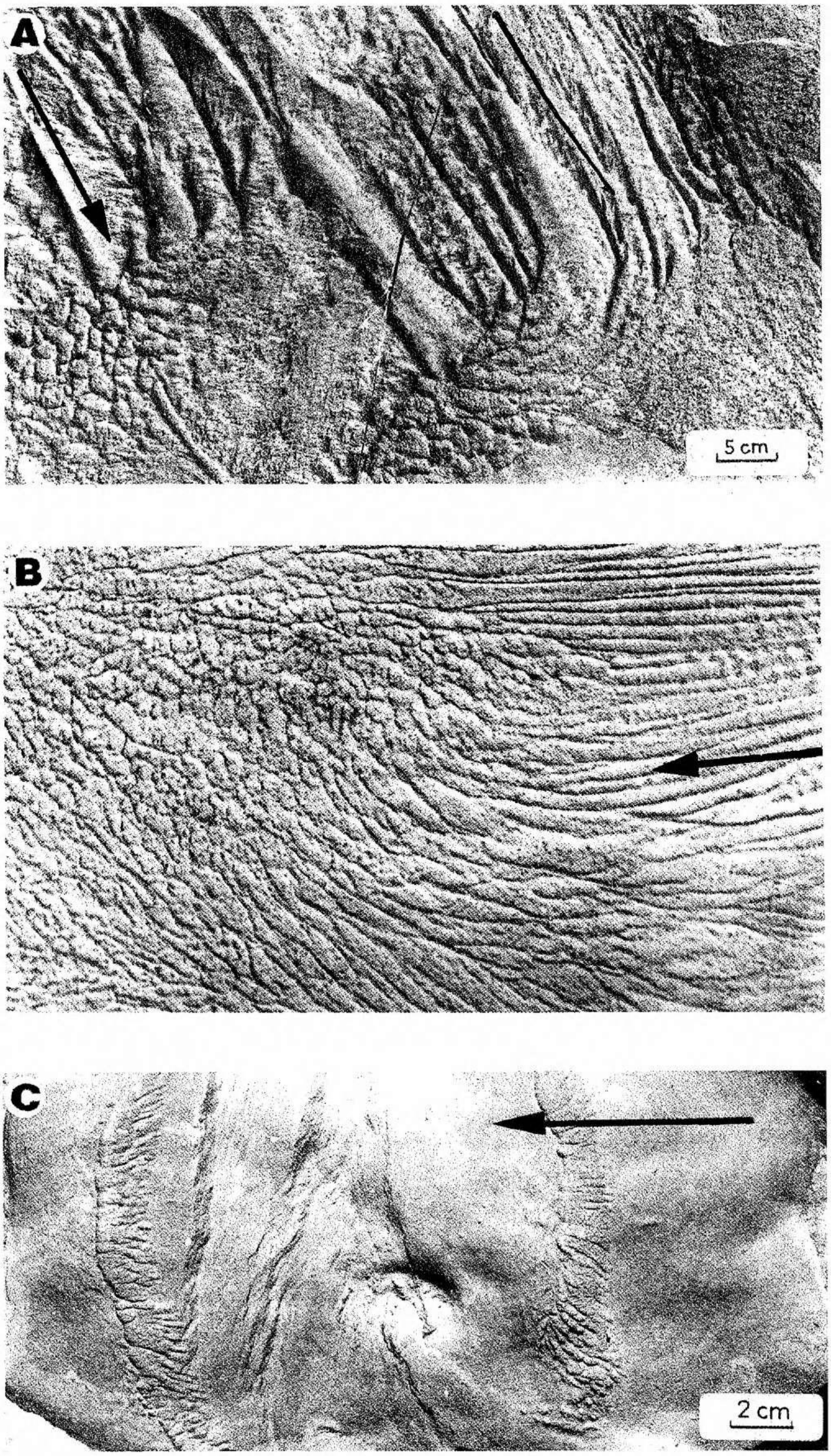


Plate 35 **(A)** Moulds of polygonal structures formed in zone of interference between deflected and undeflected flow. Oligocene Krosno Beds, Wołkowyja. **(B)** Moulds of polygonal structures formed in zone of interface between deflected and undeflected flow. Base of artificial turbidite (after Dżułyński & Simpson 1966b). **(C)** Base of loadcasted ripples with moulds of l-ridges and dilatation ridges. Oligocene Krosno beds, Rudawka Rymanowska

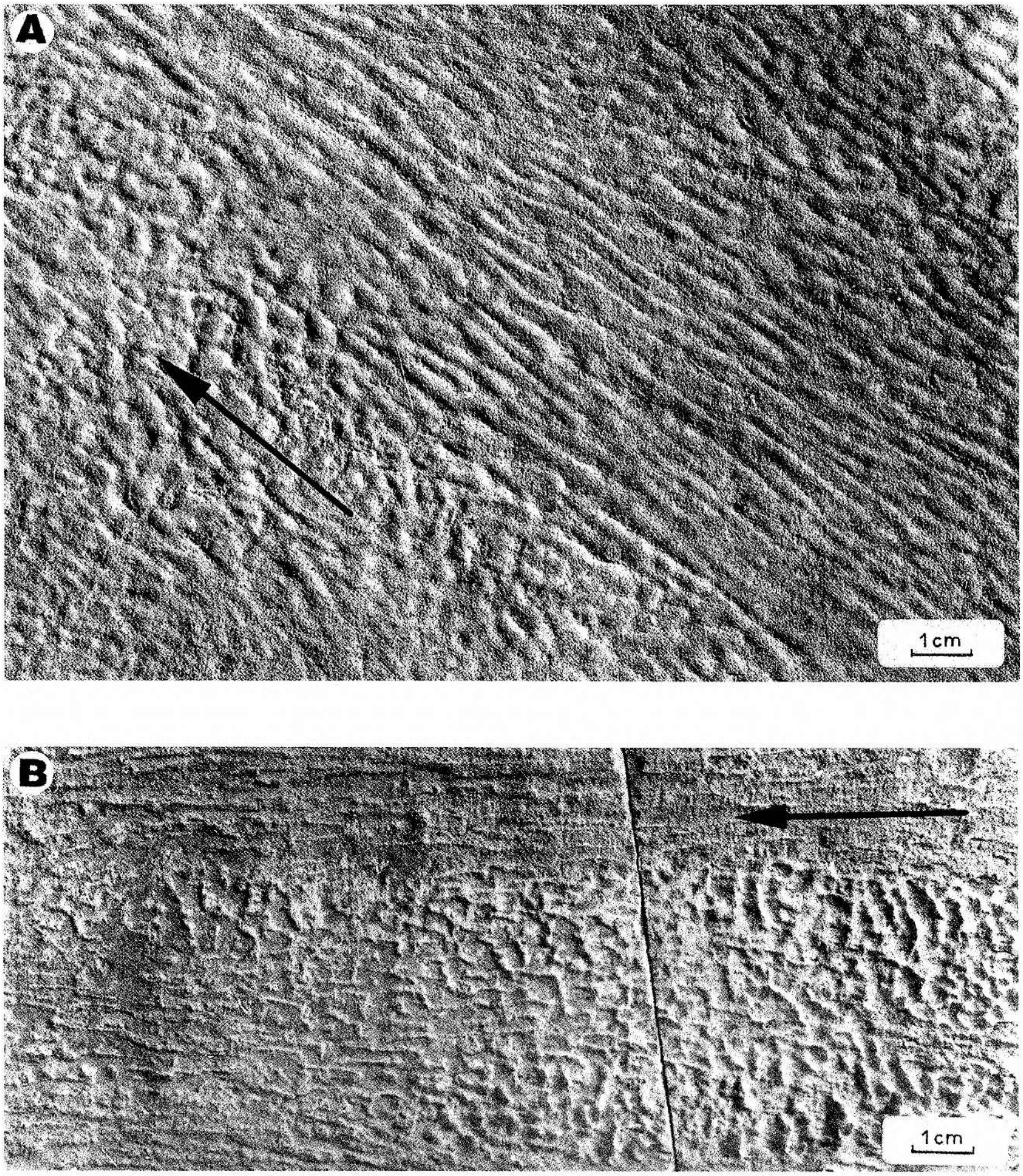


Plate 36 **(A)** Top of parting surface of sandstone bed showing alternating zones of longitudinal and polygonal ridges. Oligocene Menilite Beds, Wetlina. **(B)** Positives of alternating longitudinal and polygonal ridges. Experimental structures (after Dzulynski & Simpson, 1966)

