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WEISSLIEGENDES SANDSTONES:
A TRANSITION FROM FLUVIAL-AEOLIAN
TO SHALLOW-MARINE SEDIMENTATION
(PERMIAN OF THE FORE-SUDETIC MONOCLINE)

2. A STUDY IN SIGNIFICANCE OF ROCK COLOURATION

(Pl. 1—II and 8 Figs.)

*Przejście od sedymentacji kontynentalnej do płytkomorskiej
w obrębie białego spągowca
(perm monokliny przedsudeckiej)*

2. Geneza barwy

(Tabl. I—II i 8 fig.)

Abstract. Petrographical and chemical data, together with field evidence, suggest that the red colouration of the Rotliegendes deposits developed in situ, shortly after deposition because of alteration and breakdown of iron-rich silicates and iron-oxides grains. There is no evidence that the bulk of the Weissliegendes sandstones has ever been red in the studied area.

INTRODUCTION

The present research was initiated to study the sedimentologic and environmental aspect of the Weissliegendes sandstones, a unit of white sandstones overlying the Rotliegendes deposits and being covered by the Kupferschiefer horizon and thick Zechstein evaporites (Fig. 1). The main purpose of the study is to recognize the origin of the Weissliegendes sandstones. The area under investigation lies in the central part of the Fore-Sudetic Monocline, SW Poland, where the Permian sequence is known only from the bore-hole records (Fig. 2).

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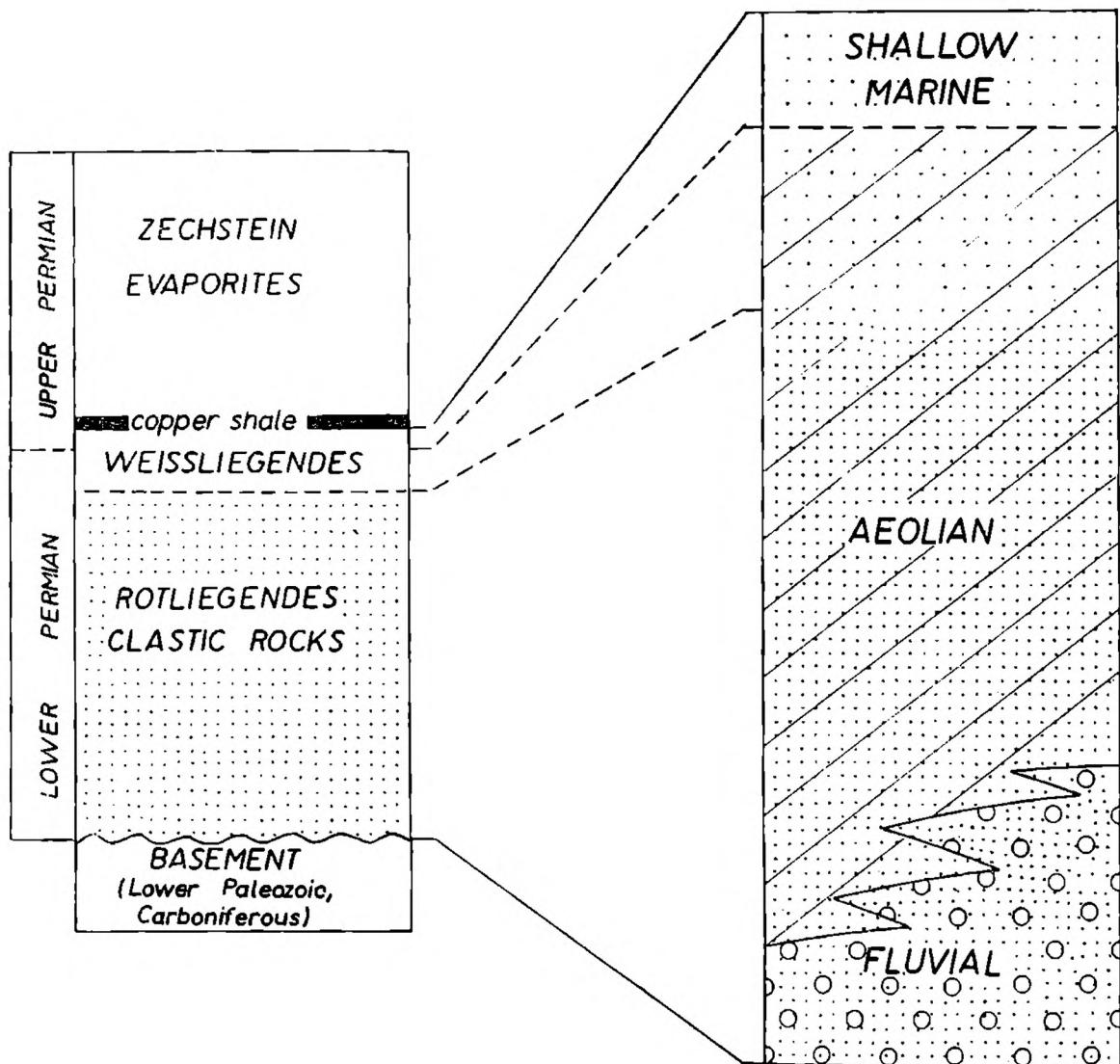
Fig. 1. General stratigraphy of the Permian, Fore-Sudetic Monocline; no scale

Fig. 1. Zgeneralizowany profil stratygraficzny utworów permskich monokliny przedsudeckiej; bez skali

In part 1 of this paper (Nemec and Porębski, 1977) the sedimentary structures and textures of the Permian clastics are examined in detail. On the basis of their vertical relationships the three environmental associations, overlying one other respectively, are distinguished within the Permian clastics: fluvial (I), aeolian (II), and shallow-marine (III) deposits. The Weissliegendes sandstones appear to be of complex origin (Nemec and Porębski, 1977). Their lower, large-scale cross-stratified portion represents the uppermost part of the aeolian association (II), while their upper portion consists of the shallow-marine deposits (i.e., association III). Within the upper part of the aeolian association there falls a transition from red coloured to white sandstones (Fig. 1).

The problem of the origin and differentiation of the rock colouration observed within the sequence has not been studied in detail by the pre-
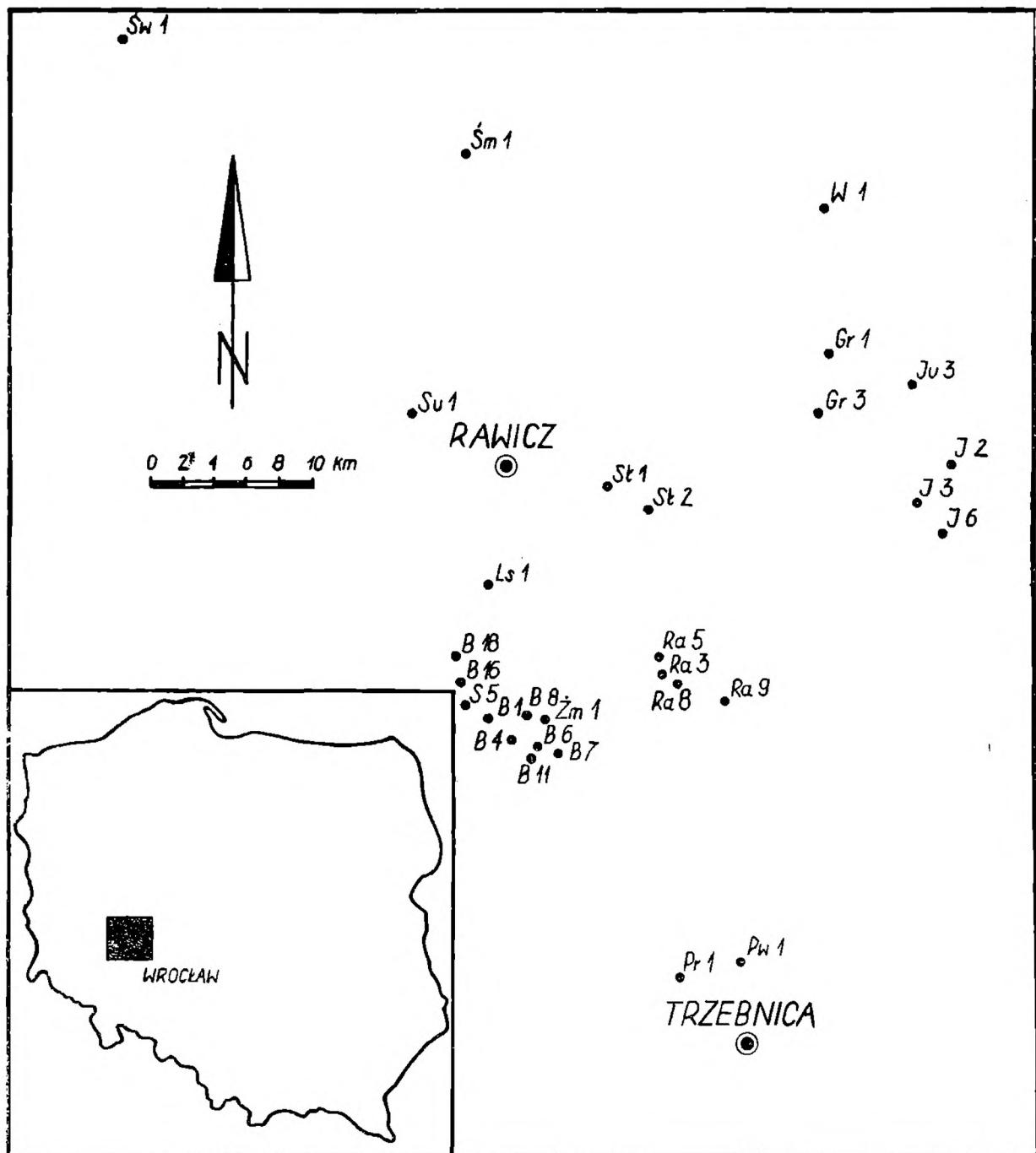
Fig. 2. Index map showing location of bore-holes

Fig. 2. Mapa sytuacyjna otworów wiertniczych

vious authors. The problem is, however, important for the three reasons at least:

(1) the Weissliegendes sandstones are distinguished, in the stratigraphic section, mainly (if not only) on the basis of their white colour;

(2) traditionally, the white colour of the Weissliegendes sandstones has been attributed to the reworking and/or bleaching of originally red sediment, the process being closely associated with the Zechstein marine transgression;

(3) the difference, in rock colouration, between the Rotliegendes red-
-beds and the overlying Weissliegendes sandstones has been attri-
buted by some authors to an important, continental/marine environ-
mental change and, if it is so, there would be also a connection
between the rock colouration and the stratigraphy.

It seems clear from the above, and from the environmental inter-
pretation presented by the authors (Nemec and Porębski, 1977), that
the problem needs explanation. Most of the present part of the paper
is therefore devoted to an examination of the origin of the rock colour-
ation. The overall aim of the investigation is to use the petrographic
and chemical data, together with the field evidence, in an attempt to
reveal any relationships which may exist between the origin of the
Permian clastics and their colour.

The results show that the red colouration developed *in situ*, shortly
after deposition of the sediments because of the alteration and break-
down of iron-rich silicates and iron-oxide grains. There is no evidence,
in the area, that the Weissliegendes sandstones have ever been red.
As a whole the situation appears to be well explained by the processes
of reddening of dune sediments with time.

PETROGRAPHY

Selected samples from measured sections of the Permian rocks were
studied in detail to reveal any vertical differentiation which may exist
in the sandstone composition. Most samples are from the association II
(aeolian deposits) and association III (shallow-marine deposits) because
this interval represents most of the Permian clastic sequence (Fig. 1)
and contains the two important transitions: a change in the rock colour-
ation and a change in the sedimentary environment (see also Nemec
and Porębski, 1977). Composition of these rocks is described below.

