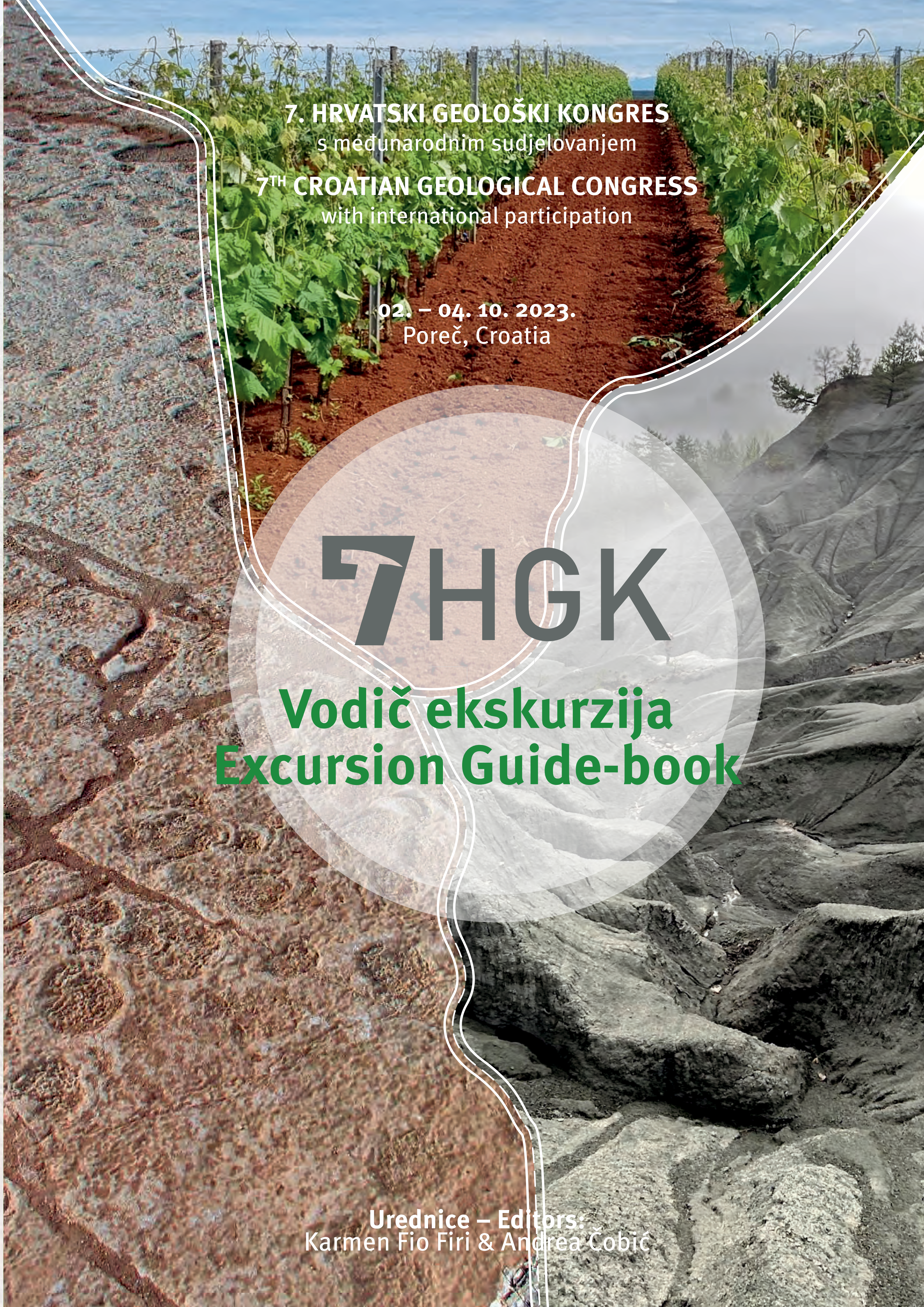

7. HRVATSKI GEOLOŠKI KONGRES
s međunarodnim sudjelovanjem

7TH CROATIAN GEOLOGICAL CONGRESS
with international participation

02. – 04. 10. 2023.
Poreč, Croatia

7HGK

Vodič ekskurzija
Excursion Guide-book

Urednice – Editors:
Karmen Fio Firi & Andrea Čobić

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CONTENT

Responsibility for the content within the field-guide lies entirely within the authors.
Authors are also responsible for the provided addresses and affiliation information.

A BRIEF INTRODUCTION TO THE GEOLOGY OF ISTRIA

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INTRODUCTION

Istria, the largest Croatian peninsula (its area in the Republic of Croatia is 3130 km^2, which is about 90 % of the total area of Istria) is known, among others, for its very interesting geology. For example, outcrops of rocks of a very wide stratigraphic range (almost 170 Ma) are very well exposed, including some very specific stratigraphic units, the deposits are mostly only slightly tectonised, almost all of the traces of dinosaurs in Croatia are discovered in the area of Istria, there are numerous quarries of very different types of building stone, the best visible boundary of the External Dinarides is located in Istria, etc.