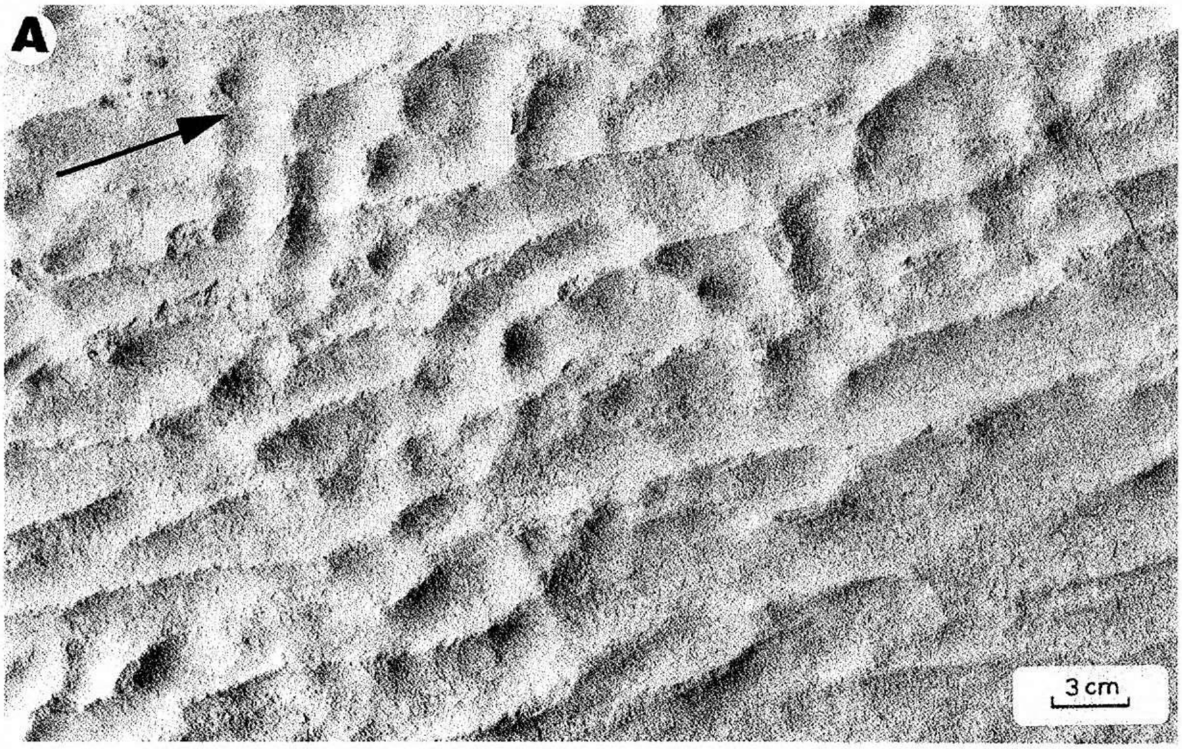
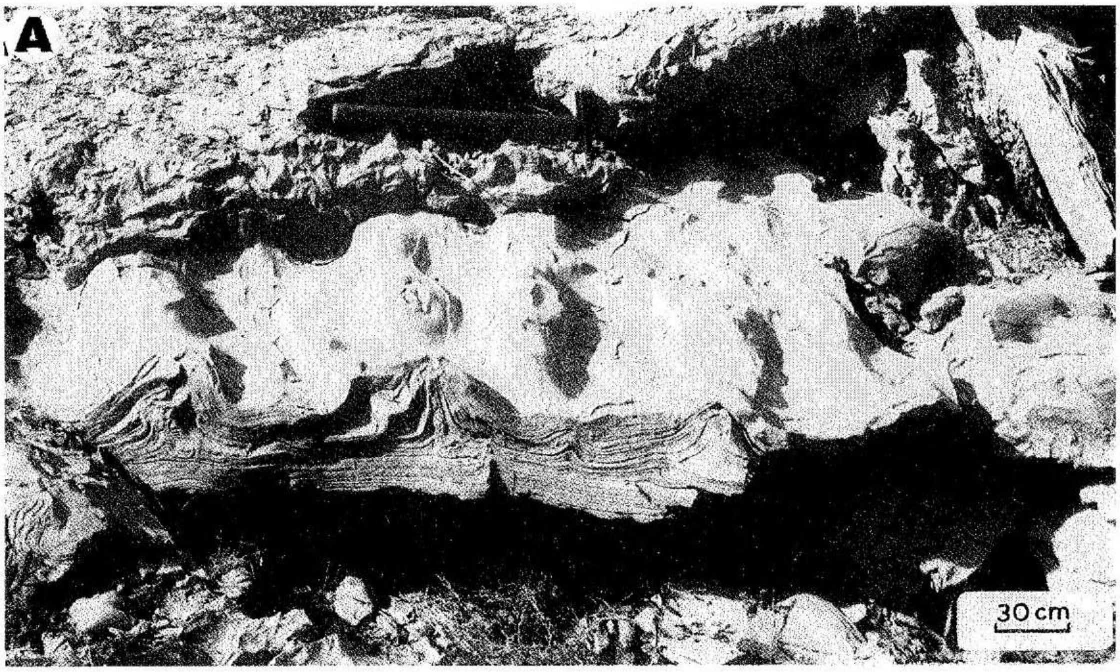


Plate 37 **(A)** Positives of longitudinal and square ridges on intrastratal parting surface. Oligocene Krosno Beds, Rudawka Rymanowska. **(B)** Longitudinal ridges in laminated sandstone, locally disturbed by rising conical knobs (comp. Pl. 39A). Top view. Oligocene Krosno Beds, Besko



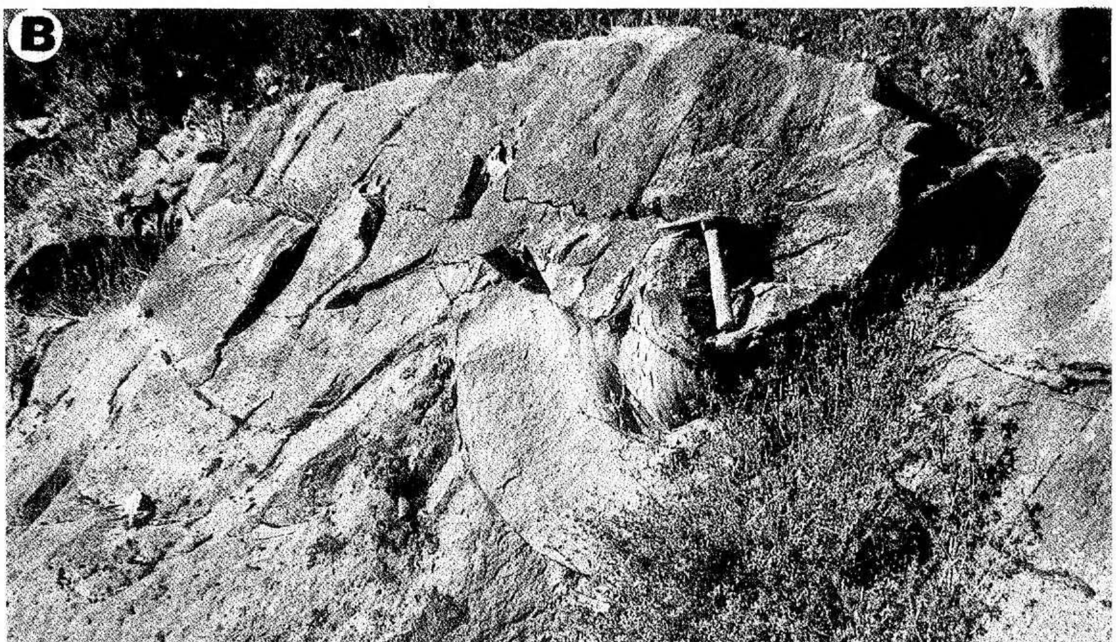


Plate 38 **(A)** Ridged convolutions in dichotomous bed whose lower member is horizontally laminated and upper one rippled. Oligocene Krosno Beds, Mymoń. **(B)** Ridged convolutions in laminated sandstone. Oligocene Krosno Beds, Sieniawa