Main constituents

The Permian sandstones studied here show a relatively uniform mi-
neral composition, being poorly differentiated in their vertical profiles.
The sandstone samples, classified on the basis proposed by Okada (1971),
range from lithic arenites to quartzose arenites. Modal analyses of 33
red (10 R 3/4 — 10 R 7/4) and intermediate (5 YR 7/2 — 10 YR 8/2)
and 13 white (N 8) sandstones (200 points, 0·3 mm^2 spacing) are shown
in Fig. 3. Upward the individual profiles a delicate, gradational increase
in the sandstone maturity is observed (see examples in Fig. 3).

Q u a r t z. — Detrital quartz forms 40 to 50% of most samples and
normally is more abundant in the uppermost part of the profiles (up
to 70%). Compound grains formed of equidimensional to elongate crys-
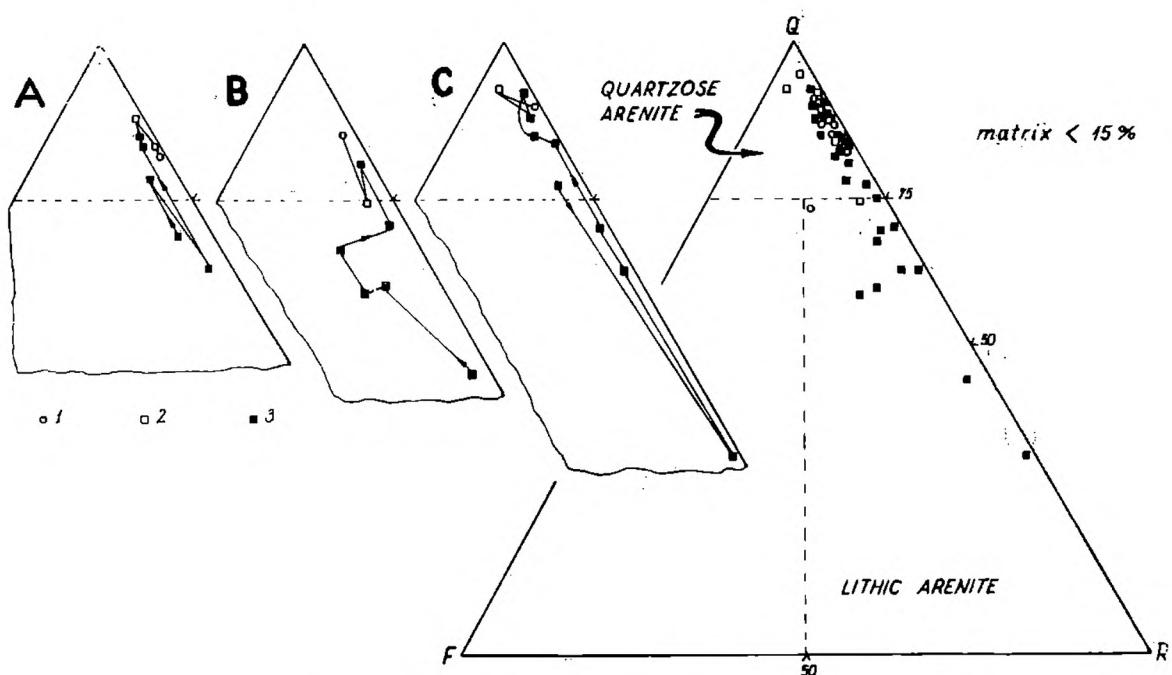
Fig. 3. Summary of modal analyses of Permian sandstones. Explanation of letter symbols: Q — detrital quartz, R — rock fragments (including cherts); F — feldspar; A, B, and C — examples of vertical change in individual profiles (Gr 1, Ju 3, and J 6 — respectively); 1 — shallow-marine ss., 2 — white aeolian ss., 3 — red aeolian ss.

Fig. 3 Modalny skład petrograficzny piaskowców permskich. Objaśnienie symboli: Q — detrytyczny kwarc, R — okruchy skał (w tym krzemionkowych), F — skalenie; A, B i C — zmienność składu piaskowców w przykładowych profilach (odpowiednio — Gr 1, Ju 3 i J 6); 1 — piaskowce płytkomorskie, 2 — białe piaskowce eoliczne, 3 — czerwone piaskowce eoliczne

tals are most abundant in the coarser sandstones, in which they commonly form as much as 20 or 30% of all the quartz. Volcanic quartz usually constitutes less than 10% of the grains and is evenly distributed throughout the section. In many samples, particularly in the uppermost ones, authigenic quartz is also present and its amount reaches 13% (5% on the average). Authigenic quartz forms either xenomorphic grains which connect, in places, detrital material, or regeneration rims around the detrital quartz grains. It often cements the individual quartz grains into a hard rock.

Rock fragments. — Rock fragments are varied and less abundant than quartz, but their amount is greater in the coarser grained sediments. The most frequent are fragments of cryptocrystalline siliceous rock (1—6%), quartzite and quartzitic schist (0·5—12%), phyllite (0—1%), and claystone (0—3%) composed of hydromicas; less regularly occur fragments of quartzose sandstone and siltstone (0—2%), granite and/or gneiss (0—1%), flow-textured porphyritic or equicrystalline rhyolites (0—1%), and a methamorphic rock similar to sericite-quartz schists (0—0·5%). Clay aggregates or pellets, subcircular in cross-section and about 0·2 to 0·4 mm across, were noted in one thin section; the pellets

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appear to be wholly sericite, but some are very dark grey under cross nicols and appear to be very fine quartz and feldspar. Such dark grey pellets are, however, extremely rare in the samples examined.

F e l d s p a r s. — These occur in every thin section, and their content generally widely ranges between 0·1 and 15%, being higher in coarser grained sediments, and decreases upward the profiles (Fig. 3). They are represented by plagioclase (albite, oligoclase) and potassium feldspar (microcline).

O t h e r   d e t r i t a l   m i n e r a l s. — These include scattered grains of detrital mica and accessory heavy minerals. Detrital mica, mostly moderately bleached and/or chloritized biotite (green in places), is common in the basal beds (fluvial deposits) but is rare in the overlying aeolian and marine deposits. In some sections both biotite and muscovite occur, the latter being always present in smaller quantity, while in some samples chlorite is the dominant detrital flake-mineral. Chlorite flakes present have a habit similar to that of the mica. In the red deposits the cleavage and margins in many biotite and chlorite flakes are altered to red ferric pigment, the cleavage being also distorted by squeezing between sand grains.

Heavy mineral content of the sandstones is very low. The heavy--mineral assemblages consist mainly of zircon, tourmaline, and opaques (mainly magnetite); in smaller quantities occur garnet, kyanite and staurolite, while rutile and hornblende are very rare. The heavy-mineral association is discussed by the authors, in more details, in a separate paper to be published later (Nemec et al., 1978).

M a t r i x. — The matrix content in most of the sandstones is usually only 2 to 4%, being even lower in the aeolian and shallow-marine sandstones (associations II and III) which generally almost lack matrix material (see also Jerzykiewicz et al., 1976). Its content is higher in the fluvial deposits, but never exceeds 15% (arenites; see Fig. 3).

C e m e n t. — The content of the rock cement ranges from 15 to 40%. Three kinds of the cement were distinguished: clay cement (0—20%), carbonate cement (0·5—20%), and quartz cement (0—13%); anhydritic cement appears to be less frequent and it occurs in a lesser amount (0—2%). Most of the sandstones contain both quartz and carbonate cements. Volumetrically the latter is more important, occurring as isolated patches and/or replacing both quartz and feldspar grains. Quartz cement typically occurs only in small amounts, being in an optical continuity with the detrital grains. The relationship of the timing between the precipitation of the quartz and calcite cements is often uncertain but in some instances calcite is replacing quartz overgrowth which indicates that probably both quartz and calcite cementation took place at an early stage in diagenesis, perhaps after or during compaction, but at a time when the sands still had a significant porosity. The solution which

brought the quartz cement has had a destructive influence on some of the detrital grains, but also caused secondary overgrowth. The dissolved quartz is thought to have been derived mostly from the sediment itself, this being drawn from the observation that a transition from weathered quartz (-ite) fragments to a silt-sized cement occurs locally. In contrast to the uppermost, shallow marine sandstones, the underlying sediments show poorer cementation and a significant portion of their porosity is even not eliminated (see also Alexandrowicz and Słupczyński, 1970, p. 703).

Vertical relations

As an aid to study the possible vertical differentiation in the mineral composition of the aeolian and shallow-marine sandstones, their two selected profiles were examined petrographically in detail. The results obtained (Figs. 3 and 4) show that there are some slight differences between the two sandstone associations in the content of individual constituents. For an objective comparison, eight additional profiles were examined using the six categories: mean content of detrital quartz (1) and rock fragments (2), and mean content of quartz (3), carbonate (4), clay (5) and anhydrite (6) cements. The values obtained were next tested, using Student's t-test, against the hypothesis that they are the same in the red sandstones of association II, white sandstones of association II, and in the sandstones of association III. The results are summarized in Table 1. In most instances the null hypothesis was accepted at the 95% confidence level. At the latter level the hypothesis was rejected only in the two instances: for comparison of the white sandstones of association II and the sandstones of association III on the basis of their carbonate cement content and their clay cement content. For all the three sandstone groups the null hypothesis was rejected only in the case of comparison of the anhydritic cement content.