Therefore, it is not unusual that several important international and national geological meetings were already held in Istria, e.g. *the 16th European Micropaleontological Colloquium* (BAUER & POLŠAK, 1979; DROBNE, 1979; DROBNE et al., 1979; HAGN et al., 1979; SOKAČ & VELIĆ, 1979), *the 4th Regional Meeting of IAS* (TIŠLJAR et al., 1983), *the 5th Meeting of Sedimentologists of Yugoslavia* (TIŠLJAR & VELIĆ, 1986), *International Symposium on the Evolution of the Karstic Carbonate Platform: Relations with other Periadriatic Carbonate Platforms* (Trieste, 1987 – VELIĆ et al., 1989), as well as *the 1st Croatian Geological Congress* (Opatija, 1995 – BIONDIĆ et al., 1995; GABRIĆ et al., 1995; TIŠLJAR et al., 1995; VELIĆ et al., 1995a, b), *the 80th Summer Meeting of the Italian Geological Society* (Trieste, 2000 – MATIČEC et al., 2000; SAKAČ & GABRIĆ, 2000; ŠPARICA et al., 2000; TIŠLJAR et al., 2000a, b; TUNIS et al., 2000; VELIĆ et al., 2000; VLAHOVIĆ et al, 2000a, b), the *6th International Congress on Rudists* (Rovinj, 2002 – KORBAR et al., 2002; MORO et al., 2002; TIŠLJAR et al., 2002a, b; VELIĆ et al., 2002; VLAHOVIĆ et al., 2002a, b) and 22nd IAS Meeting of Sedimentology (Opatija 2003 – BERGANT et al., 2003; DURN et al., 2003; VELIĆ et al., 2003; VLAHOVIĆ et al., 2003).

Field trips presented in this Guidebook of the 7th Croatian Geological Congress are based on the traditional division by KREBS (1907), who divided Istria into three large-scale units according to the colour of the prevailing lithology, based on historical Italian names (Fig. 1). **Field trip A1** (GULAM et al., this Vol.) is focused on **Grey Istria** (*Istria grigia*) characterised by turbidite deposits, **Field trip A2** (DURN et al., this Vol.) on **Red Istria** (*Istria rossa*) characterised by carbonate rocks covered by the red Mediterranean soil – *terra rossa*, while participants of the **Field trip A3** (PALENIK et al., this Vol.) will visit the area of the **White Istria** (*Istria bianca*) composed mostly of light-coloured carbonate rocks as well as two interesting quarries located in western Istria.

The geological literature on Istria spans more than two centuries, including some extraordinary studies like Guido Stache's "Die Liburnische Stufe und deren Grenz-Horizonte" (STACHE, 1889). General data on the geology of Istria, including the most important previous publications, can be found in the sheets and explanatory notes of the 1:100,000 scale Basic Geological Map of SFRY, published some sixty years ago: Trst (PLENIČAR et al., 1969, 1973), Ilirska Bistrica (ŠIKIĆ et al., 1972; ŠIKIĆ & PLENIČAR, 1975), Rovinj (POLŠAK & ŠIKIĆ, 1969, 1973), Labin (ŠIKIĆ et al., 1969a; ŠIKIĆ & POLŠAK, 1973), Cres (MAGAŠ, 1968, 1973) and Pula (POLŠAK, 1967a, 1970). The geology of Istria has also been presented in several important papers based on investigations conducted for the Basic Geological Map of SFRY, e.g. POLŠAK (1965a, b, 1967b), ŠIKIĆ & BLAŠKOVIĆ (1965), ŠIKIĆ et al. (1969b), and MAGDALENIĆ (1972). The list of older references should be completed by the discussions of the geology of E–SE Istria, written by Marijan Salopek, especially the classic paper on thrust sheets in Ćićarija and Učka Mts. (SALOPEK, 1954).

From the mid-1960's to the mid-1980's there were no systematic geological investigations of regional significance in Istria. However, it should be noted that, in the 1970's, detailed lithofacies and biofacies investigations of shallow water carbonates in Istria started (e.g., TIŠLJAR, 1976, 1978a, b; SOKAČ & VELIĆ, 1978; TIŠLJAR & VELIĆ, 1986, 1987).

Recent investigations for the new Basic Geological Map of the Republic of Croatia (lithostratigraphic map in 1:50,000 scale) started in Istria in the mid-1980's. In the

We dedicate this work to the memory of our dear friend, the late Professor Josip Tišljar (1941–2009), fellow of the Croatian Academy of Sciences and Arts, who was the fourth member of the Istrian team.

course of these investigations new and interesting results have been obtained, which have been increasingly published in Croatian and international journals (VELIĆ & TIŠLJAR, 1987, 1988; MARINČIĆ & MATIČEC, 1988, 1989, 1991; BARIŠIĆ et al., 1994; MATIČEC, 1994; MATIČEC et al., 1996; TIŠLJAR et al., 1994, 1998; VELIĆ & VLAHOVIĆ, 1994; VLAHOVIĆ et al., 1994). Three sheets of the new Basic Geological Map of the Republic of Croatia covering the territory of Istria have been published so far: Rovinj-1 (MATIČEC et al., 2017), Rovinj-2 (BERGANT et al., 2020) and Rovinj-3 (MATIČEC et al., 2015).