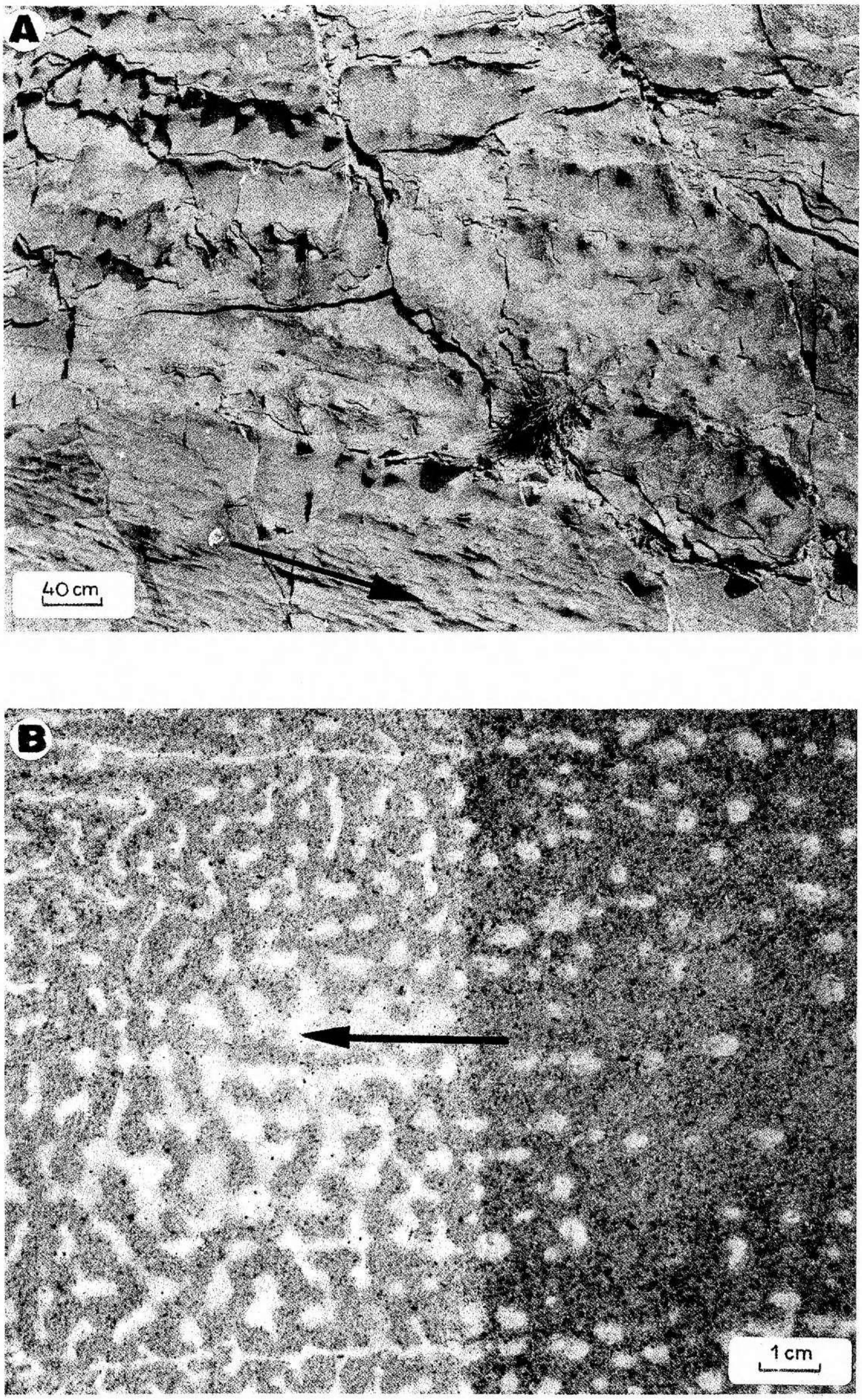


Plate 39 **(A)** Besko-type convolutions. Oligocene Krosno Beds, Besko. Top view. **(B)** Experimentally produced deformation pattern similar to that of Besko type convolutions. Explanations in text

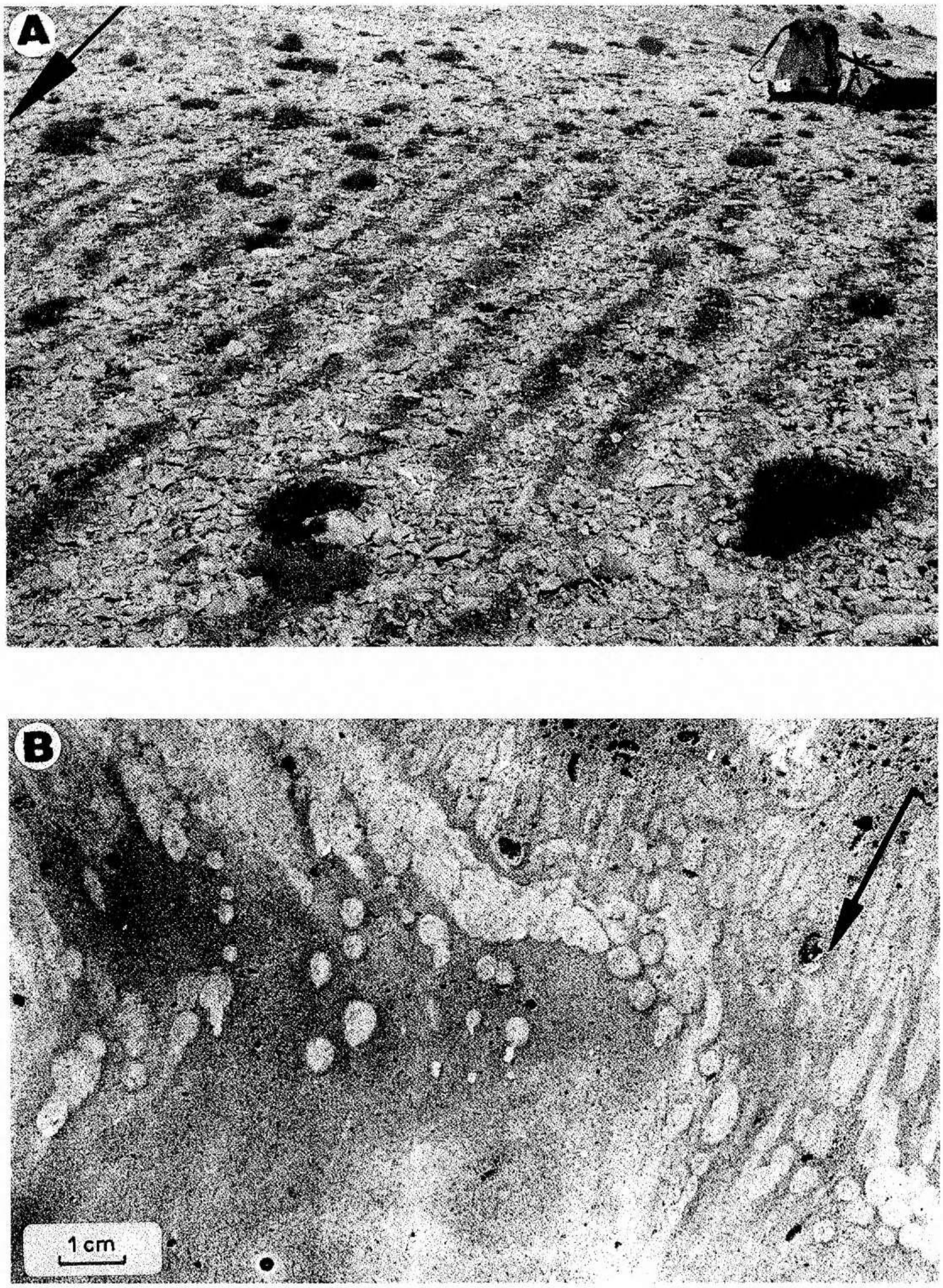


Plate 40 (**A**) Sorted stripes. Engadin. Swiss Alps. (**B**) Experimentally produced sorted stripes. Explanations in text