From the data above it becomes clear that the Permian sandstones show poor differentiation in their mineral composition. The only significant difference observed is that in the carbonate and clay cements content; namely, between the sandstones of association III (content significantly greater) and the white sandstones of association II (content significantly lesser). Except for the anhydritic cement content, the latter sandstones do not significantly differ from the remaining (i.e., red) sandstones of the aeolian association. The higher cement content in the shallow-marine sandstones may be explained as being partly due to the washing phase of the sediment redeposition (see Nemec and Porębski, 1977), associated with the winnowing of fines, in the nearshore environment of its final deposition. As such it may be attribued to the originally higher porosity of this shallow-marine deposit. The other conclusion
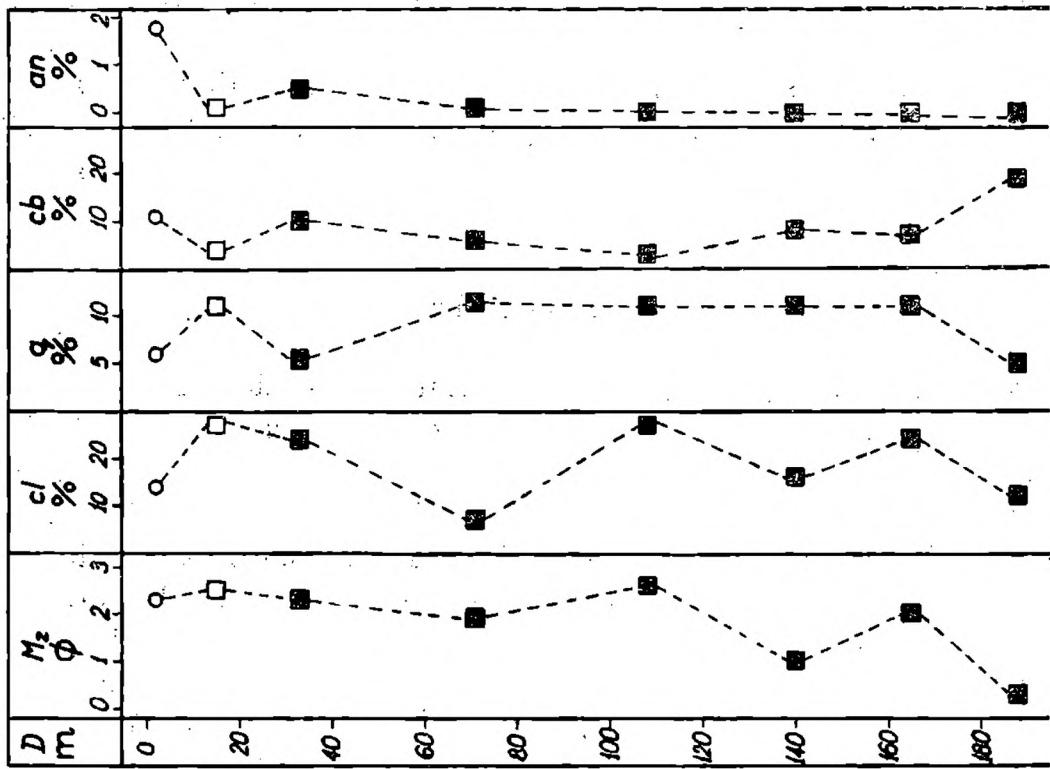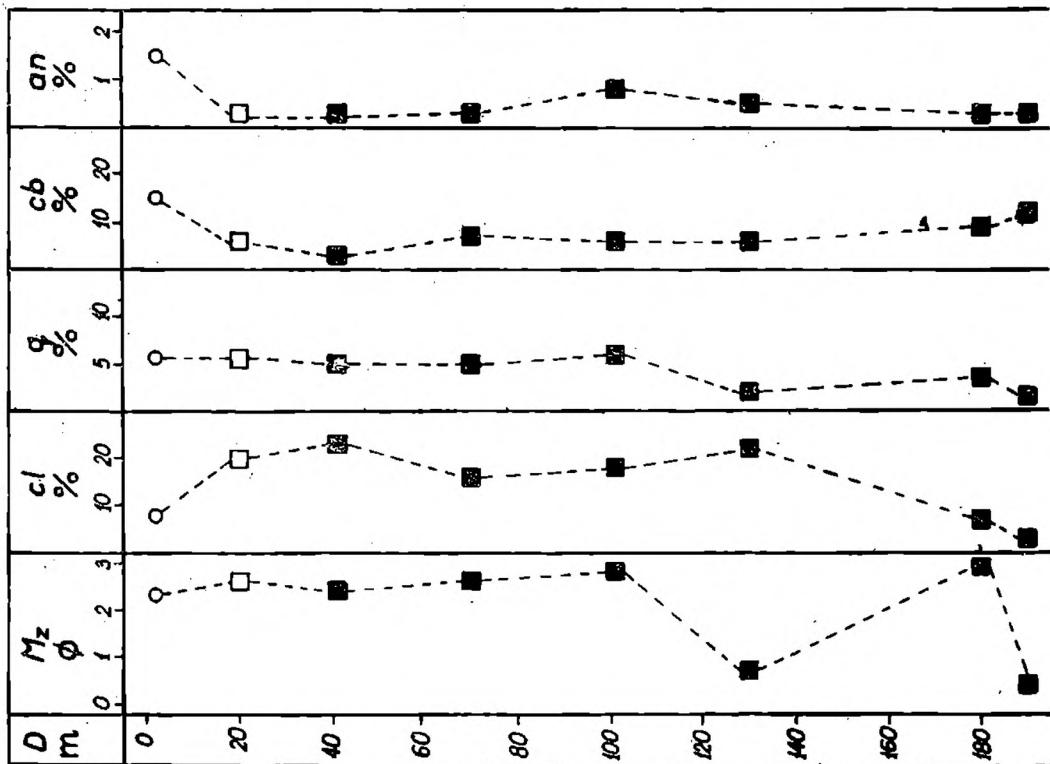
Summary of results of modal analysis of 23 sandstone samples Table 1

Skład petrograficzny piaskowców na podstawie 23 próbek   Tabela 1

Sandstone constituent	Group III: shallow-marine, white sandstones 7 samples		Group IIb: white aeolian sandstones 10 samples		Group IIa: red aeolian sandstones 6 samples	
	mean /%/	variance	mean /%/	variance	mean /%/	variance
/1/ Detrital quartz	59.1	54.3	60.9	99.8	59.6	100.5
/2/ Rock fragments	8.7	3.8	9.1	8.4	9.4	10.7
/3/ Quartz cement	8.2	14.7	9.2	12.6	6.1	4.1
/4/ Carbonate cement	13.2	86.1	4.8	6.4	2.4	5.9
/5/ Clay cement	8.6	15.7	15.7	37.0	16.7	25.7
/6/ Anhydrite cement	0.6	0.4	0.1	0.1	0.3	0.1

Student's t - tests for differences in mean contents of mineral constituents in grouped sandstone samples

Sandstone groups compared	Calculated t - values /^{x}/ Sandstone constituents					
	/1/	/2/	/3/	/4/	/5/	/6/
III vs. IIb	0.379/NS/	1.104/NS/	0.492/NS/	2.581/S/	2.787/S/	5.860/S/
IIb vs. IIa	0.241/NS/	0.202/NS/	1.672/NS/	1.693/NS/	0.335/NS/	4.000/S/

/^{x}/ /NS/ - not significant   /P>0.05/

/S/ - significant

from the data is the increasing-upwards maturity, both mineralogical and textural (see also Nemec and Porębski, 1977), of the Permian sandstones studied here.

Clay mineralogy

Fifteen samples of red, intermediate and white sandstones, representing the sandstone associations II and III, were analysed for clay minerals in order to find any possible relationships which may exist between clay mineralogy and rock colouration, and/or depositional environment.

Fig. 4. Cement content in Permian sandstones; A — profile Gr 1, B — profile J 6. Explanation of letter symbols: D — depth below the Kupferschiefer (in meters); Mz — mean size (optically determined); cl — clay cement; q — quartz cement; cb — carbonate cement; an — anhydrite cement; for other symbols see Fig. 3.

Fig. 4. Zawartość spoiwa chemicznego w piaskowcach permskich; A — profil Gr 1, B — profil J 6. Objaśnienie symboli literowych: D — głębokość poniżej łupków miedzionośnych (w metrach); Mz — średnia średnica ziarna (ustalona optycznie); spoiwa: cl — ilaste, q — kwarcowe, cb — węglanowe, an — anhydrytowe; graficzne oznaczenia punktów jak na fig. 3

— 522 —

S. 2
sw 340
TG 200

S. 6
sw 750
TG 100

S. 5
sw 940
TG 100

S. 4
sw 650
TG 100

S. 15
sw 320
TG 200

S. 16
sw 660
TG 100

S. 18
sw 930
TG 100

S. 35
sw 830
TG 200

S. 37
sw 730
TG 200

S. 13
sw 650
TG 100

S. 41
sw 760
TG 100

S. 11
sw 450
TG 100

S. 12
sw 660
TG 100

S. 8
sw 750
TG 200

S. 9
sw 840
TG 200

0 100 200 300 400 500 600 700 800 900 1000
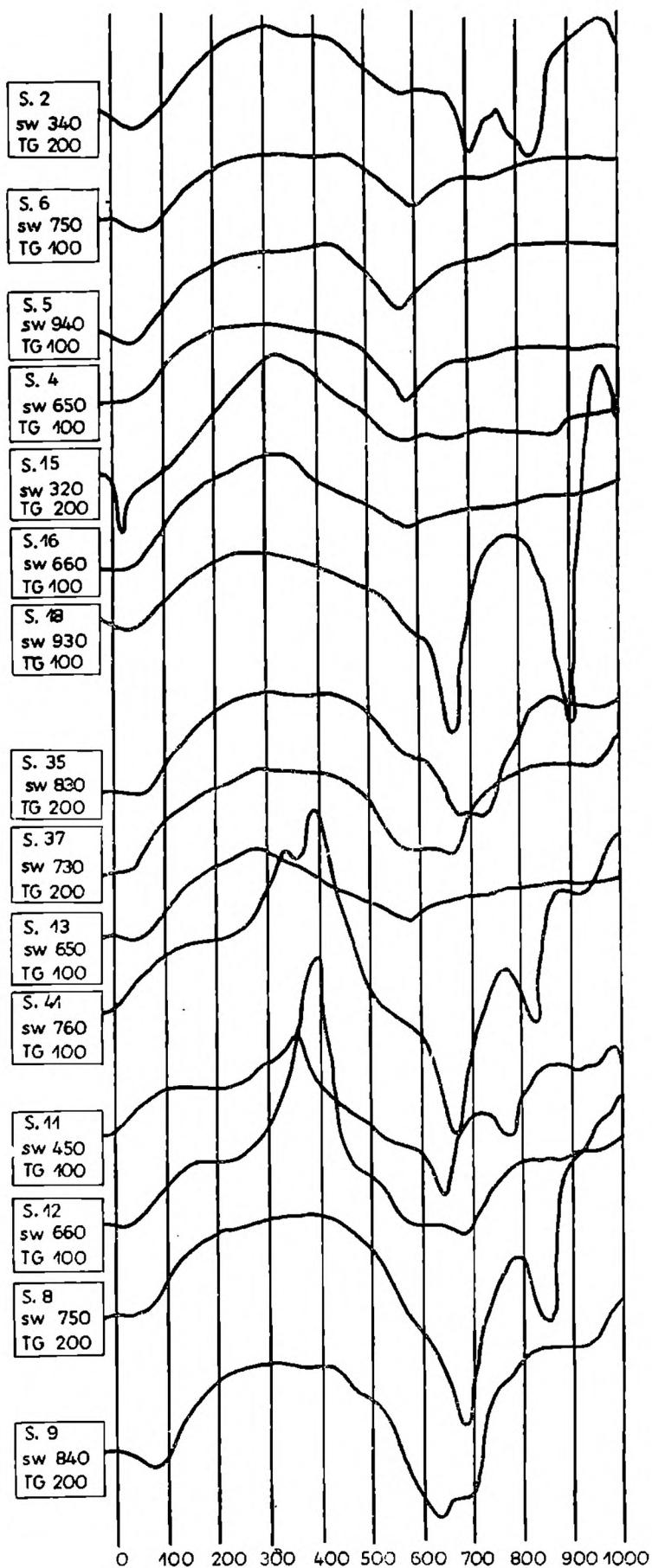
The fresh samples were disagregated by gentle wet grinding and < 0.062 mm separations were then made in usual way. Qualitative determination of clay minerals was accomplished by means of differential thermal analysis (DTA). All the samples were run at almost identical laboratory conditions: heating rate of 10° C/1 minute, DTA and DTG sensivities of 1/5 and 1/5 respectively, TG sensivity equal to 100 or 200, sample weight ranging from 320 to 940 mg (684 mg on the average). All the samples were dried 0·5 hour at about 105° C before the DTA (see e.g. van der Marel, 1956, p. 22).

Although neither the DTA method nor the grain-size fraction used give a sufficient basis for obtaining exact data, the DTA results are generally thought to be satisfactory for comparative purpose (e.g. van der Marel, 1956; Stoch, 1974).

Results od clay DTA analysis   Table 2

Skład mineralny frakcji ilastej oznaczony metodą DTA   Tabela 2

Sandstone association	Sample no.	Colour	Montmo-rillonite	Kaolini-te	Illite	Chlorite	Other traces
III	2	white	x?			y	
III	35	white		x?		x	
III	11	white	x?	x	x	x	org. subst.
III	8	white		x		x?	
III	41	white		x		x?	org. subst.
III	13	white			x		
II	15	white		tr	tr		org. subst.
II	16	white			x?		
II	4	v.light pink			x		
II	5	pink			x		
II	6	light red			x		
II	12	red		x			org. subst.
II	37	red		x			
II	18	red		x		tr?	
II	9	red		x	tr		

The only clay minerals identified are illite, kaolinite and chlorite; montmorillonite was questionably identified in two samples. Some of the samples contain traces of organic matter (perhaps bitumens). Figure 5 shows the DTA curves and the results obtained are summarized in Table 2.

Fig. 5. DTA curves. Explanation of symbols: s — sample number (see Table 2): sw — sample weight (in miligrams); TG — TG sensitivity

Fig. 5. Wykresy DTA. Objaśnienie symboli: s — numer próby (zob. tabela 2); sw — waga próby (w miligramach); TG — czułość TG

No significant relationships between clay mineralogy and rock colouration were found in the samples examined, this being in agreement with many studies of, environmentally associated, red and non-red (white or grey) sediments (see e.g. Friend, 1966; Turner, 1974). Although such a relationship might have existed, with certain probability, within the sequence but, even if it was present, now it is largely obscured by later, long-lived diagenetic processes (see also Turner, 1974, pp. 230—231).