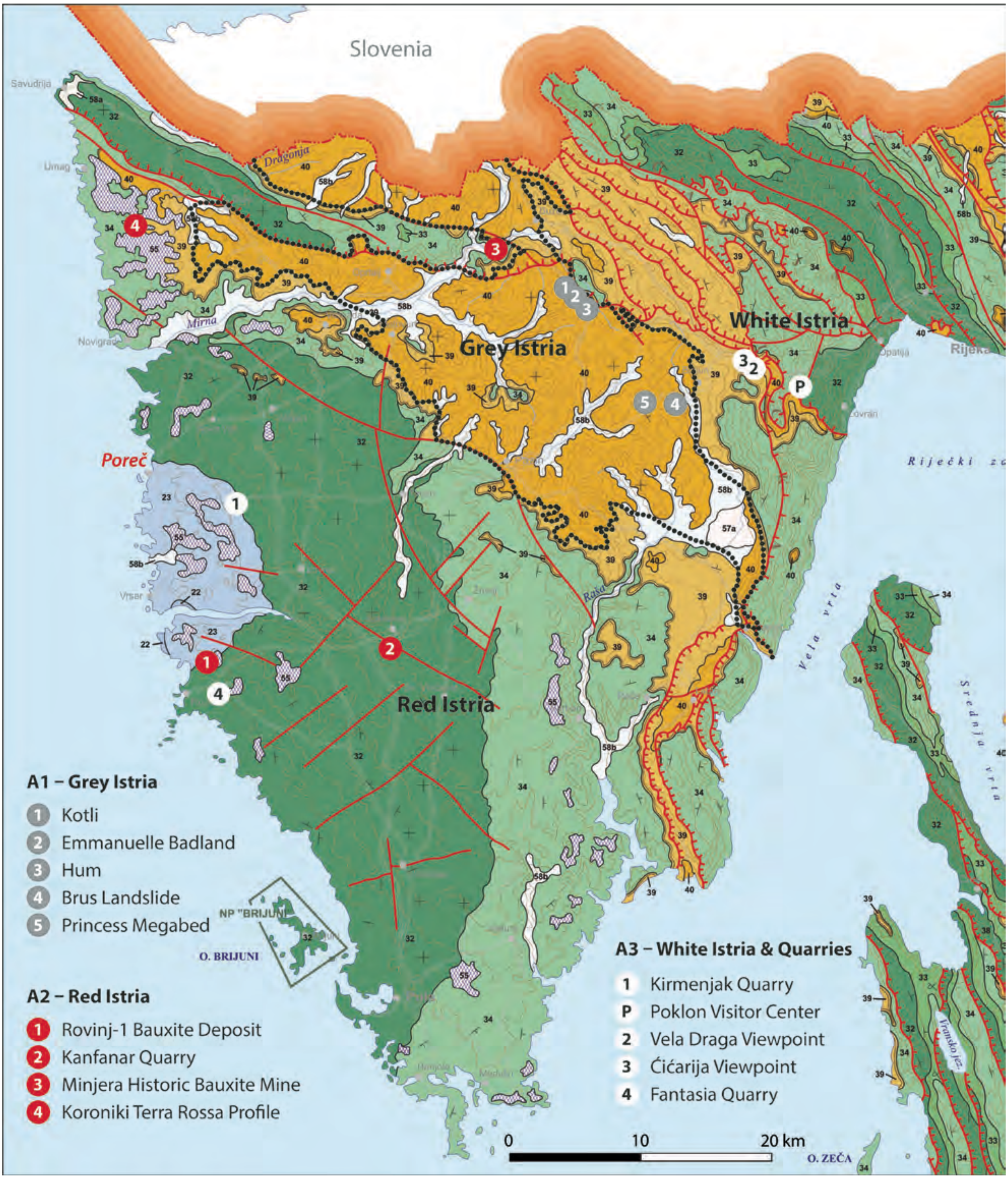

Fig. 1. Geological map of Istria (after CROATIAN GEOLOGICAL SURVEY, 2009) with location of field trip stops: grey circles Field trip A1 – Grey Istria (GULAM et al., this Vol.), red circles Field trip A2 – Red Istria (DURN et al., this Vol.), white circles Field trip A3 – White Istria & Quarries (PALENIK et al., this Vol.). Legend for lithological units (for more details see VELIĆ & VLAHOVIĆ, 2009): 22 – Thick-bedded limestones and dolomites (Middle Jurassic); 23 – Limestones and dolomites (Upper Jurassic); 32 – Limestones and dolomites (Lower Cretaceous); 33 – Dolomites and post-sedimentary diagenetical breccia (upper Albian–lower Cenomanian); 34 – Rudist limestones (Cenomanian–upper Santonian); 39 – Kozina limestones, Foraminiferal limestones and Transitional beds (lower–middle Eocene); 40 – Flysch deposits (middle–upper Eocene); 55 – Terra rossa (Holocene); 57a – Lake deposits (Holocene); 58a – Deluvial–proluvial deposits (Holocene); 58b – Alluvial deposits (Holocene).

New data were obtained also by several specialised investigations (e.g., DALLA VECCHIA et al., 1993, 2000a, b, 2001, 2002; DALLA VECCHIA & TARLAO, 1995; DALLA VECCHIA, 1998; DINI et al., 1998; MEZGA & BAJRAKTAREVIĆ, 1999; DURN et al., 2000; MORO & ĆOSOVIĆ, 2000; MEZGA et al., 2003). Most important recent literature on geology of Istria can be found in this guidebook, in description of field trips **A1** (GULAM et al., this Vol.), **A2** (DURN et al., this Vol.) and **A3** (PALENIK et al., this Vol.).

This introduction to geology of Istria is based on the review papers written as an introduction to field trip guidebooks of the 1st Croatian Geological Congress and 22nd IAS Meeting of Sedimentology (VELIĆ et al., 1995a, 2003).

STRATIGRAPHIC AND PALAEOGEOGRAPHICAL EVOLUTION OF ISTRIA

Although intense synsedimentary and postsedimentary Palaeogene and Neogene tectonics have significantly affected the area of the former Adriatic Carbonate Platform and its overlying deposits, resulting in formation of mostly very complex tectonic structures, there are some localities with quite well-preserved stratigraphic records enabling recognition of important events in the geological history. Istria is probably the most important among them, located on the NW part of the Adriatic coast.

From the geological point of view, Istria can be divided into three regions (Fig. 1):

- the Jurassic–Cretaceous (and partly Eocene) carbonate plain of western and southern Istria,
- the Eocene flysch basin in central Istria, and
- the Cretaceous–Eocene predominantly carbonate area characterised by intense compressional tectonics in eastern and northeastern Istria (Učka and Ćićarija Mts.).

The geological peculiarities of these regions had been noticed historically by the inhabitants of Istria who, as already mentioned, coined specific names for them (KREBS, 1907). *Red Istria* represents the western and southern Istrian plain named after the red Mediterranean soil *terra rossa* covering a large part of the Jurassic, Cretaceous and Eocene carbonates. *Grey Istria* (or sometimes also referred to as *Green Istria* due to the heavy vegetation) is covering the area of central Istria, composed of Eocene turbidite deposits. *White Istria* encompasses area in eastern and northeastern Istria, characterised by karstified outcrops of light-coloured Cretaceous–Eocene limestones.