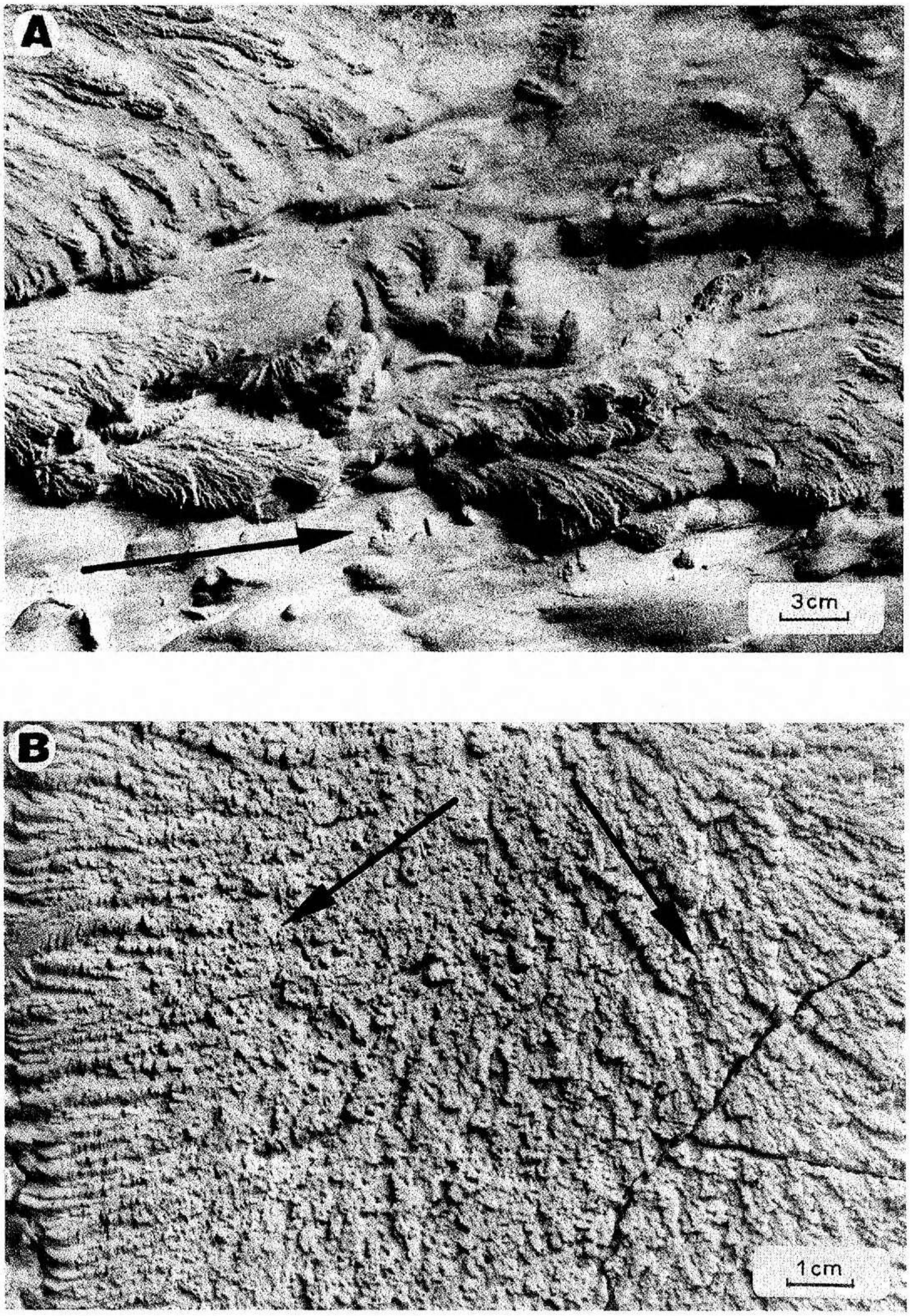


Plate 41 (**A**) Frondescent marks. Oligocene Krosno beds, Rudawka Rymanowska. (**B**) Experimental frondescent markings (after Dżułyński & Walton 1965)

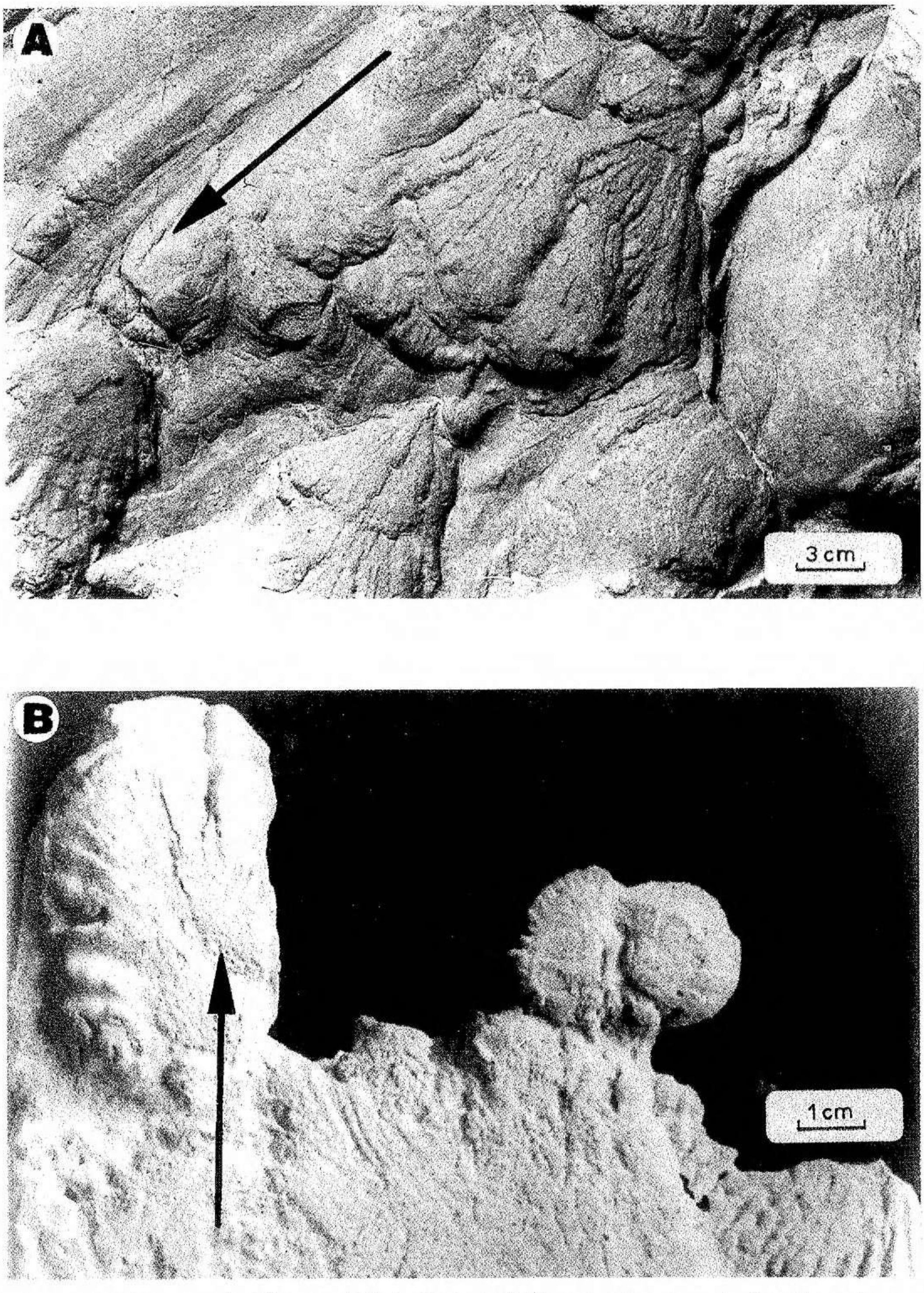


Plate 42 **(A)** Bulbous frondescent marks. Oligocene Podhale flysch. **(B)** Bulbous frondescent marks. Experimental structure