As regards the sedimentary environment of the Permian sequence, no clear relationships were found between the clay mineralogy and the sandstone environmental associations considered. In some of the samples of the shallow-marine sandstones (association III), however, chlorite has been identified while it is rare in the clay fraction of the underlying aeolian sandstones (association II). At the present stage of the investigation it is difficult to say whether the presence of chlorite in these marine sediments is environmentally sensitive. Such a possibility may be suggested, however, due to the work of Grim (1958), Milne and Earley (1958), Weaver (1958a, b), Jacobs (1970, 1974), Siever and Kastner (1972), Eugster and Chou (1973), Leone et al. (1975). On the other hand, the depth of burial and the DTA restrictions and exactness may be also the factors. Additional studies of the clay minerals from the Permian sequence are needed in order to prove or disprove this suggestion.

THE PROBLEM OF ROCK COLOURATION

The present part of the paper is a kind of revision of the previous opinions on the problem. The study is an attempt to examine, in detail, the variation in the rock colouration observed and to interpret it from the point of view of the sedimentary environment of the Permian sequence. In contrast to many previous investigations, the origin of the white colour of the Weissliegendes sandstones has been studied here in the context of the origin of the underlying red beds which should be, logically, a clue to the explanation of the whole problem.

Red beds are typified by the European Lower Permian but are common at all levels of the stratigraphic column. Although they are most commonly terrestrial clastic deposits, they include also a wide variety of marine or deltaic facies so it is reasonable to assume that a variety of mechanisms were ultimately responsible for their formation, and that conclusions should only be made about specific examples.

Some related studies

The colouration of ancient red beds is usually caused by red ferric oxide pigment in the form of hematite (α—Fe_2O_3). Recently, in moist tropical areas and hot deserts much ferric iron is present in the soils

as tan to brown amorphous iron hydroxide or red oxide. However, the nature of the processes which might produce reddness has been the centre of much controversy and remains a discussed problem.

Three main lines of interpretation can be distinguished in the literature:

1) traditionally, geologists have envisaged red beds being formed by derivation of detrital hematite directly from deeply weathered, red tropical lateritic soils (latosols), as suggested by Raymond (1927), Wahlstrom (1948, p. 1175, 1188), Krynine (1949, p. 65, 67; 1950), Hubert (1960, pp. 221—234), and Van Houten (1961);

2) the pigment may be diagenetic hematite that has converted *in situ* from detrital yellow or brown iron oxides derived from deeply weathered, but not necessirily red, soils, as suggested by Van Houten (1964, pp. 658—659; 1968, pp. 404—405; 1972);

3) alternatively, the red pigment may be authigenic hematite that has formed diagenetically owing to *in situ* alteration of iron-bearing, detrital silicate minerals, as suggested by Walker (1967a) and Walker et al. (1967);

4) the pigment may be also a product of a combination of two or more of these processes.

The first of these hypotheses is not applicable to the Lower Permian rocks because, as indicated by numerous authors, no evidence exists that red beds are transported and accumulating as red detritus from source areas today (e.g. Gray, 1967; Walker, 1967a, b; Van Houten, 1972), and there is no evidence, in the present case, to suggest that lateritic soils were available for transport from the source areas. Picard (1965, p. 471) states that secondary (second cycle) red beds are characterized by (1) interbedded rocks of red and drab colours, (2) a genetic relation to known previously existing red beds, (3) conglomeratic facies being dominant in the sequence, and (4) restriction of particular colours to genetic units of deposition. The Permian sedimentary sequence, being under the study, does not exhibit any of the characteristics. For similar reasons the second theory is rather unlikely here and it is argued below that the red colouration of the investigated Lower Permian rocks is best explained in terms of the third hypothesis.

Evidence reported by Walker (1967a, b; 1973), Walker and Ribbe (1967), Walker et al. (1967), and Walker and Honea (1969) from the subsequent studies in the Sonoran Desert in Baja California, Mexico, and in desert regions in North Africa and the Middle East, demonstrate that hot, dry climates are particularly favourable for the formation of red beds because in such climates the Eh — pH of the interstitial environment both above and below the water table are likely to favour the formation and preservation of hematite. The same works have also strongly suggested that weathering of hornblende and biotite are an

important, and perhaps the most important (Norris, 1969, p. 10), source of iron in sediments. It was also pointed out by Walker (1967a) that perhaps in all ancient red beds associated with either evaporites or extensive aeolian sandstones the pigment has formed diagenetically in sediments that were deposited in arid or semiarid climates. Walker and co-authors (see above) have provided some of the most carefully documented evidence for the development of the red sediments and concluded that reddening can and does take place in both hot dry and hot moist climates.

Walker's hypothesis has been widely supported by other investigators (e.g. Norris, 1969; McBride, 1974; Turner, 1974; Schluger, 1976). Although presented as possible alternatives to that of Walker, the interpretations recently offered by some authors (e.g. Friend, 1966; Blatt et al., 1972; Turner, 1974) do not reject Walker's hypothesis but rather tend to support it; in situ crystallization of iron hydroxide under oxidizing conditions and the post-depositional alteration of iron-silicates and oxides were recognized to be responsible for the formation of red beds in all these situations, and the differences in oxidation/reduction potential have been explained in terms of the position of the water table. It cannot be concluded from the data presented by Thompson (1970) and Blatt et al. (1972, p. 368) that Walker's hypothesis is unsatisfactory with respect to the red beds studied by these authors (see Turner, 1974, p. 218).

The role of water in the alteration processes has been also a subject of some controversy. But although Schmalz (1968) has argued on theoretical grounds that hematite could not form stably in the presence of liquid water, most of the authors fully agreed with and imply that water (or moisture at least) is necessary to bring about the diagenetic changes resulting in the reddening of desert sediments (Dunham, 1953; Alimen, 1957; Feys, 1964; Walker, 1967a; Berner, 1969a; and numerous others).

A full summary of recent research on red beds has been presented by Glennie (1970, pp. 171—193). In the present paper only some of the main studies have been reviewed.

Field evidence

The basal fluvial deposits (Fig. 1), as well as most of the overlying aeolian sandstones, exhibit rather uniform, reddish brown colouration. The latter sandstones show however, in their uppermost portion, a change in their colouration, from reddish brown (10 R 3/4) and moderate orange pink (10 R 7/4) to greyish orange pink (5 YR 7/2), very pale orange (10 YR 8/2) and light grey (N 8). The light grey sandstones, usually referred to as „white", are uppermost and correspond to the

lower portion of the Weissliegendes sandstones, i.e., alpha-type sandstones of Jerzykiewicz et al. (1976) in the Lublin mining district; these are overlain by the shallow marine sandstones showing a similar, light grey, colour. The transition zone varies in thickness from profile to profile (see Nemec and Porębski, 1977, Fig. 3), the colour change being usually gradational and only sporadically sharp. In some of the profiles, however, an abrupt change was observed within the red-coloured sandstones, from brick red (in the lower portion) to reddish brown (in the upper portion), this being similar to the evidence reported by Price (1962) and Norris (1969) from modern desert dune fields and related to the dune age.

In one of the profiles examined (Pw 1, Fig. 2; see also Nemec and Porębski, 1977, Fig. 3) the colour change is not uniformly gradational. Although the aeolian sandstones are uniformly reddish brown in their lower part (about 3/4 of the sequence), their upper portion (about 14 m thick) exhibits purple red streaks and spots present on the red background. Upwards, the red-colour background changes to pink and very pale pink, this being accompanied by a similar change in the colour of streaks and spots, i.e., from red to very light red (see also Nemec and Porębski, 1977, Pl. I, figs. 1—3). In their uppermost part (several meters thick), these cross-stratified sandstones show uniform, very pale pink to light grey, colouration and no spots or streaks are visible. They are overlain by the unit of light grey, shallow-marine sandstones (about 0·5 m thick). The red streaks and spots, mentioned above, are usually parallel to the sandstone stratification, their foci being concentrated along the coarser, sandy laminae. The evidence above suggests incomplete reddening of the deposit rather than a bleaching phenomenon, the spots being probably the result of the local redistribution of ferric iron due to the ground-water fluctuations. There are no signs of iron concentration (spots, nodules, etc.) near the top of the non-red sandstones which might have suggested the presence of bleaching processes (see also Berner, 1969b) in the area.

Petrographic evidence

Petrographic data concerning the origin of the red pigment in the Permian red beds, being under the study, have been collected from an analysis of 45 thin sections from core samples. Thin section data indicate that the colour of the Lower Permian red beds is due to hematite, and/or geothite in places, pigment which occurs in a variety of textural sites. Particular grain coatings of very finely crystalline hematite occur on grains both in red sandstone and mudstone but are absent from the white sandstone. Plate I, fig. 1, shows a typical red sandstone with hematite pigment.

In most of the sections hematite pigment typically appears as stain both on matrix material and framework grains, therefore it is difficult to determine clearly whether the detritus was stained prior to or after deposition. All the sections contain red pigment, but less than one third are useful for consideration of its origin and time of formation.

The iron-rich silicates present in the Permian sedimentary sequence are biotite and chlorite; the others, such as hornblende or epidote, are extremely rare. Their content is generally very low in the deposits, but relatively higher in the white sandstones. The content seems also to be directly proportional to the degree of preservation (i.e., unalteration) of the mineral grains, decreasing with the intensity of the sandstone reddening.

In the studied red sediments biotite and chlorite show a variety of stages of replacement by hematite, from totally fresh grains, to almost complete replacement. Characteristic stages of this alteration are: (1) reddening of cleavages and grain margins; (2) development of opaque hematite along cleavages; (3) formation of densely packed, irregularly shaped, opaque hematites. Some of these textures are shown in Plate I. Even where biotite or chlorite is totally replaced by hematite, this is easily recognized because they retain the flake-like phyllosilicate form, and the same distortion due to compaction, as the unaltered grains (Pl. II, fig. 1). In some sections of the fluvial sandstones all the flake--like biotites (and chlorites or muscovites, if present) show an apparent arrangement parallel to the sandstone stratification, and in such a case the hematite flakes also occur in the same position (Pl. II, fig. 1); moreover, the successive stages of biotite-to-hematite or chlorite-to-hematite alterations, such as described above, can be also observed here.

The fact that biotites are scarce in intensively red sandstones, whilst hematite flakes are relatively frequent, suggests that hematite largely replaced biotite in these rocks. Relatively frequent occurrence of biotites in the intermediate (light red, pink) sandstones and their presence in white sandstones show that the oxidizing conditions necessary for their alteration did not persist after burial. In the intermediate sandstones, biotites show a varying degree of alteration in any section, while they are almost fresh in white sediments.

Hornblendes, the more unstable iron-bearing detrital grains, are very rare in the studied red sediments. Fresh hornblende crystals are preserved however, in places, inside larger quartz grains. When occur in red sediment, they are normally etched and densely hematite stained, and/or peripherally altered to authigenic clay (Pl. II, figs. 2—4). In the latter case, the accompanying released iron is often still preserved in a form of opaque oxides or hematite „halos", surrounding the parent grain or clay aureola. The extensive character of the alteration indicates, as in

the case of hornblende, biotite and chlorite, that it has taken place intrastratally rather than at the ground surface (see also Walker, 1967a). Similar minute-grained hematite, indicating post-depositional formation of red pigment, occurs also on and/or around fragments of volcanic rock and grains of magnetite. All detrital iron-bearing grains, therefore, probably have contributed in some degree to the source of iron in the hematite pigment.