The Istrian succession consists predominantly of carbonate rocks ranging in age from late Middle Jurassic to Eocene, with subordinate Eocene clastic rocks (turbidites), and Quaternary *terra rossa* and loess deposits. The Istrian late Middle Jurassic to Eocene succession can be divided into four megasequences, i.e. large-scale stratigraphic sequences bounded by important unconformities of different duration, and covered by Quaternary deposits (VELIĆ et al., 1995a, 2003). Evidence for their chronostratigraphic determination will be discussed later. The following large-scale sequences have been distinguished (Figs. 2, 3):

1) M1 Megasequence (Bathonian–lowermost Kimmeridgian);
2) M2 Megasequence (upper Tithonian–lower/upper Aptian);
3) M3 Megasequence (upper Albian–upper Santonian);
4) M4 Megasequence (Eocene).

M1 Megasequence (Bathonian–lowermost Kimmeridgian)

The oldest Istrian megasequence, M1 (Figs. 2, 3), is mainly characterised by a shallowing- and coarsening-upward trend, including in places the *Rovinj breccia* (VELIĆ & TIŠLJAR, 1988), ending in major unconformity U1 with bauxite occurrences and deposits. This Bathonian–lowermost Kimmeridgian megasequence is represented predominantly by different types of shallow-water limestones, which crop out in western Istria, between Poreč and Rovinj (Fig. 2).

During the Bathonian and Callovian restricted shallow subtidal and lagoonal environments prevailed, characterised by medium- to thick-bedded mudstones and fossiliferous wackestones (the *Monsena Unit* – VELIĆ & TIŠLJAR, 1988). Bathonian deposits include evidences of mild synsedimentary tectonics (MARINČIĆ & MATIČEC, 1991). Similar depositional environments continued into the early Oxfordian, with deposition of peloidal packstones and wackestones (the *Lim Unit* – VELIĆ & TIŠLJAR, 1988). During the middle and late Oxfordian prograding sand bars composed of ooids and bioclasts were formed in high-energy shallows and the marginal parts of lagoons (tidal bars of TIŠLJAR & VELIĆ, 1987, or the *Muča Unit* of VELIĆ & TIŠLJAR, 1988). The shallowing-upward tendency continued to the end of the Oxfordian, and during the probable earliest Kimmeridgian resulted in the formation of the regressive *Rovinj* (*Vrsar*) *breccia*, representing the end of this megasequence. Breccia is composed of clasts from the immediate footwall (*Lim* and *Muča limestones*). Complete subaerial exposure and karstification followed, which is shown by the formation of locally high relief associated with an accumulation of source-material for the formation of clayey bauxites. In some places important quantities of bauxite have been formed, e.g., near Rovinj, and to a lesser extent NW from Rovinjsko selo, near Gradina, as well as between Vrsar and Funtana (ŠINKOVEC, 1974). For more data concerning this level see Stops 1 and 2 in VLAHOVIĆ et al. (2003), Stop 1 in DURN et al. (2003) and Stop 1 in DURN et al. (this Vol.).

Middle Jurassic and Oxfordian successions in other parts of the AdCP are also characterised by shallow-water carbonates, deposited mostly within restricted inner parts of the platform (rarely within higher-energy environments as in the southern Adriatic area, with abundant ooids and bioclasts). However, the Kimmeridgian represented a period of significant palaeogeographic changes within the