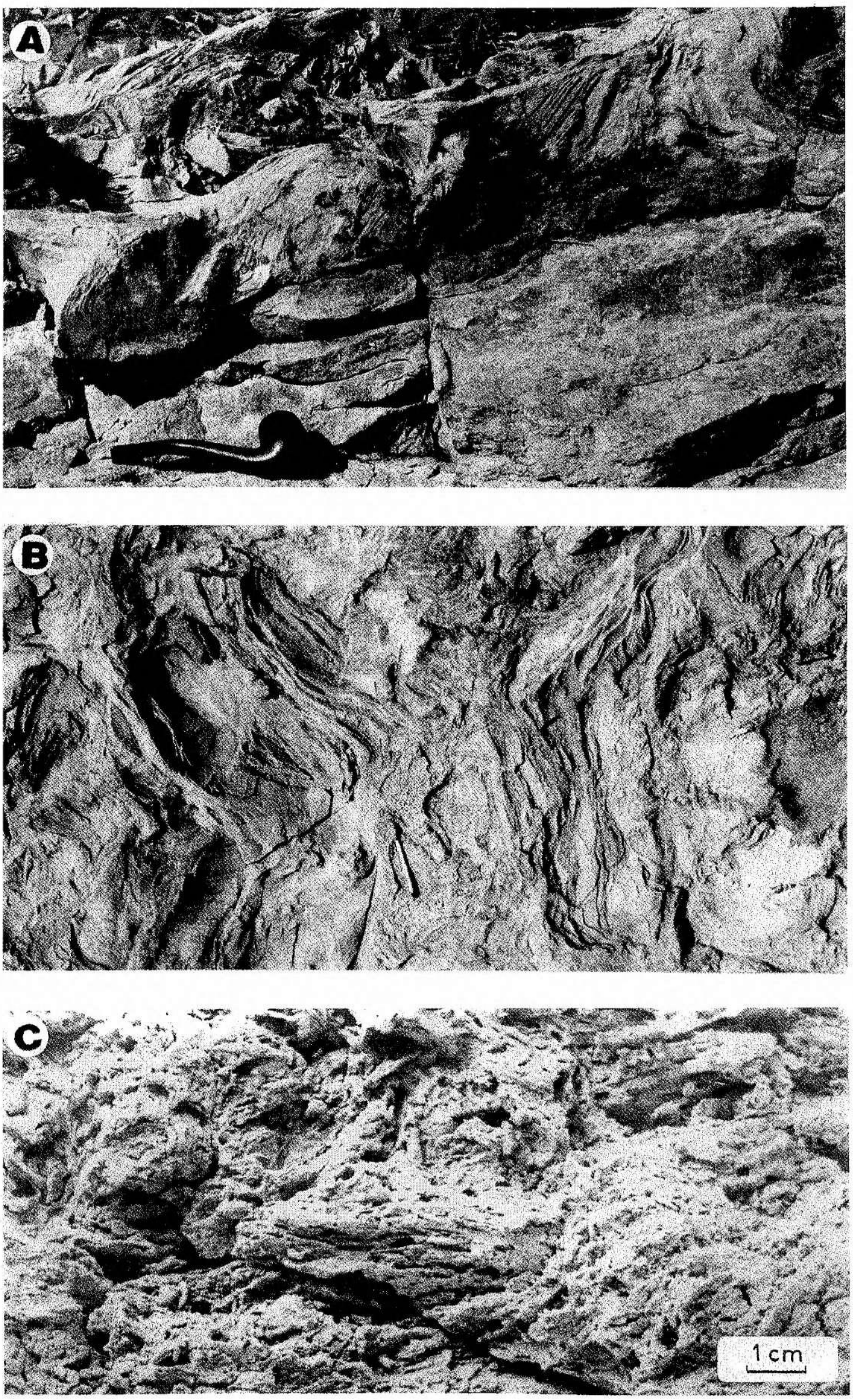


Plate 43 Crinkled lamination on top of massive sandstone. (**A** - side view; **B** - top view). Oligocene Magura beds, Bednarka. (**C**) Experimental crinkled lamination. Top view

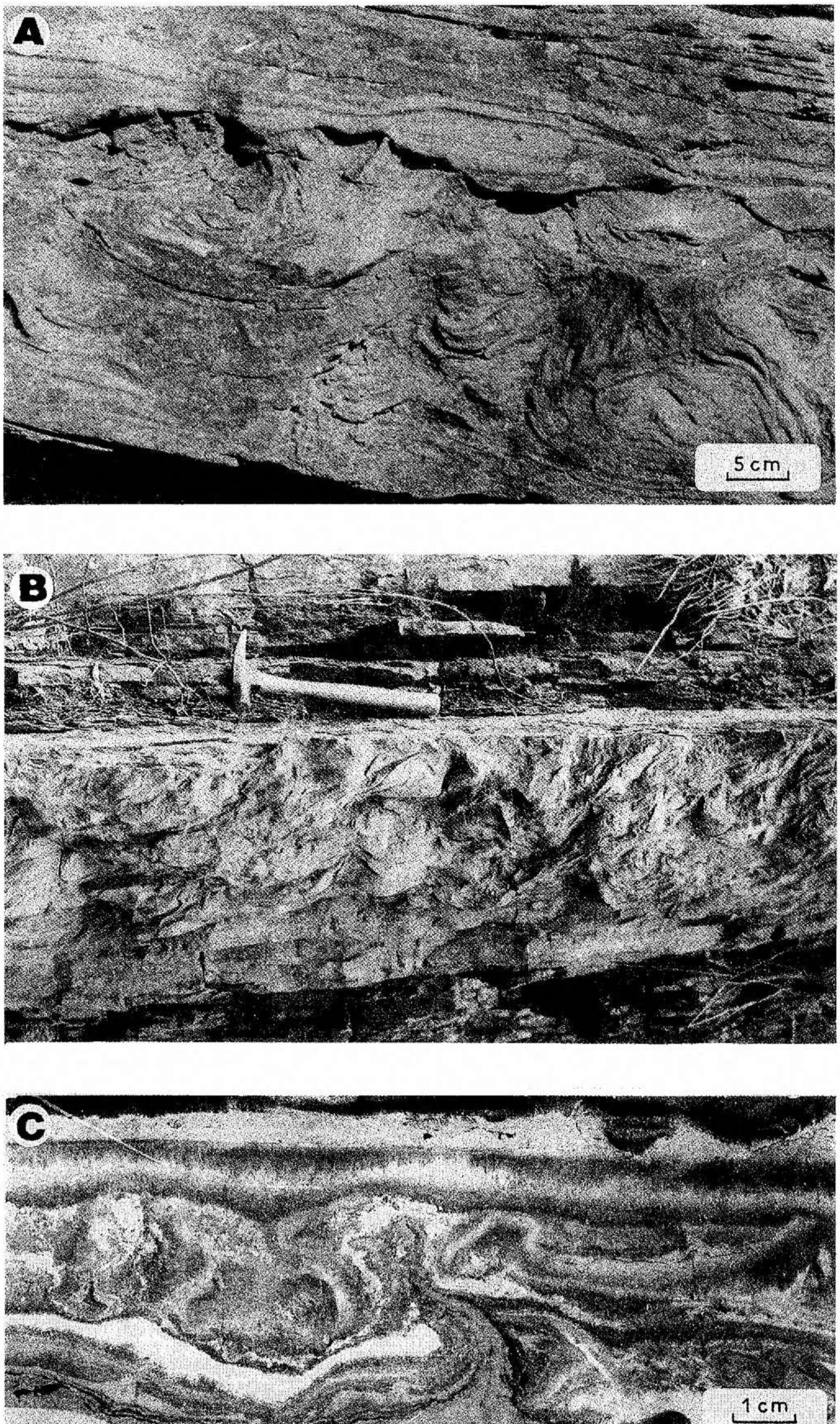


Plate 44 **(A)** Erosional contact between convoluted and horizontally laminated member of amalgamated bed. Oligocene Krosno beds, Sieniawa. **(B)** Flat surface produced by self-truncation of convoluted layers at water/sediment interface in absence of current erosion. Cretaceous, Inoceramian beds, Szczawa (original position of strata is vertical). (C) Flat self-truncation surface produced experimentally. Explanations in text