Paucity of hornblende in the red deposits, where it should be a relatively common accessory mineral (known in the source area and in the non-reddened deposit), suggests that it has been largely destroyed by intrastratal alteration. As shown by many authors, hornblende (and augite) is a common accessory mineral in modern fluvial sediments that are derived from igneous and metamorphic terrain, even where source areas are characterized by hot moist climate and deep weathering (see Walker, 1967a, pp. 365—366). Moreover, decomposition of hornblende in the red sediments forming today is a well-known fact (see Walker, 1967a).

The occurrence of mineral grains with varying degrees of hematization suggests that the alteration (i.e., oxidation) might have been partly pre-depositional. This feature could represent the mixing of detritus which has suffered different degrees of weathering, as suggested by Friend et al. (1963) and Friend (1966) for the Devonian red beds of Arran. However, there is no real evidence of material derivation from upland red soils to the Lower Permian basin. If, as considered in the literature (e.g. Dziedzic, 1961; Glennie, 1970, 1972; Brongersma-Sanders, 1969), the Lower Permian climate in Europe was regionally semi-arid to arid, then weathering in the source area must have been negligible. Probably the weathering might started within the basin, and the later redeposition of sediment (with continuous contribution of fresh material) provoked the differences in the degree of post-depositional alteration. Alternatively, most of the weathering might have taken place only during diagenesis, for interstitial Eh — pH conditions may vary considerably and there is no reason to expect a uniform alteration (see also Walker, 1967a, pp. 361—363; Schluger and Roberson, 1975; Schluger, 1976).

In some sections of the Permian red beds the hematite occurs as small grains and/or euhedral crystals which in places replace the periphery of detrital silicate grains (Pl. II, figs. 5—6). The presence of similar hematite crystals have been reported by Walker (1967a, Pl. 2, fig. 5) from the Late Paleozoic red beds of Colorado; Walker states that „such crystals clearly have formed *in situ* and demonstrate that the interstitial environment was favourable for the formation of hematite after deposition of the sediments" (Walker, 1967a, p. 366). Large

irregular patches of hematite (Pl. II, fig. 1) occupying intergrain position could be grains of hematitized shale which have been crushed during compaction (Turner, 1974, p. 221) or, alternatively, some of these grains could be „remobilized" detrital hematites, i.e., hematite grains which have changed their shape under the pressure of compaction (Picard, 1965). In places, the hematite grains were observed to form in the matrix material, due to extensive replacement of the clay (particularly chlorite-rich) matrix; in such instances the hematites often contain small inclusions of only partly replaced and/or unreplaced matrix material (Pl. II, fig. 2).

Biotite and chlorite, particularly minute-grained, are abundant in red mudstones and shaly siltstones, and they have provided perhaps the major source of iron for the hematite pigment in these lithologies. Red shaly siltstones also occur as intraformational flakes in coarser-grained rocks. Robb (1949), Walker (1967a), and Turner (1974) also have called attention to the importance of biotite as a source of iron for authigenic hematite, particularly in fine-grained red beds. On the other hand, the presence of reddened siltstone fragments, and some hematite-stained mineral grains, sticking in poorly reddened argillaceous matrix indicates that their partial alteration to hematite (or to geothite or limonite, at least; see Walker, 1967a, pp. 360—361) might have taken place prior to their final deposition, or rather redeposition. In either case an early diagenetic alteration is probable. The mineralogy of the red ferric oxide pigment in the Permian rocks is often difficult to determine. Inasmuch as even these red beds show only traces of hematite, it seems probable that the pigment is often not hematite, or at least not well-crystallized hematite. The pigment seems to be, in places, some form of ferric oxide that either is amorphous or too poorly crystallized to give diagnostic optical features or X-ray diffraction pattern and includes forms that may be intermediate between yellow-brown, amorphous ferric hydrate (limonite?) and red, well-crystallized hematite. The combination of types of iron oxide seems to be the dominant pigment in the Lower Permian red beds.

Many grain-to-grain contacts apparently lack hematite pigment (Pl. II, fig. 5) and even when the hematite coatings occur on quartz or feldspar grains, they show no rounding by abrasion. It thus seems clear that the formation of these coatings took place post-depositionally. Miller and Folk (1955, p. 344), Walker (1967a, p. 366), Glennie (1970, p. 185), and Turner (1974, p. 221) noted similar hematite-free grain-to-grain contacts in red beds and also suggested that these indicated post-depositional origin of the hematite grain coatings. A similar conclusion is presented by Przybyłowicz (1968, p. 811) who studied the internal character of the opaque grains from the Permian sandstones in an adjacent area; she found that the opaque nature of these grains resulted

from the post-depositional hematite coatings and infiltrations. It is also noteworthy that, in contrast to the red sediment, no real hematite coatings have been observed on detrital grains in the white sandstones, and they are also less common in the intermediate (pink, salmon) sediments.

Iron analyses

Nineteen samples of white, intermediate and red sandstones were analysed for ferrous and ferric iron. The samples were divided into three groups (see Table 3) corresponding to the three different depth--levels of the Permian sandstones: white sandstones of association III (uppermost in the profiles), white sandstones of association II, and red (including intermediate) sandstones of associations II and I (see also Fig. 1). These groups are not differentiated with respect to the Fe^{2+}

Results of iron analyses — Table 3

Zawartości różnych form żelaza — Tabela 3

Sample No.	Colour	Sandstone association		Fe^{2+} /%/	Fe^{3+} /%/	Fe^{2+}/Fe^{3+}
2	white	III		1.33	0.09	14.77
19	white	III		1.26	0.12	10.50
51	white	III		1.17	0.10	11.70
48	white	III		1.17	0.08	14.62
			Mean	1.23	0.09	12.88
			Standard Dev.	0.0025	0.0075	0.0175
3	white	II		2.30	0.10	23.00
49	white	II		1.16	0.11	10.54
16	white	II		1.31	0.14	9.35
52	white	II		1.12	0.16	7.00
			Mean	1.47	0.12	12.47
			Standard Dev.	0.0025	0.0063	0.0025
4	v.light pink	II		1.21	0.42	2.88
5	pink	II		1.18	0.63	1.87
6	light red	II		1.18	0.92	1.28
50	red	II		1.22	1.73	0.70
20	white, red mottled	II		1.27	0.63	2.01
21	pink, red mottled	II		1.20	1.32	0.91
22	red	II		1.22	1.44	0.84
23	red	II		1.13	1.73	0.65
53	red	II		1.16	1.21	0.95
7	red	I		1.37	1.44	0.95
59	red	I		1.47	1.65	0.89
			Mean	1.23	1.19	1.26
			Standard Dev.	0.0072	0.0062	0.0154
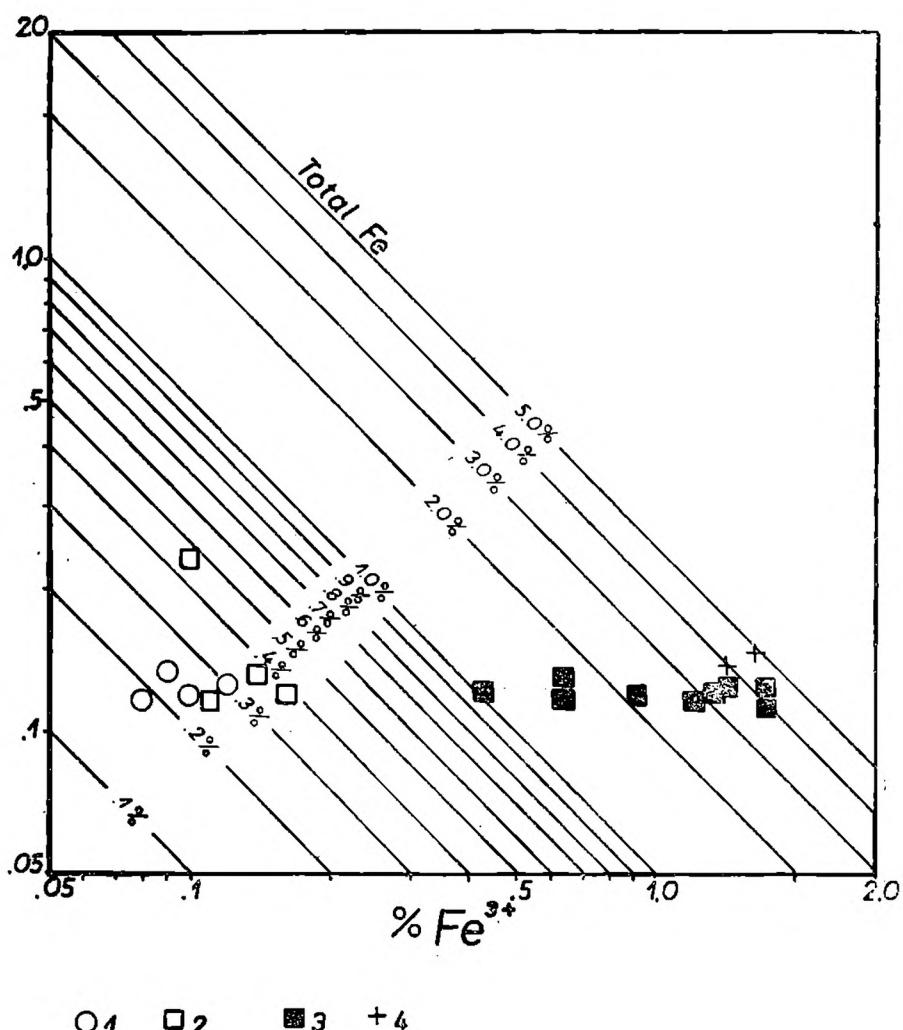
Fig. 6. Plot of Fe^{3+} vs. Fe^{2+} in Permian sandstone samples of different colour; see Table 3 for data. 1 — white shallow-marine ss.; 2 — white aeolian ss.; 3 — red aeolian ss.; 4 — red fluvial ss.

Fig. 6. Wykres zależności Fe^{2+} od Fe^{3+} w permskich piaskowcach różnej barwy; zob. dane w tabeli 3. 1 — białe piaskowce płytkomorskie; 2 — białe piaskowce eoliczne; 3 — czerwone piaskowe eoliczne; 4 — czerwone piaskowce rzeczne

content, but there are significant differences in the ferric iron (Fe^{3+}) content and in the Fe^{2+}/Fe^{3+} ratio. The results are presented in Table 3 and expressed graphically in Figs. 6—8. The red sandstones have averagely of 1·07% more Fe^{3+} than the white sandstones of association II, and 1·10% more Fe^{3+} than the overlying sandstones of association III. Thus, the Fe^{3+} content decreases upward the profiles. Also the Fe^{2+}/Fe^{3+} ratio in the red sandstones clearly differs from that in the remaining sandstone groups.