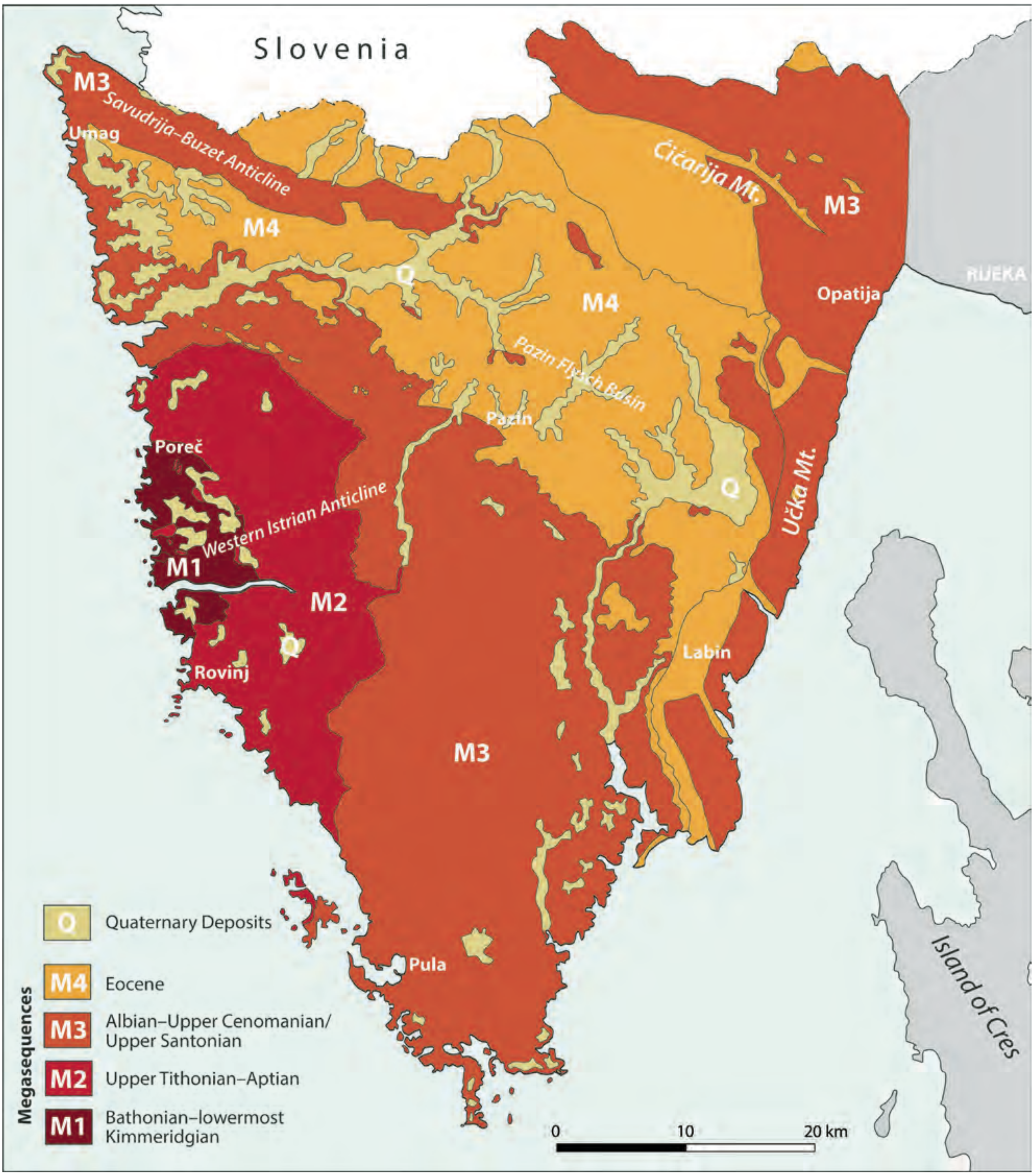

Fig. 2. Map of the Istrian peninsula showing four depositional megasequences separated by regional unconformities, modified after VELIĆ et al. (1995, 2003).

formerly relatively uniform platform realm due to the obduction of ophiolites along the northeastern Adriatic Microplate margin (SCHMID et al., 2008).

The subaerial exposure with bauxite formation in Istria corresponds to the level characterised by significant palaeokarstification in large part of the AdCP (e.g., Kimmeridgian of Biokovo Mt. – TIŠLJAR et al., 1989; TIŠLJAR & VELIĆ, 1991; Stop 8 in BENČEK et al., 2003; NNE part and the platform margin in central and SE Slovenia – DOZET & MIŠIČ, 1997; NW Bosnia – VRHOVČIĆ et al., 1983; E Herzegovina – NATEVIĆ & PETROVIĆ, 1967; W and N Montenegro – VUJISIĆ, 1972; MIRKOVIĆ & MIRKOVIĆ, 1987).

Penecontemporaneously in the central part of the platform deeper troughs with temporary connection to the open sea were formed, characterised by deposition of limestones with cherts and ammonites (FURLANI, 1910; CHOROWICZ & GEYSSANT, 1972; VELIĆ et al., 1994; BUCKOVIĆ, 1995; Stops 2 and 5 in BUCKOVIĆ et al., 2003). These deeper areas were surrounded by reefs