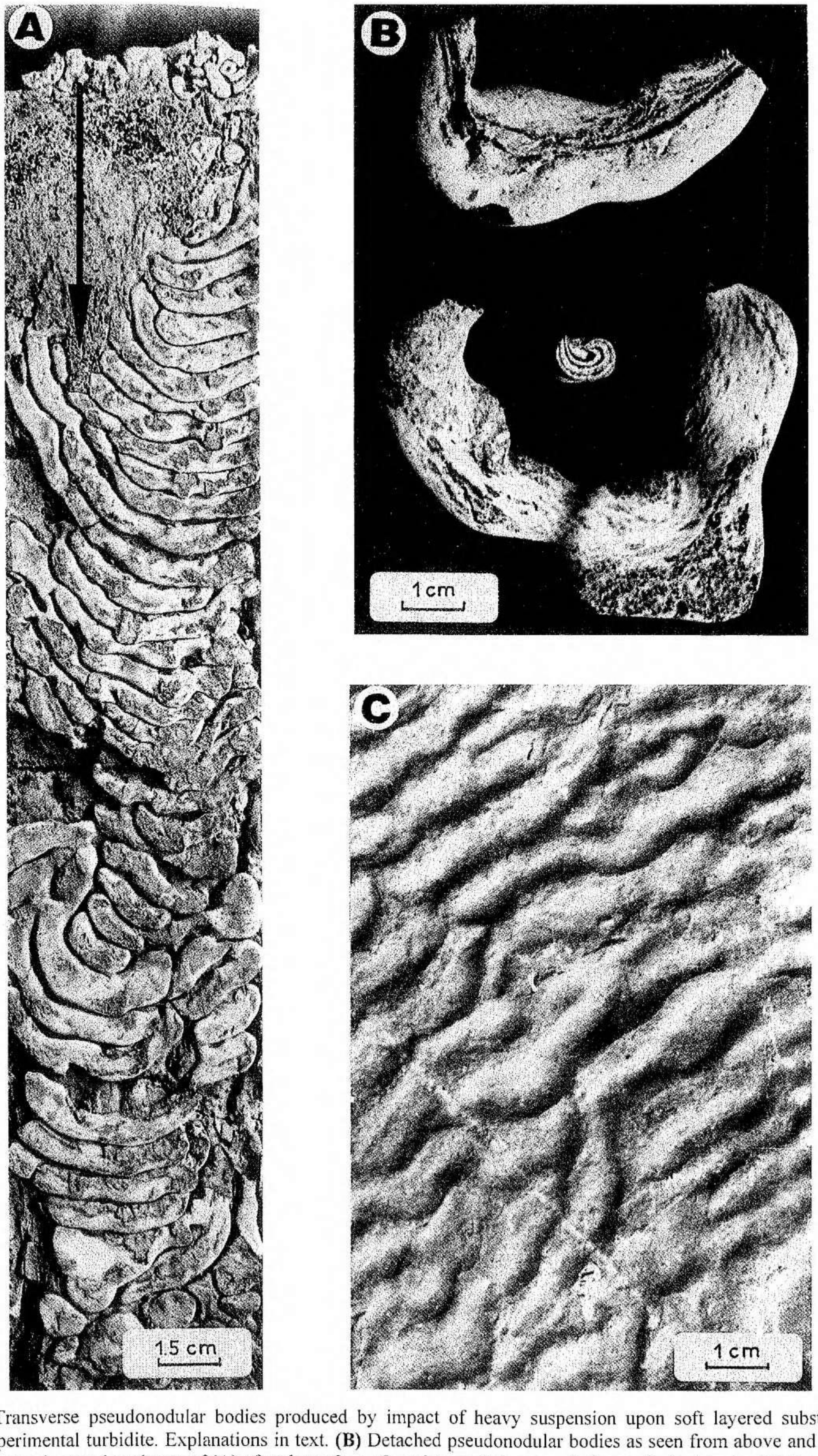


Plate 45 (**A**) Transverse pseudonodular bodies produced by impact of heavy suspension upon soft layered substratum. Base of amalgamated experimental turbidite. Explanations in text. (**B**) Detached pseudonodular bodies as seen from above and in cross-section (center). (**C**) Presumed natural analogue of (**A**) of undersurface of sandstone. Paleocene, Beloveza beds, Lipnica Wielka

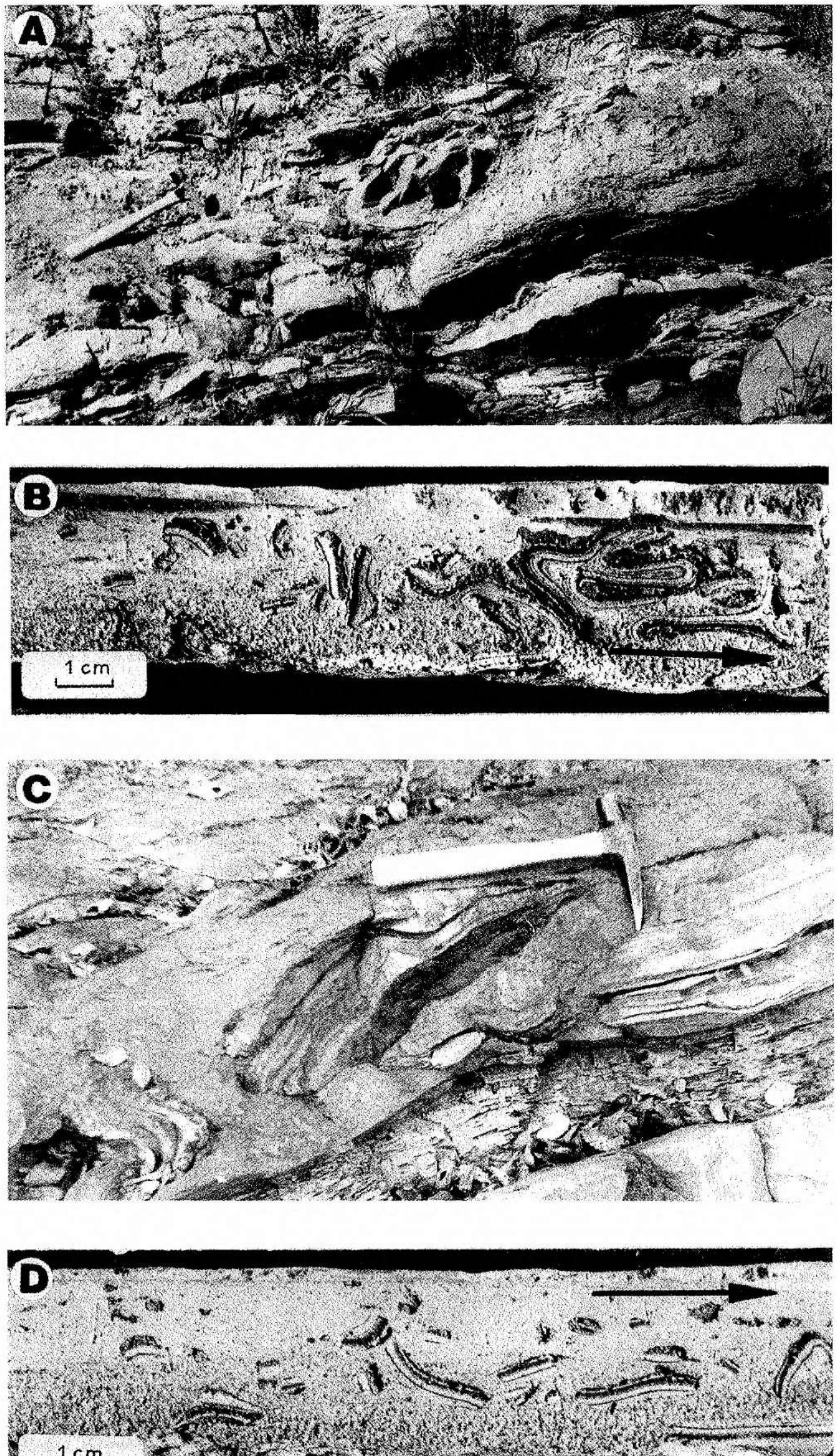


Plate 46 **(A)** Contorted pack of ripped-up layers in sediment-flow deposit. Oligocene Magura beds, Ropica Górna. **(B)** Analogous structure produced experimentally. **(C)** Angular clasts incorporated into sediment flow. Oligocene Magura beds. Ropica Górna. **(D)** Analogous structure as above produced experimentally