It is possible that the relative enrichment of Fe^{3+} occurred in two ways:

(1) selective reduction (and possibly dissolution) of Fe^{3+} in entirely red sediment which ultimately become white (Miller and Folk, 1955; Picard, 1965);
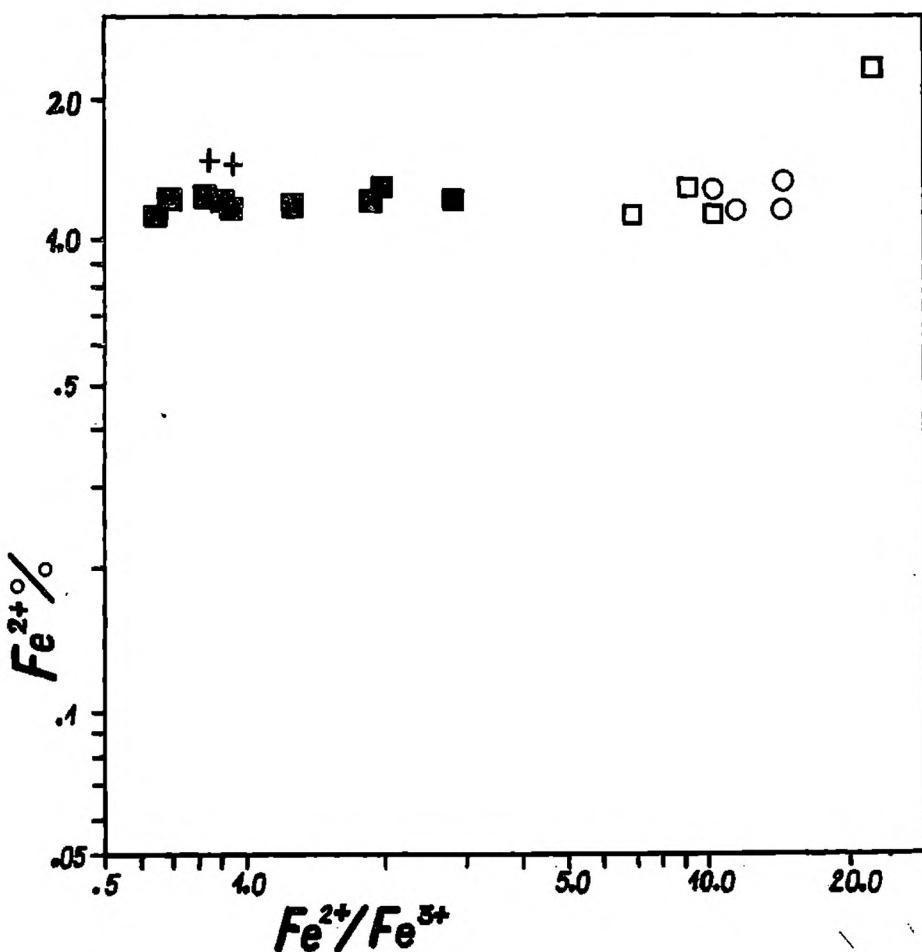
Fig. 7. Plot of Fe^{2+}/Fe^{3+} vs. Fe^{2+} in Permian sandstone samples; see Fig. 6 for explanations

Fig. 7. Wykres zależności Fe^{2+} od Fe^{2+}/Fe^{3+} w piaskowcach permskich; objaśnienia na fig. 6

(2) process of *in situ* oxidation of iron-bearing, silicates, in originally non-red sediment, which were deposited, and remained (except for the uppermost part of the sequence), in an oxidizing environment for a relatively long period of time (Walker, 1967a; Walker and Honea, 1969; see also previous section).

The former process is unlikely to account for the results obtained. Thompson (1970) and McBride (1974) noted an apparent retention of iron in sandstones whose red colour was lost by reduction. Thus, we would expect the same total iron content in both red and white sandstones or a relatively higher Fe^{2+} content in the white sediment. It is clear, however, from the present results that this is not so. As mentioned earlier, there are also no signs of large — scale iron dissolution and redistribution as well as no field evidence of iron concentration out of the sequence. Moreover, Kolthoff et al. (1969, p. 252) reported that iron is insoluble even in the reduced state at pH value above 6.5—7.0. The carbonate cement content in the studied white sandstones suggests in-

4*
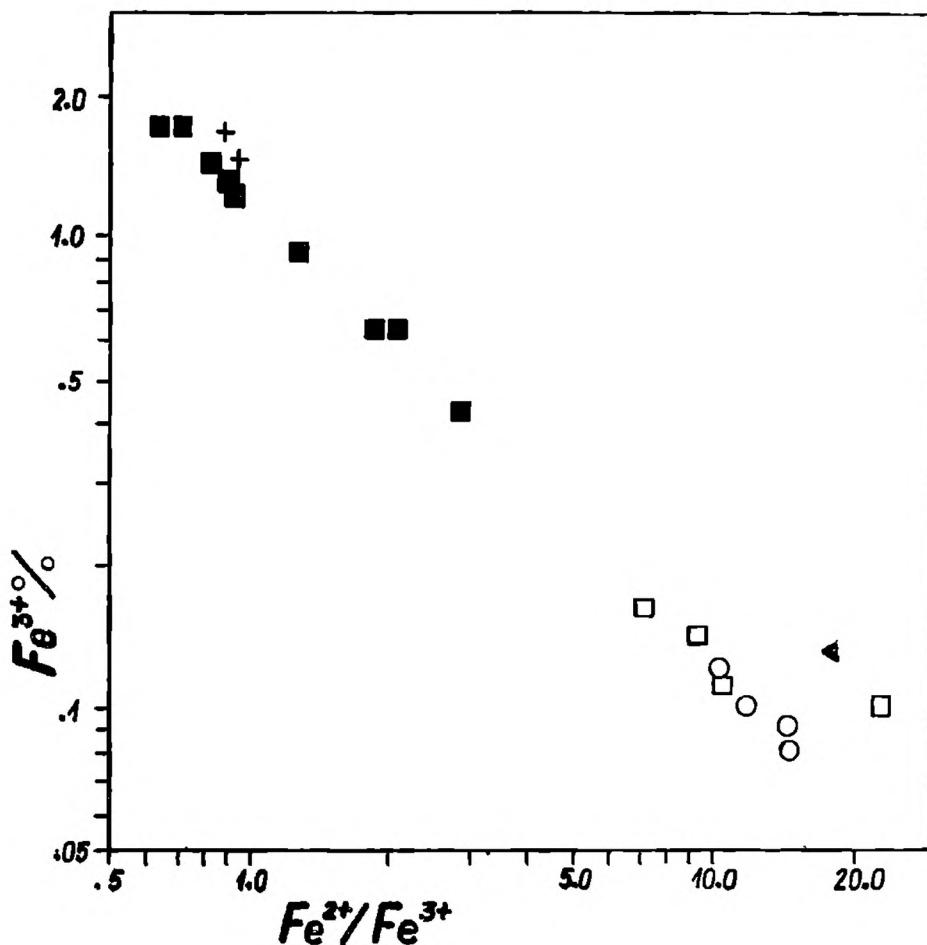
Fig. 8. Plot of Fe^{2+}/Fe^{3+} vs. Fe^{3+} in Permian sandstone samples; see Fig. 6 for explanations

Fig. 8. Wykres zależności Fe^{3+} od Fe^{2+}/Fe^{3+} w piaskowcach permskich; objaśnienia na fig. 6

terstitial pH values above 8 during diagenesis. Additionally, there are no signs of sphaerosiderite nodules and/or paleosiderite layers formation in basal parts of the white sandstones, which might have indicated bleaching of red sediment by collecting iron from the adjacent parts (see e. g. Slànskà, 1976), and there are also no evidence of organic material having ever been present in the studied sandstones.

Thus, an origin by reduction, and/or redistribution, of iron does not explain the present differences in iron content in the Permian sedimentary sequence. The evidence from the iron analyses rather indicates that there is a close relationship between the Fe^{3+} enrichment and the formation of red colouration. Thus, the increase in Fe^{3+} content is perhaps due to the sediment reddening process being discussed in the previous section. This also suggests that the alteration processes, which formed the Lower Permian red beds, were not active within the now non-red portion of the sequence (see also previous section).

DISCUSSION

A synthesis of all the pertinent evidence clearly suggests that the origin of red pigment in the studied Lower Permian red beds is due to *in situ* alteration of iron-bearing detrital grains. The only problem which might need consideration is the white colour of the uppermost portion of the Permian sandstones i.e., Weissliegendes sandstones.

Contrary to the previous opinions, that these white sandstones are the result of bleaching of originally all-red deposit, the present authors represent another point of view based on the evidence discussed above. Theoretically, the originally red sediment might have been bleached either by mechanical grain abrasion or by chemical reduction of ferric iron, or by a combination of these processes. The former possibility should be rejected in the case of such the fine-grained sediment because, as shown by Kuenen (1960, p. 438) and by Norris and Norris (1961, p. 612), the mechanical abrasion of the grain red-coatings is inconspicuous even in water-free (aeolian) environment. Also the second possibility seems to be unlikely in the light of the experimental work of Berner (1969 b, 1971). From both theoretical and experimental points of view, there is no reason to assume that such the large-scale bleaching phenomena have been associated with the Zechstein marine transgression. The influence of the infiltrated-downwards marine water and/or of the anaerobic environment of the Kupferschiefer sediment do not explain the great, and largely variable, thickness of the white sediment (see also B e r n e r, 1969b). The thickness of the Weissliegendes sandstones varies from profile to profile, ranging from less than 1 m up to 40 m, and sporadically to 60 m. As shown above, there are also no signs of ferric iron reduction and, generally, no evidence of ferric iron having ever been present, in this white sediment, in a significant amount.

In summary, there is no evidence that the Weissliegendes sandstones have ever been red (see also Magraw, 1975). Their white colour is thought to be closely related to the post-depositional origin of the underlying Rotliegendes red beds. In this line of evidence the white sandstones represent the non-reddened, uppermost portion of the Permian clastic sequence, while the intermediate sandstones correspond to the transitional zone of only partly reddened sediment. As a whole, the model is simple and well-explained by the processes of reddening of dune sediments with age, the phenomenon being recently reported from many modern and ancient desert environments (e. g. Norris and Norris, 1961; Price, 1962; Norris, 1969; Bigarella et al., 1971 see also numerous examples cited by Folk, 1976, p. 605).

Bleaching of red sediment due to the Zechstein marine transgression cannot be, however, completely excluded, but it might have occurred

only locally, i. e., where the sedimentary sequence was reddened to a relatively high level being close to the ground surface. Thus, bleaching is thought 'to be possible, but rather as a local, small-scale phenomenon.

REFERENCES — WYKAZ LITERATURY

Alexandrowicz S. W., Słupczyński K. (1970), O występowaniu Lingula credneri Geinitz w osadach dolnego cechsztynu okolic Ostrowa Wielkopolskiego. On the presence of Lingula credneri Geinitz in the lower Zechstein deposits in the vicinity of Ostrów Wielkopolski. Acta geol. polon., 20, pp. 694—710.

Alimen H. (1957), The prehistory of Africa. Hutchison, 438 p., London.

Berner R. A. (1969 a), Goethite stability and the origin of red beds. Geoch. Cosmoch. Acta, 33, pp. 267—273.

Berner R. A. (1969 b), Migration of iron and sulfar within anaerobic sediments during early diagenesis. Am. J. Sci., 268, pp. 19—42.

Berner R. A. (1971), Principles of chemical sedimentology. McGraw — Hill, 240 p., New York.

Bigarella J. J., Duarte G. M., Becker R. D. (1971), Structural characteristics of the dune, foredune, interdune, beach, beach-dune ridge and sand ridge deposits. Geociens. Bol. Par., 28/29, pp. 9—72.

Blatt H., Middleton G., Murray R. (1972), Origin of sedimentary rocks. Prentice — Hall, 634, p., New Jersey.

Brongersma-Sanders M. (1969), Permian wind and the occurrence of fish and metals in the Kupferschiefer and Marl Slate. [In:] Sedimentary ores: ancient and modern (revised). Proc. of the 15th Inter-University Geol. Congress, 1967. University of Leicester, England, pp. 61—71.

Dunham K. C. (1953), Red coloration in desert formations of Permian and Triassic age in Britain. Congr. Geol. Inter. 1952, sect. VII, Fasc. VII, pp. 25—32.

Dziedzic K. (1961), Utwory dolnopermskie w niecce śródsudeckiej. Lower Permian of the Intra Sudetic Basin. Studia Geol. Polon., VI, 121 p.

Eugster H. P., Chou I. M. (1973), The depositional environments of Precambrian banded iron-formation. Econ. Geol., 68, pp. 1144—1168.

Feys R. (1964), The paleopedology of coal basin. [In:] A. E. M. Nairn (Ed.) Problems in paleoclimatology. Interscience Publ., 705 p., New York.

Folk R. L. (1976), Reddening of desert sands: Simpson Desert, N. T., Australia. J. Sediment. Petrol., 46, pp. 604—615.

Friend P. F. (1966), Clay fractions and colours of some Devonian red beds in the Catskill Mountains, USA. Quart. J. Geol. Soc. Lond., 122, pp. 273—292.

Friend P. F., Harland W. B., Hudson J. D. (1963), The Old Red Sandstone and the highland boundary in Arran, Scotland. Trans. Edinb. Geol. Soc., 19, pp. 363—425.