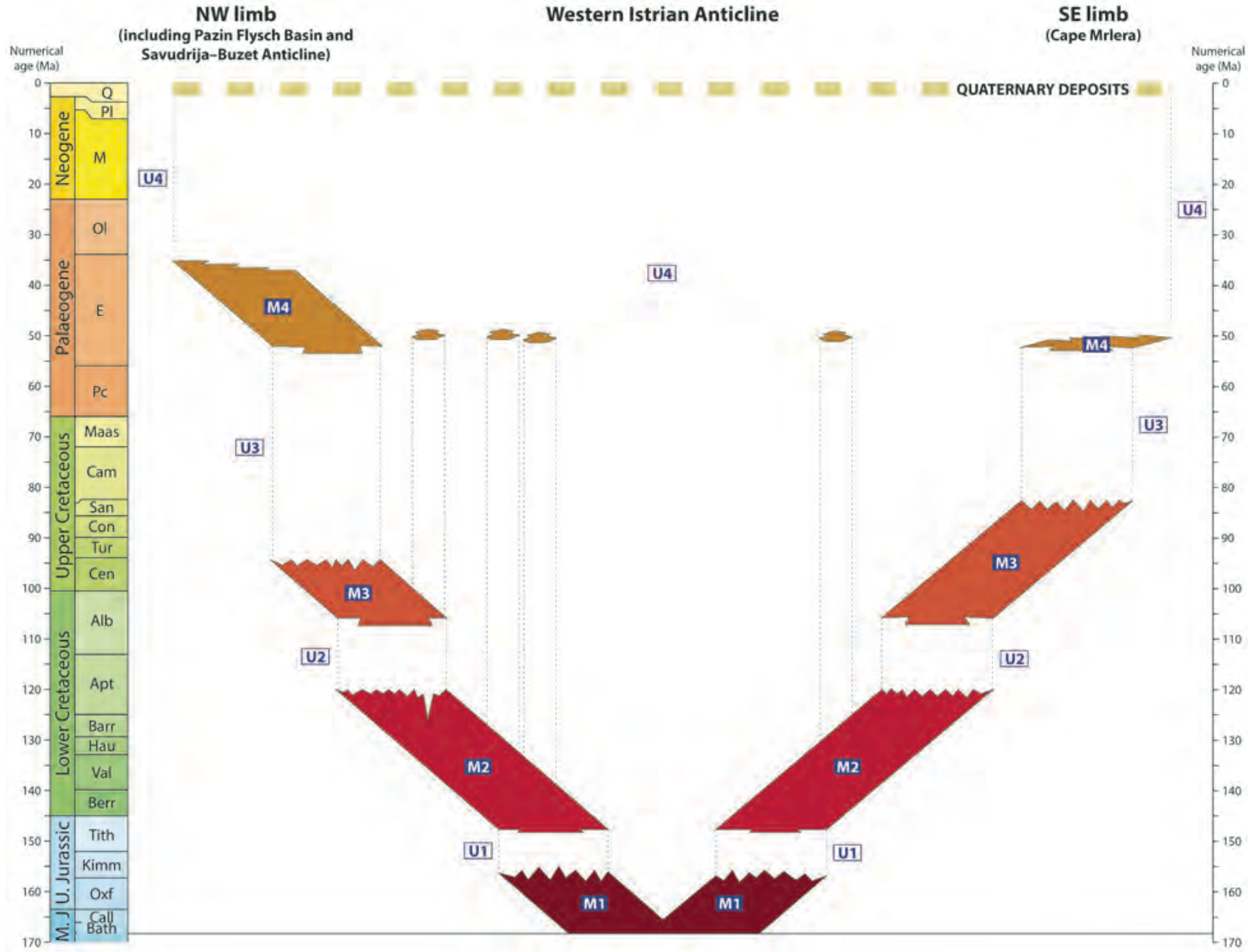

Fig. 3. Schematic geological columns of the Istrian peninsula showing four depositional megasequences (M1–M4) separated by regional unconformities (U1–U4); modified after VELIĆ et al. (2003). Vertical scale in Ma.

and ooid shoals, and were, during the late Kimmeridgian and Tithonian, completely filled by their progradation.

The Kimmeridgian succession along the NE margin of the AdCP is characterised by transgression over the area which was subaerially exposed since Early Jurassic (BUKOVAC et al., 1974, 1984).

During the Tithonian the entire area was again united into a relatively uniform shallow water carbonate platform, representing the beginning of the M2 Megasequence in Istria.

M2 Megasequence (upper Tithonian–lower/upper Aptian)

The second megasequence of Istria is very complex due to its facies heterogeneity and significant thickness. Peritidal deposits characterised by peloid limestones and LLH-stromatolites predominate, with subordinate breccia with clayey matrix formed during short-lasting subaerial exposures (in Tithonian, Hauterivian, Barremian), early- and late-diagenetic dolomites (during Berriasian), and grainstones (bioclastic sand bar deposits typical for upper Valanginian and upper Barremian). Deposits of the M2 Megasequence crop out in the form of an arc (Fig. 2) from Poreč, near Kanfanar and Bale, to the coast from Rovinj to the Brijuni islands.

M2 Megasequence started in the late Tithonian with an oscillating transgression, i.e. dm–m scale shallowing-upward cycles deposited in subtidal, intertidal and supratidal environments (see Stops 1 and 2 in VLAHOVIĆ et al., 2003 and Stop 1 in DURN et al., this Vol.). These limestones, known as architectural-building stone *Pietra d'Istria* or *Kirmenjak*, are composed of black-pebble breccia/conglomerates, mudstones and fenestral mudstones with desiccation cracks (including probably the oldest dinosaur tracks in Istria – MEZGA et al., 2003; Stop 1 in PALENIK et al., this Vol.). The uppermost part of the upper Tithonian limestones is heavily late-diagenetically dolomitised (Stop 1 in VLAHOVIĆ et al., 2003).

Relative shallowing during the Berriasian and older Valanginian resulted in the alternation of subtidal and intertidal limestones which were almost completely late-diagenetically dolomitised and early-diagenetic dolomites deposited in supratidal environments. This alternation of late- and early-diagenetic dolomites is known as the *Fantasia dolomite* (VELIĆ & TIŠLJAR, 1988; see Stop 3 in VLAHOVIĆ et al., 2003 and Stop 4 in PALENIK et al., this Vol.).

In the late Valanginian mainly subtidal parasequences prevailed, sporadically characterised by a coarsening-upward tendency. A similar situation, but with numerous short-lasting subaerial exposure surfaces, continued in the

Hauterivian and a major part of the Barremian, when shallowing-upward cycles were characterised by frequent LLH-stromatolites (see Stop 4 in VLAHOVIĆ et al., 2003), numerous subaerial exposure surfaces and peritidal breccia. Footprints of dinosaurs have been found in such Barremian rocks on the island of Veli Brijuni, as well as bones on the sea floor near the western coast of Istria (Stop 5 in VLAHOVIĆ et al., 2003).

By the end of the Barremian, bioclastic carbonate sand bars characterised by cross-bedding were deposited in shallow subtidal–intertidal environments. The transition to the Aptian was characterised by a change in the depositional system into restricted lower subtidal and/or lagoonal environments with sporadic pelagic influences, due to a relatively important relative sea-level rise connected with oceanic anoxic event OAE-1a (HUCK et al., 2010). Therefore, the lower Aptian deposits are characterised by thick-bedded to massive floatstones with *Bacinella irregularis* oncoids and requieniid rudists (*Toucasia* sp.), which are well-known as the architectural-building stone *Istarski žuti* (*Yellow Istrian* – named after its yellowish colour).

Upper Aptian deposits in Istria indicate a relatively rapid shallowing, resulting in subaerial exposure. This regional unconformity (U2 on Fig. 3) was the result of a relative sea-level fall caused by the interaction of eustatic changes and synsedimentary tectonics on the Istrian part of the Adriatic Carbonate Platform. These movements resulted in the variable duration of shallow-water environments on different parts of the platform, as well as in the different intensity of erosion of the Aptian and Barremian deposits. Finally, the end of the M2 Megasequence was marked by deposition of carbonate breccia and conglomerates, with clay and dark grey swamp deposits, which are well exposed in numerous quarries in western Istria, from Punta Furlan, Baderna, Heraki, Selina, Kanfanar, Bale, Negrin and Barbariga to Veli Brijun. Between Selina and Negrin, uneroded relics of Upper Aptian deposits have been found. The deepest erosion, locally to the Barremian, was reported from west of Heraki to Červar (VELIĆ et al., 1989).

For more information on the youngest, Aptian part of this succession and the regional unconformity U2 see Stop 6 in VLAHOVIĆ et al. (2003), Stops 3 and 4 in DURN et al. (2003) and Stop 2 in DURN et al. (this Vol.).