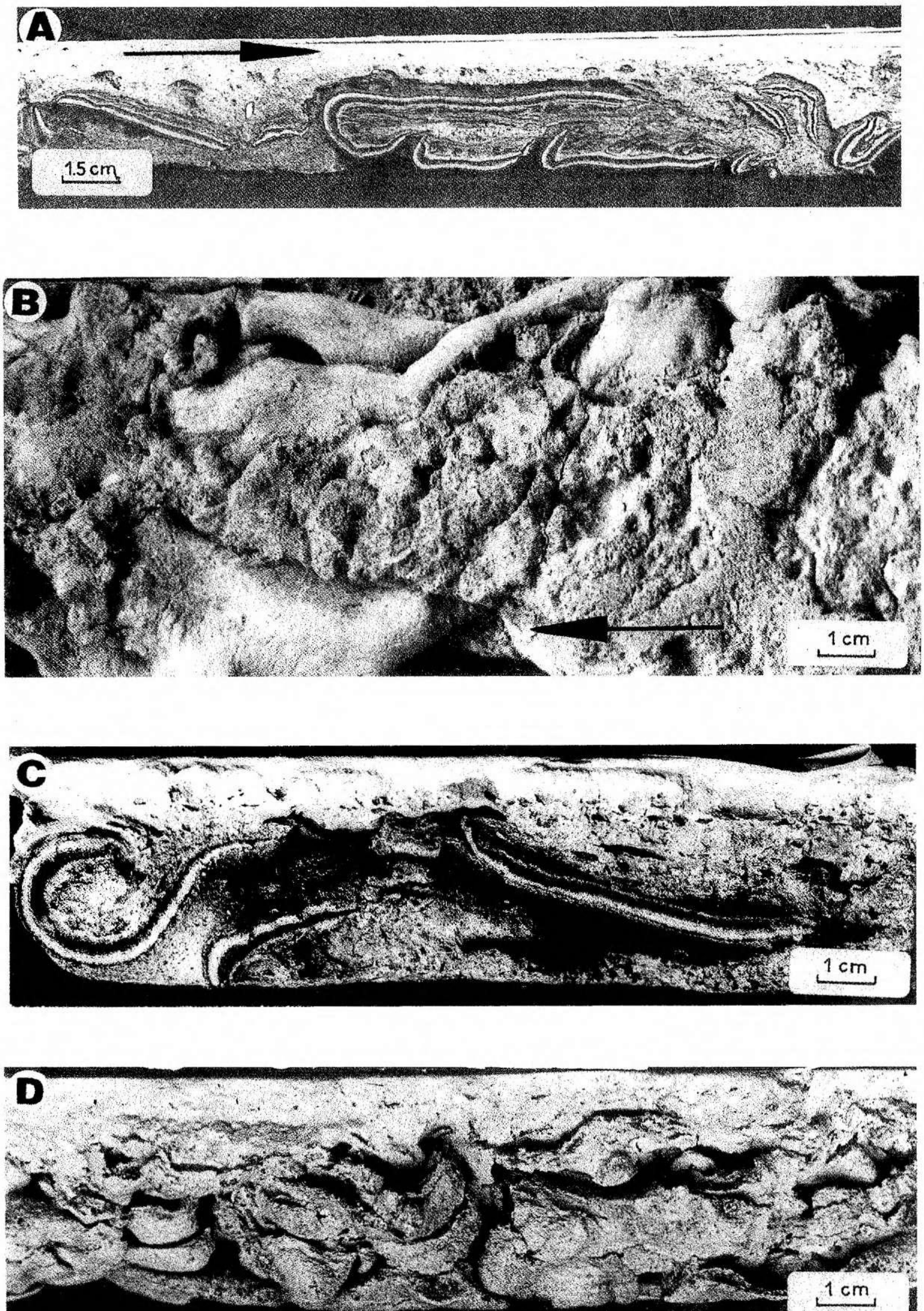


Plate 47 **(A)** Experimentally produced prolapse fold (after Dżułyński & Radomski, 1966). **(B)** Flow parallel longitudinal folds. Base of heavy sediment flow deposited upon layered substratum. Base of experimental flow deposit (after Dżułyński & Radomski, 1966). **(C)** Transverse section of experimental longitudinal folds. **(D)** Transverse cross-section of longitudinal structures produced by flow of heavy suspension upon stratified substratum. Experimental structure