Glennie K. W. (1970), Desert sedimentary environments. Elsevier, 222 p., Amsterdam.

G l e n n i e K. W. (1972), Permian Rotliegendes of Northwest Europe interpreted in light of modern desert sedimentation studies. *Am. Assoc. Petrol. Geol. Bull.,* 56, pp. 1048—1071.

G r a y R. S. (1967), Petrography of the Upper Cenozoic non-marine sediments in the San Pedro Valley, Arizona. *J. Sediment. Petrol.,* 37, pp. 774—789.

G r i m R. E. (1958), Concept of diagenesis in argillaceous sediments. *Am. Assoc. Petrol. Geol. Bull.,* 42, pp. 246—253.

H u b e r t J. F. (1960), Petrology of the Fountain and Lyons Formations, Front Range, Colorado. *Colorado School Mines Quart.,* 55, 242 p.

J a c o b s M. B. (1970), Clay mineral investigations of Cretaceous and Quaternary deep-sea sediments of the North American Basin. *J. Sediment. Petrol.,* 40, pp. 864—868.

J a c o b s M. B. (1974), Clay mineral changes in Antarctic deep-sea sediments and Cenozoic climatic events. *J. Sediment. Petrol.,* 44, pp. 1079—1086.

J a r z y k i e w i c z T., K i j e w s k i P., M r o c z k o w s k i J., T e i s s e y r e A. K. (1976), Geneza osadów białego spągowca monokliny przedsudeckiej. Origin of the Weissliegendes deposits in the Fore-Sudetic Monocline. *Geol. Sudetica,* XI, pp. 57—97.

K u e n e n Ph. H. (1960), Experimental abrasion. 4. Eolian action. *J. Geol.,* 68, pp. 427—449.

K o l t h o f f I. M., S a n d e l l E. B., M e e k e r E. J., B r u c k e n s t e i n S. (1969), Quantitative chemical analysis. McMillan, 320 p., London.

K r y n i n e P. D. (1949). The origin of red beds. *New York Acad. Sci. Trans.,* Ser. 2, 11, pp. 60—68.

K r y n i n e P. D. (1950), Petrology, stratigraphy, and the origin of the Triassic sedimentary rocks of Connecticut. *Connect. Geol. and Nat. History Survey Bull.,* 73, 239 p.

L e o n e M., A l a i m o R., C a l d e r o n e S. (1975), Genesis of chlorite pellets from Mesozoic bedded cherts of Sicily. *J. Sediment. Petrol.,* 45, pp. 618—628.

M a g r a w D. (1975), Permian of the offshore and coastal region of Durham and SE Northhumberland. *J. Geol. Soc.,* 131, pp. 397—414.

M a r e l H. W. van der (1956), Quantitative defferential thermal analyses of clay and other minerals. *Amer. Miner.,* 41, pp. 222—244.

M c B r i d e E. F. (1974), Significance of color in red, green, purple, olive, brown, and gray beds of Difunta Group, Northeastern Mexico. *J. Sediment. Petrol.,* 44, pp. 760—773.

M i l l e r D. N., F o l k R. L. (1955), Occurrence of detrital magnetite and ilmenite in red sediments: new approach to the significance of red beds. *Amer. Assoc. Petrol. Geol. Bull.,* 39, pp. 338—345.

M i l n e I. H., E a r l e y J. W. (1958), Effect of source and environment on clay minerals. *Am. Assoc. Petrol. Geol. Bull.,* 42, pp. 328—338.

N e m e c H., N e m e c W., P o r ę b s k i S. J. (1978), Weissliegendes Sandstones: a transition from fluvial-aeolian to shallow-marine sedimentation (Permian of Fore-Sudetic Monocline). 3. Interpretation in light of heavy-mineral data. *Rocz. Pol. Tow. Geol.,* in press.

N e m e c W., P o r ę b s k i S. J. (1977), Weissliegends Sandstones: a transition from fluvial-aeolian to shallow-marine sedimentation (Permian of Fore-Sudetic Monocline). 1. Sedimentary structures and textural differentiation. *Rocz. Pol. Tow. Geol.,* vol. 47.

N o r r i s R. M. (1969), Dune reddening and time. *J. Sediment. Petrol.,* 39, pp. 7—11.

N o r r i s R. M., N o r r i s K. S. (1961), Algodones dunes of southeastern California. *Geol. Soc. Am. Bull.,* 72, pp. 605—620.

O k a d a H. (1971), Classification of sandstone: analysis and proposal. *J. Geol.*, 79, pp. 509—525.

P i c a r d M. D. (1965), Iron oxides and fine-grained rocks of Red Peak and Crow Mountain sandstone members, Chugwater (Triassic) Formation, Wyoming. *J. Sediment. Petrol.*, 35, pp. 464—479.

P r i c e W. A. (1962), Stages of oxidation coloration in dune and barrier sands with age. *Geol. Soc. Amer. Bull.*, 73, pp. 1281—1284.

P r z y b y ł o w i c z T. (1968), O możliwości rozdzielenia czerwonego i białego spągowca w rejonie Ostrowa Wielkopolskiego na tle badań petrograficznych. The differentiation of the Rotliegende from the „Weissliegende" by petrographic studies. *Acta geol. polon.*, XVIII, pp. 801—815.

R a y m o n d P. E. (1927), The significance of red color in sediments. *Amer. J. Sci.*, 13, pp. 234—254.

R o b b G. L. (1949), Red bed coloration. *J. Sediment. Petrol.*, 19, pp. 99—103.

S c h l u g e r P. R. (1976), Petrology and origin of the red beds of the Perry Formation, New Brunswick, Canada, and Maine, USA. *J. Sediment. Petrol.*, 46, pp. 22—37.

S c h l u g e r P. R., R o b e r s o n H. E. (1975), Mineralogy and chemistry of the Patapsco Formation, Maryland, related to the ground-water geochemistry and flow system: a contribution to the origin of red beds. *Geol. Soc. Amer. Bull.*, 86, pp. 153—158.

S i e v e r R., K a s t n e r M. (1972), Shale patrology by electron microprobe: pyrite--chlorite relations. *J. Sediment. Petrol.*, 42, pp. 350—355.

S l à n s k à J. (1976), A red-bed formation in the South Bohemian Basins, Czechoslovakia. *Sediment. Geol.*, 15, pp. 135—164.

S t o c h L. (1974), Minerały ilaste (*in Polish only*). Wydaw. Geol., 492 p., Warszawa.

T h o m p s o n A. M. (1970), Geochemistry of color genesis in red bed sequence, Juanita and Bald Eagle Formations, Pennsylvania. *J. Sediment. Petrol.*, 40, pp. 599—615.

T u r n e r P. (1974), Origin of red beds in the Ringerike Group (Silurian) of Norway. *Sediment. Geol.*, 12, pp. 215—235.

V a n H o u t e n F. B. (1961), Climatic significance of red beds. [In:] A. E. M. Nairn (En.) Descriptive palaeoclimatology. Interscience Publish.. pp. 89—139, New York.

V a n H o u t e n F. B. (1964), Origin of red beds — some unsolved problems. [In:] A. E. M. Nairn (Ed.) Problems in palaeoclimatology. Interscience Publish., pp. 647—661, New York.

V a n H o u t e n F. B. (1968), Iron oxides in red beds. *Geol. Soc. Amer. Bull.*, 79, pp. 399—416.

V a n H o u t e n F. B. (1972), Iron and clay in tropical savanna alluvium, northern Colombia: a contribution to the origin of red beds. *Geol. Soc. Amer. Bull.*, 83, pp. 2761—2772.

W a h l s t r o m E. E. (1948), Pre-Fountain and recent weathering on Flagstaff Mountain near Boulder, Colorado. *Geol. Soc. Amer. Bull.*, 59, pp. 1173—1190.

W a l k e r T. R. (1967 a), Formation of red beds in modern and ancient deserts. *Geol. Soc. Amer. Bull.*, 78, pp. 353—368.

W a l k e r T. R. (1967b), Color of recent sediments in tropical Mexico: a contribution to the origin of red beds. *Geol. Soc. Amer. Bull.*, 78, pp. 917—920.

W a l k e r T. R. (1973), Intrastratal alterations in the Fountain Formation (Pennsylvanian age), Denver area, Colorado. *Geol. Soc. Amer., Abs. with Programs (Southeastern Sect.)*, 5, pp. 521—522.

W a l k e r T. R. (1974), Formation of red beds in moist tropical climates: a hypothesis. *Geol. Soc. Amer. Bull.*, 85, pp. 633—638.

W a l k e r T. R., H o n e a R. M. (1969), Iron content of modern deposits in the Sonoran Desert: a contribution to the origin of red beds. *Geol. Soc. Amer. Bull.*, 80, pp. 535—544.

W a l k e r T. R., R i b b e P. H. (1967), Formation of iron-rich authigenic clay by intrastratal alteration of hornblende in an arid climate: a contribution to the origin of hematite-stained matrix in red beds. *Geol. Soc. Amer., Spec. Pap.* 101, *Abstr. 1966*, p. 233.

W a l k e r T. R., R i b b e P. H., H o n e a R. M. (1967), Geochemistry of hornblende alteration in Pliocene red beds, Baja California, Mexico. *Geol. Soc. Amer. Bull.*, 78, pp. 1055—1060.

W e a v e r Ch. E. (1958a), Geologic interpretation of argillaceous sediments. Part I. Origin and significance of clay minerals in sedimentary rocks. *Amer. Assoc. Petrol Geol., Bull.*, 42, pp. 254—271.

W e a v e r Ch. E. (1958b), Geologic interpretation of argillaceous sediments. Part II. Clay petrology of Upper Mississippian — Lower Pennsylvanian sediments in central United States. *Amer. Assoc. Petrol. Geol. Bull.*, 42, pp. 272—309.

STRESZCZENIE

Niniejsza praca stanowi drugą część studium poświęconego zagadnieniom warunków sedymentacji permskich piaskowców białego spągowca monokliny przedsudeckiej. Piaskowce te są ogniwem przejściowym pomiędzy dolnopermskimi kontynentalnymi utworami barwy czerwonej i morskimi osadami górnego permu (cechsztynu). W profilu stratygraficznym są one wyróżniane głównie, jeżeli nie wyłącznie, na podstawie swej białej barwy. Zdaniem autorów piaskowce białego spągowca są osadem o złożonej genezie (Nemec, Porębski, 1977). Ich dolna część, reprezentowana przez piaskowce z wielkoskalowym warstwowaniem przekątnym, leży na analogicznie wykształconych piaskowcach barwy czerwonej. Granica pomiędzy tymi piaskowcami na ogół przejawia się w stopniowych zmianach w barwie osadu: od brunatnoczerwonej (w niższych partiach profili) poprzez odcienie barwy czerwonej i różowej, do jasnożółtej i białej w górnych partiach profili. Miąższości tej strefy przejściowej wahają się od kilku do kilkunastu metrów (por. Nemec, Porębski, 1977, fig. 3). Wspomniany zespół piaskowców interpretowany jest jako osad środowiska eolicznego. Górna część piaskowców białego spągowca, odznaczająca się obecnością fauny morskiej, bioturbacji i strukturami sedymentacyjnymi o wodnej genezie, stanowi niewątpliwie płytkomorski, transgresywny osad cechsztynu.

Białą barwę piaskowców białego spągowca interpretowano tradycyjnie jako efekt odbarwienia pierwotnie w całości czerwonego osadu, co miało nastąpić w wyniku jego redepozycji podczas transgresji morskiej, bądź też wskutek postdepozycyjnej infiltracji wody morskiej o charakterze redukcyjnym. Stąd też występowanie białych piaskowców w stro-

pie utworów czerwonego spągowca wiązano zazwyczaj ściśle z bezpośrednim wpływem transgresji cechsztyńskiej, a samą ich barwę uznawano w tym względzie niejako za kryterium środowiskowe.