M3 Megasequence (upper Albian–upper Santonian)

The third megasequence of Istrian deposits (Figs. 2 and 3) is thick and characterised by a very variable facies succession.

After extensive subaerial exposure during the late Aptian and early Albian, at first a gradual, and later a complete ingression occurred during the middle Albian, which was fully accomplished by the beginning of the late Albian. Thus the shallow-water platform carbonate system was re-established for the next 10–20 Ma over the whole area of the Adriatic Carbonate Platform that today belongs to Istria.

Several depositional systems can be recognized within the M3 Megasequence:

a) peritidal and foreshore sedimentary system during the middle and late Albian;
b) significant differentiation of sedimentary systems during the latest Albian and early to middle Cenomanian;
c) drowned platform system during the late Cenomanian and early Turonian;
d) mostly shallow-water sedimentary system during late Turonian, Coniacian and Santonian.

Middle and upper Albian deposits

The lower part of this unit is characterised by an oscillating transgression in the middle Albian, covering the previously completely subaerially exposed area of Istria. In the late Albian a thick sequence of thin-bedded (5–20 cm) grainy limestones was deposited in peritidal and foreshore environments. These are mostly well sorted, fine-grained intraclastic-peloidal packstone–grainstones alternating with foraminiferal-peloid packstones–wackestones, with rare LLH-stromatolites. The uppermost part of the Albian deposits is commonly represented by limestone breccia (mostly formed in peritidal environments during severe storms), grainy limestones with gently inclined cross-bedded sets and current ripples, and diagenetic quartz deposits (Stop 5 in DURN et al., 2003).

Uppermost Albian and lower–middle Cenomanian deposits

The transition from the Early to the Late Cretaceous was marked by the establishment of different sedimentary environments in northern and southern Istria (VLAHOVIĆ et al., 1994).

In northern Istria (the Umag–Savudrija–Buzet area), stable peritidal conditions continued into the earliest Cenomanian. The younger part of the early Cenomanian and the older part of the middle Cenomanian were characterised by the facies differentiation of a formerly united depositional area. In the western part of northern Istria, subtidal mudstones and peloid wackestone–packstones with benthic foraminifera and rudist debris were succeeded by cycles with thicker subtidal members consisting of mudstones and foraminiferal-bioclastic wackestones and thinner intertidal fenestral mudstones and LLH-stromatolites. In the central part of northern Istria, near Marušići, a large prograding carbonate sand body composed of well-sorted bioclastic packstone–grainstones was deposited in upper shoreface and foreshore environments. In the eastern part of northern Istria bioclastic, storm-deposited wackestones, packstones and grainstones, and locally mudstones and chert nodules and layers were deposited on a gently inclined inner carbonate ramp. By the end of the Cenomanian, such a highly differentiated depositional area was filled up, resulting in the re-establishment of united environments characterised by an irregular alternation of light-coloured mudstones and rudist floatstones.

In southern Istria, the latest Albian and early to middle Cenomanian shoreface depositional system was formed, influenced by synsedimentary tectonics, resulting in deposition of slumps, tempestites, carbonate sand and rudist clinoforms and biolithite bodies (TIŠLJAR et al., 1998; VLAHOVIĆ et al., 2011). The boundary between these deposits and the middle to upper Cenomanian limestones is sharp, represented by a change from massive and thick-bedded clinoform bioclastic bodies into thin-bedded (5–25 cm) limestones deposited in low energy shallow-water environments. In the following succession, peloid-bioclastic wackestone–packstones and peloid grainstones predominate, while 30–70 cm thick rudist biostromes and chondrodontid coquinas are infrequent.

A more detailed description and interpretation of the uppermost Albian to middle Cenomanian part of the succession is presented at Stops 7 and 8 in VLAHOVIĆ et al. (2003)

An interesting feature of this unit in southern Istria are chert nodules and lenses found within well-bedded peloid packstones ("limestones with chert" – POLŠAK, 1965b). Although these limestones contain calcitised radiolarians and sponge spicules (which is why they were formerly interpreted as being of deep-marine origin), they appear in a succession of typical peritidal facies. The chert nodules and lenses were formed by early-diagenetic silicification of carbonate mud in low-energy lagoons (TIŠLJAR, 1978a).

Upper Cenomanian and lower Turonian deposits

By the end of the Cenomanian and at the beginning of the Turonian, a drowned platform depositional system was established over most of the Adriatic Carbonate Platform, including southern lstria, resulting in the deposition of "limestones with ammonites" (POLŠAK, 1965b), i.e. mudstone–wackestones with planktonic fauna (common calcispheres and planktonic foraminifera with rare ammonites), as a result of a global eustatic rise related to the oceanic anoxic event OAE-2 (GUŠIĆ & JELASKA, 1993).

These deposits are described at Stop 9 in VLAHOVIĆ et al. (2003).

This drowning event, which left traces on a large part of the Adriatic Carbonate Platform, was not recorded in northern Istria – on the contrary, upper Cenomanian beds became subaerially exposed and covered by bauxites and transgressive Eocene deposits (see Stop 6 in DURN et al., 2003; Stop 3 in DURN et al., this Vol.). The succession of Cenomanian deposits in northern Istria compared to southern Istria indicate the important role of synsedimentary tectonics recorded in neighbouring areas (ŚRODOŃ et al., 2018), which locally modified changes in bathymetry caused by global eustacy to a significant extent (VLAHOVIĆ et al., 1994).