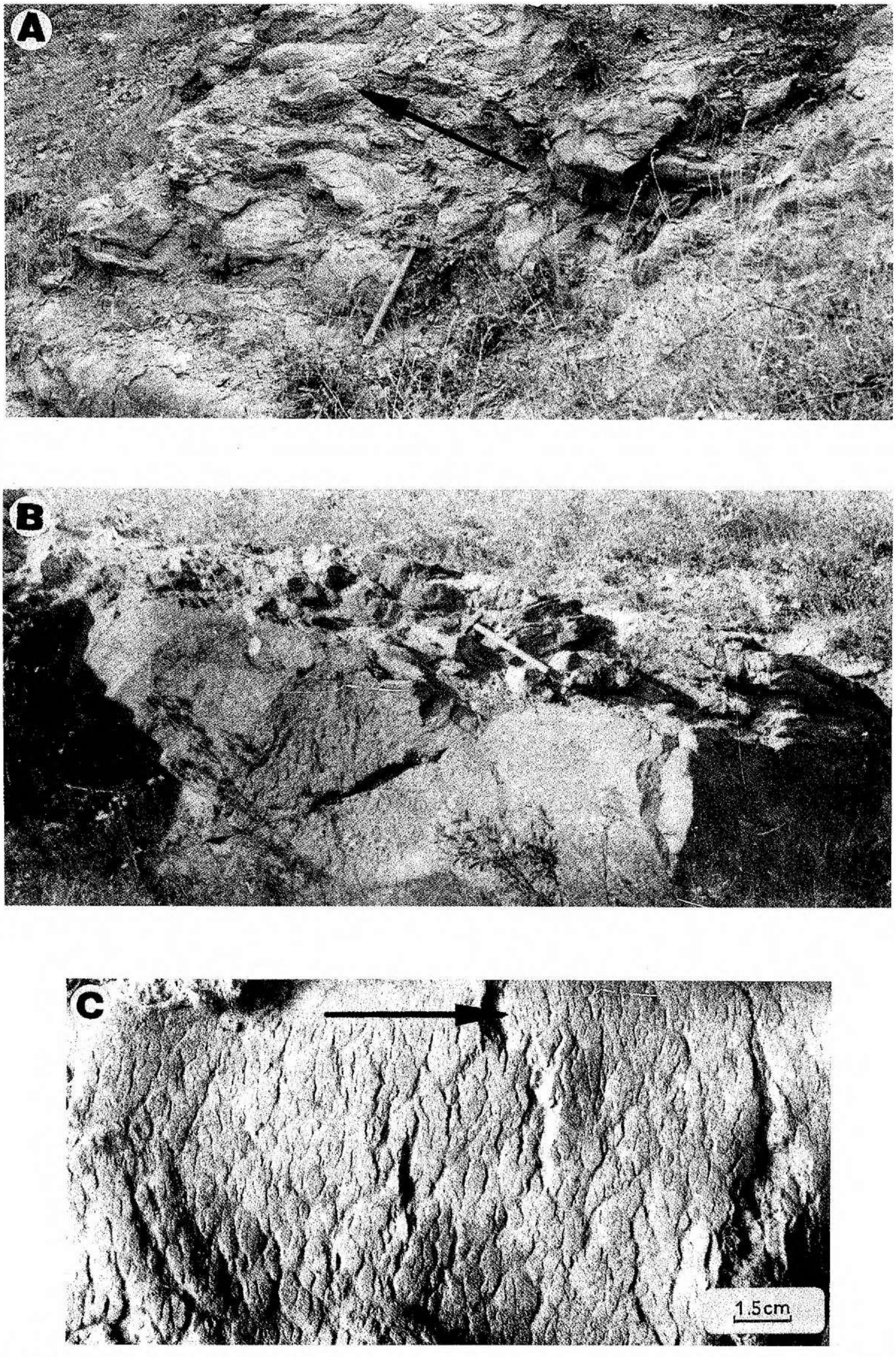


Plate 48 **(A)** Longitudinal folds in plane view. Base of overturned sandstone. Eocene Magura beds, Krynica (after Dżułyński & Radomski, 1966). **(B)** Transverse cross-section of structure shown above. **(C)** Open shear fissures produced by impact of heavy suspension upon semi-consolidated layer. Experimental structure

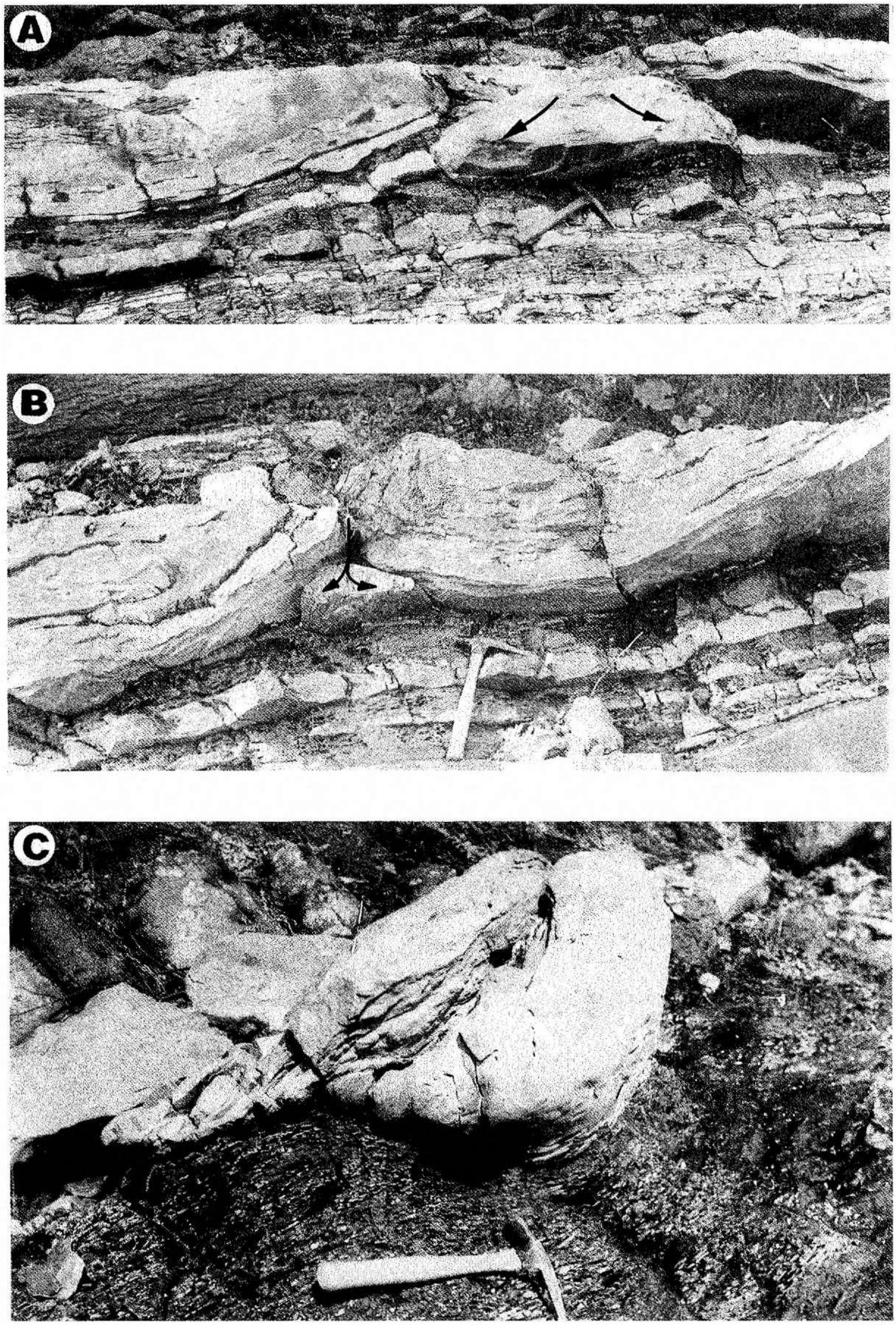


Plate 49 **(A)**, **(B)** Downward injection of haevy slurry and formation of canoe-shaped structures. Oligocene Magura beds, Ropica Górna. **(C)** Kidney-shaped clast ripped up and rised by impact of heavy ssediment flow. Corrugations on undersurface of covering bed represent moulds of sole markings and moulds of deformed top surface of clast. Cretaceous Inoceramian beds, Ropica Górna

Announcement

The new Grzybowski Foundation Library at the Geological Museum of the Jagiellonian University

The Grzybowski Foundation Library was established in 1995, with the aim of providing a central resource for researchers in the field of micropalaeontology. The GF Library is conveniently located in a research office at the Geological Museum of the Jagiellonian University, just upstairs from the Library of the Polish Geological Society. Although the Grzybowski Foundation affiliates itself and cooperates with the Geological Society of Poland, the Library is a separate institution and is housed in separate quarters. The Grzybowski Foundation has also transferred a number of geological journals and publications that were obtained through private donations to the Library of the Polish Geological Society.

The library collection contains the book and reprint collection of the late Professor Stanisław Geroch. A manuscript catalog of the *Geroch reprint collection* has been produced, and a copy is available for inspection in the Library of the Geological Society of Poland. Through the kindness of private individuals and additional purchases, the library now contains the following sets of journals:

Cushman Foundation Special Publications
Journal of Micropalaeontology
Journal of Foraminiferal Research
Micropaleontology
Marine Micropaleontology
Paleobiology
Paleoceanography
Palaeontographical Society Special Publications
Revue de Micropaleontologie
Stereo Atlas of Ostracod Shells

The GF is also in possession of a set of *Initial Reports of the Deep Sea Drilling Project*, which owing to space limitations, is housed in the Institute of Geological Sciences, Jagiellonian University.

It should be also mentioned, that during 1996-97 we expect to obtain a number of additional journals form certain persons as well as Institutions, e.g. complete set of *Micropaleontology* and *Micropaleontology Special Publications*, selected volumes of *Palaeontologica Polonica* and others. The library is also well stocked with over 150 textbooks, monographs, and occasional publications on the subject of Micropalaeontology, with special emphasis on the Foraminiferida.

The GF Library is open on weekdays from 9 AM until 3 PM, and it is equipped with a reading desk, a microscope, and a MacIntosh computer with reference data bases on the subject of Foraminifera. The GF Library is open to any researcher who is active in the field of micropalaeontology, irrespective of professional affiliation. However, book borrowing privileges are currently restricted to members of the Geological Society of Poland. Donations of all micropalaeontological literature are gratefully accepted, especially any reprints and publications dealing with Carpathian micropalaeontology. It is envisaged that the holdings of the GF Library will expand further in 1996-97, and that the library will begin to serve as a centre for micropalaeontological research in Krakow and its surroundings.

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For further information regarding the Grzybowski Foundation Library, or Grzybowski Foundation Special Publications, please contact the Curator of the Geological Museum, Urszula Mazurkiewicz, MSc, ul. Oleandry 2a, 30-063 Kraków.

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