Niniejsza praca jest próbą weryfikacji tych poglądów w świetle uprzednio przedstawionej interpretacji środowiskowej, a punktem wyjścia było ustalenie genezy barwy czerwonej utworów podścielających biały spągowiec. Innym zagadnieniem, potraktowanym w pracy skrótowo, jest zróżnicowanie petrograficzne badanych piaskowców.

PETROGRAFIA

Piaskowce białego i czerwonego spągowca mają stosunkowo jednorodny skład mineralogiczno-petrograficzny. W klasyfikacji modalnej O k a d y odpowiadają one arenitom kwarcowym, podrzędnie arenitom litycznym (fig. 3). Szkielet ziarnowy składa się głównie z kwarcu (40— 50%, maksymalnie do 70%). Udział okruchów skalnych waha się w dość znacznym zakresie, a wśród nich występują: kwarcyty i łupki kwarcytyczne (0.5—12%), fylity (0—1%), iłowce (0—3%), rzadziej piaskowce kwarcytyczne i mułowce (0—2%), granity i gnejsy (0—1%), ryolity (0— 1%) oraz łupki kwarcowo-serycytowe (0—0.5%). Udział skaleni (albit, oligoklaz, mikroklin) zmienia się w granicach 0.1—15% i jest z reguły wyższy w grubiej ziarnistych piaskowcach. W niewielkich ilościach występują detrytyczne łyszczki, głównie schlorytyzowany biotyt oraz akcesoryczne minerały ciężkie. Zawartość masy wypełniającej w większości prób waha się od 2 do 4% i chociaż jest wyższa w piaskowcach asocjacji rzecznej, to jednak nigdy nie przekracza 15%. Udział spoiwa chemicznego zmienia się od 15 do 40%. Dominuje cement ilasty (0—20%), węglanowy (0.5—20%) i krzemionkowy (0—13%), a udział spoiwa anhydrytowego jest niski (0—2%). Porównanie składu mineralnego w kategoriach średnich zawartości kwarcu, okruchów skalnych, spoiwa węglanowego, ilastego i anhydrytowego ustalonych dla trzech grup piaskowców (czerwonych i białych piaskowców asocjacji eolicznej oraz piaskowców płytkomorskich) wykazało zasadnicze mineralogiczne podobieństwo całości tych piaskowców (tab. 1). Uzyskane dane wskazują również na wzrastającą w górę profili dojrzałość mineralogiczną badanej sekwencji piaskowcowej.

Przy zastosowaniu metody DTA zbadano skład mineralny frakcji ilastej w 15 próbach piaskowców asocjacji eolicznej (II) i płytkomorskiej (III). Zidentyfikowano illit, kaolinit i chloryt, a w dwóch próbach niepewnie oznaczono także montmorylonit (fig. 5, tab. 2). Nie stwierdzono zależności pomiędzy barwą piaskowców a występującym w nich zespołem minerałów ilastych. Pokrywa się to z obserwacjami innych autorów dotyczącymi współwystępujących ze sobą czerwonych i białych (lub szarych) osadów (np. Friend, 1966; Turner, 1974). W większo-

ści prób piaskowców asocjacji płytkomorskiej stwierdzono jednak chloryt, który z kolei sporadycznie występuje w próbach białych i czerwonych piaskowców asocjacji eolicznej. Biorąc jednak pod uwagę możliwość zmian diagenetycznych w osadzie, a także ograniczenia związane z samą metodą DTA, trudno jest jednoznacznie stwierdzić, czy obecność (lub brak) chlorytu we frakcji ilastej może być w tym przypadku traktowana jako wskaźnik środowiskowy. Aczkolwiek istnieją ku temu przesłanki (por. Grim, 1958; Milne, Earley, 1958; Weaver, 1958 a, b; Jacobs, 1970; Siever, Kastner, 1972), jednak dopiero szczegółowe badania minerałów ilastych piaskowców permskich mogą przynieść potwierdzenie lub odrzucenie tej sugestii.

GENEZA BARWY

Zgodnie z dotychczasową interpretacją biała barwa piaskowców białego spągowca mogła być wynikiem odbarwienia najwyższej partii osadów czerwonego spągowca na drodze: (1) mechanicznego usunięcia substancji żelazistej z osadu w warunkach abrazji morskiej lub (2) redukcji żelaza trójwartościowego na skutek postdepozycyjnej infiltracji roztworów euksynicznych. Możliwość pierwsza jest jednak nie do przyjęcia w świetle eksperymentalnych prac Kuenena (1960). W przypadku bowiem osadów tak drobnoziarnistych jak badane, mechaniczne usuwanie powłoczek żelazistych z ziarn zachodzi w stopniu znikomym nawet w warunkach abrazji eolicznej, gdzie poszczególne ziarna pozbawione są ochronnych powłoczek wodnych (por. też Norris, Norris, 1969, s. 612). Podobnie odbarwienie na skutek infiltracji roztworów redukcyjnych wydaje się niemożliwe na tak dużą skalę (por. wyniki eksperymentalnych prac Bernera, 1969 a, 1971). Przeciwko tej ostatniej ewentualności przemawia dodatkowo szereg innych faktów, jak: (a) wysoce zróżnicowana miąższość białego spągowca — od kilkudziesięciu centymetrów do kilkudziesięciu metrów, (b) wyraźnie wyższa zawartość Fe^{3+} w piaskowcach barwy czerwonej przy braku różnic w zawartości Fe^{2+} pomiędzy osadem czerwonym i białym (fig. 6—8, tab. 3), i przy jednoczesnym (c) braku jakichkolwiek śladów odprowadzenia z osadu tak znacznej ilości żelaza. Pozostaje zatem pytanie; jaka jest geneza barwy piaskowców białego spągowca i jaki jest jej związek ze środowiskiem sedymentacji?

Kluczem do wytłumaczenia obserwowanych zmian w barwie piaskowców permskich jest niewątpliwie wyjaśnienie genezy samej barwy czerwonej utworów podścielających biały spągowiec. Czerwone zabarwienie osadów kopalnych jest zwykle wywołane obecnością w nich pigmentu hematytowego. Charakter procesów prowadzących do koncentracji hematytu w osadzie jest jednak do tej pory przedmiotem dyskusji (por. Krynine, 1947; Van Houten, 1964, 1968; Walker, 1967 a). Uzyskane dane wskazują, że barwa czerwona piaskowców permskich jest wyni-

kiem występowania w nich hematytu, a w niewielkim stopniu getytu, o różnej lokalizacji w obrębie szkieletu ziarnowego i masy wypełniającej. Występujące w osadzie ziarna krzemianów bogatych w żelazo (biotyt, chloryt, hornblenda) wykazują różne stadia postdepozycyjnego zastępowania ich przez autigeniczny hematyt. Poszczególne stadia tego procesu to: (1) czerwienienie ziarn wzdłuż powierzchni łupliwości i partii peryferycznych, (2) rozwój nieprzeźroczystego hematytu wzdłuż powierzchni łupliwości, (3) powstawanie gęsto upakowanego, nieregularnie wykształconego hematytu pokrywającego w formie skupień powierzchnie ziarn i rozrastającego się w obrębie masy wypełniającej (tabl. I i II). Ekstensywny charakter wspomnianych przemian wskazuje, że procesy powstawania autigenicznego hematytu zachodziły raczej wewnątrz złożonego osadu aniżeli na jego powierzchni (por. Walker, 1967 a). Wiele kontaktów międzyziarnowych pozbawionych jest pigmentu hematytowego, a w przypadku gdy ziarna kwarcu lub skaleni mają obwódki hematytowe, te ostatnie na ogół nie wykazują śladów abrazji. Powstanie tych obwódek miało niewątpliwie miejsce po złożeniu osadu (por. Miller, Folk, 1955; Walker, 1967 a). Występowanie skupień hematytu wokół fragmentów skał wulkanicznych i ziarn magnetytu świadczy, że niemal wszystkie detrytyczne ziarna bogate w żelazo były w różnym stopniu źródłem dla pigmentu hematytowego. W profilu pionowym badanych piaskowców obserwuje się ku górze wzrost udziału ziarn krzemianów zawierających żelazo. Wynika to stąd, że ku górze są one w coraz mniejszym stopniu zastąpione przez autigeniczny hematyt, a więc lepiej zachowane w osadzie. W piaskowcach białych nie stwierdzono prawdziwych obwódek hematytowych wokół ziarn, a udział takich minerałów jak: hornblenda, epidot i granat jest tu wyższy niż w piaskowcach czerwonych. Wskazuje to wyraźnie na odwrotną zależność pomiędzy ilością ziarn krzemianów bogatych w żelazo w osadzie a stopniem jego zaczerwienienia, tj. zabarwienia przez hematyt. Można zatem sądzić, że niższy udział ziarn krzemianów bogatych w żelazo w piaskowcach czerwonych jest wynikiem daleko posuniętego ich zastąpienia przez hematyt. Z przedstawionych danych można wnioskować, że piaskowce białego spągowca reprezentują najwyższą, nie zabarwioną partię sekwencji eolicznej. Postępujące w czasie zabarwianie piasków wydmowych związane było ściśle z wahaniami wód gruntowych (por. Walker, 1967 a) i stopniowo zatrzymywane w miarę narastania transgresji cehsztyńskiej. Diachroniczny przebieg procesów zabarwiania, a przede wszsytkim ich związek z morfologią pól wydmowych są powodem aktualnie obserwowanych znacznych różnic w miąższości piaskowców białego spągowca.

W podsumowaniu należy stwierdzić, że w badanym obszarze nie znaleziono dowodów na to, że piaskowce białego spągowca były w swej przeważającej masie kiedykolwiek czerwone (por. też Magraw, 1975).

Fig. 1. Ziarna hematytu o cechach typowych dla współwystępujących łyszczyków (płatkowy pokrój, orientacja i deformacje kompakcyjne). Płytka cienka, jeden nikol. Działki skali odpowiadają 0,02 mm

Fig. 2. Hornblende grain (h) being etched, densely hematite stained and peripherally altered to authigenic clay, and hematites (a) sticking in matrix and containing small matrix „inclusions . Thin section, one nicol. Scale bars correspond to 0.01 mm

Fig. 2. Silnie zmienione (częściowo w autigeniczny ił) i utkane hematytem ziarno hornblendy (h) oraz tkwiące w matrix ziarna hematytu (a) zawierające drobne „inkluzje" materiału stanowiącego matrix. Płytka cienka, jeden nikol. Działki skali odpowiadają 0,01 mm

Fig. 3. Hornblende? (h) almost completely altered to authigenic clay; the accompanying released iron is still preserved in a form of opaque hematite surrounding the parent grain. Thin section, one nicol. Scale bars correspond to 0.01 mm

Fig. 3. Hornblenda? (h) niemal zupełnie zmieniona w autigeniczny ił, gęsto utkana i otoczona hematytem. Płytka cienka, jeden nikol. Działki skali odpowiadają 0,01 mm

Fig. 4. Peripherally altered hornblende surrounded by hematite stained halo in fine-grained pink sandstone. Thin section, one nicol. Scale bars correspond to 0.005 mm

Fig. 4. Hornblenda peryferycznie zmieniona silnie w autigeniczny ił; uwolnione żelazo tworzy obwódkę hematytową wokół ziarna. Płytka cienka, jeden nikol. Działki skali odpowiadają 0,005 mm

Fig. 5—6. Detrital quartz grains (q) partly replaced by hematite crystals (a); in fig. 5 hematite developed due to alteration of pyrite or magnetite, perhaps, as indicated by crystal shapes. Thin section, one nicol. Scale bars correspond to 0.02 mm

Fig. 5—6. Autigeniczny hematyt (a) zastępujący peryferycznie ziarna kwarcu (q); na fig. 5 hematyt mógł powstać przez zastąpienie pirytu lub magnetytu, na co wskazuje pokrój kryształu. Płytka cienka, jeden nikol. Działki skali odpowiadają 0,02 mm