Upper Turonian, Coniacian and Santonian deposits

During the late Turonian, Coniacian and Santonian a shallow-water platform depositional system was re-established over large part of modern-day Istria. It was represented by well-bedded limestones with an alternation of thin layers of mudstone, bioclastic wackestone/packstone and stromatolite laminae in the older part, and mostly thin-bedded rudist coquinas/microcoquinas/biostromes in the younger part of the succession (TIŠLJAR, 1978a). Rudist biostromes are infrequent, because they were mostly destroyed in relatively high-energy environments, with biodetritus deposited in their vicinity. The youngest part of the Cretaceous succession is missing, as a result of a long-lasting subaerial exposure caused by synsedimentary tectonics.

For more information on the upper Santonian deposits see Stop 10 in VLAHOVIĆ et al. (2003).

M4 Megasequence (Eocene)

The fourth megasequence in lstria comprises a relatively thick succession of carbonate and clastic rocks (Figs. 2 and 3). Its greatest part crops out in the Pazin Flysch Basin and neighbouring areas. For more information on the Eocene succession see Stops 11 and 12 in BERGANT et al. (2003) and Field trip A1 (GULAM et al., this Vol.).

Stratigraphic hiatus between the Upper Cretaceous and Eocene deposits (DROBNE, 1977; in places even Lower Cretaceous and Eocene deposits – MATIČEC et al., 1996) is variable from area to area. Different members of the Eocene succession were transgressively deposited over different parts of the Cretaceous succession, due to the synsedimentary tectonic movements (MATIČEC et al., 1996). Consequently, the succession of Eocene deposits is very variable both laterally and vertically. In general, the deposits can be divided into the so-called Kozina deposits, Foraminiferal limestones, Transitional beds and Flysch.

Kozina deposits are only locally present, since they were deposited in the lowest parts of the palaeorelief. They are characterised by deposition during oscillating transgression, and are mostly represented by fresh-water to brackish and lagoonal deposits of the lower Eocene age ("Cuisian"/Ypresian, DROBNE, 1977). These deposits are presented at Stop 6 in DURN et al. (2003) and Stop 3 in DURN et al. (this Vol.).

Foraminiferal limestones in Istria can be divided into three or four lithostratigraphic types deposited from the "Cuisian"/Ypresian to the middle–late Lutetian (ĆOSOVIĆ & DROBNE, 1998), which are mostly in superpositional relationships. These are usually referred to as miliolid, alveolinid, nummulitid and orthophragminid (or discocyclinid) limestones. The foraminiferal limestones are composed of whole and broken larger foraminifera tests, with subordinate detritus of molluscs, ostracods, echinoderms, bryozoans and corallinacean algae, as well as glauconite grains and planktonic foraminifera in the uppermost part. The succession of foraminiferal limestones represents a gradual change of different environments, from the restricted inner part of the carbonate platform (miliolid limestones), through shallower and deeper parts of shoreface environments (alveolinid and nummulitid limestones) to deeper parts of a relatively open carbonate ramp (orthophragminid limestones). Although described varieties are not always of the same age due to the different palaeogeographic position, a general deepening-upward trend is always present. Such a tendency is a consequence of the complex interaction of synsedimentary tectonics and a relatively low sedimentation rate (due to deposition in environments which were no longer ideal for carbonate production, as well as a significant redeposition of the material to deeper parts of the basin).

Transitional beds comprise a range between slope and deep marine deposits. So-called 'marls with crabs' represent the lower part of the Transitional beds as a thin package of nodular-shaped clayey limestones composed of fine-grained matrix with variable amount of glauconite. The fossil content is composed of planktonic foraminifera, bioclasts of benthic organisms and the often well preserved shells of crabs and echinoderms. The upper part of the Transitional beds consists of up to several tens of meters thick massive 'globigerina marls' rich in planktonic foraminifera and glauconite grains, deposited in significantly deeper environments of middle to late Lutetian age (middle Eocene).

lstrian flysch deposits crop out in the Pazin Flysch Basin, Brkini Flysch Basin, at Učka Mt., and partly on Ćićarija Mt. They are generally characterised by an alternation of hemipelagic marls and gravity-flow deposits. The prevailing turbiditic succession of calcareous and calcareous–siliciclastic sandstones and marls is randomly intercalated with thick carbonate debrite beds, usually referred to as megabeds. The flysch deposits are middle to upper Eocene in age (BENIĆ, 1991). The depth of their deposition was concluded to be of bathyal range based on morphotype associations of smaller benthic foraminifera (ŽIVKOVIĆ, 1996). The total thickness of Istrian flysch deposits is estimated at up to 350 m. For more information on turbidite deposits and very important postsedimentary processes recorded in them see Field trip A1 (GULAM et al., this Vol.)

Quaternary deposits

Deposits of all four megasequences of Istrian carbonates and clastic deposits are irregularly covered by a relatively thin Quaternary deposits. The most important are loess and terra rossa (for more information on Quaternary deposits see Stops 2, 7 and 8 in DURN et al., 2003, and Stop 4 in DURN et al., this Vol.), although there are also other types of palaeosols and soils (including swamp deposits – e.g. MEISCHNER, 1995).

BASIC CHARACTERISTICS OF TECTONICS OF ISTRIA

The tectonic pattern of the Croatian part of Istria is composed of three structural units (Fig. 2). The Western Istrian Anticline comprises the largest part of western and southern Istria, being composed of carbonate deposits of the Middle and Upper Jurassic in its oldest part, surrounded by Cretaceous and Eocene carbonates. The area of the Učka Anticline along the eastern margin of the Istrian peninsula represents a more deformed part of the same unit. The second unit, the Pazin Flysch Basin, is composed of a relatively thin Eocene limestones and thick flysch deposits, cropping out in the central and NW parts of the peninsula. The third unit, the only part of Istria belonging to the Dinarides, is composed of structures of Ćićarija Mt. and the Učka Klippe in the northern and eastern part of Istria (see Stop 10 in VLAHOVIĆ et al., 2003, and Stops 2 and 3 in PALENIK et al., this Vol.), built mostly of Upper Cretaceous and Eocene carbonates.

The oldest evidence of tectonic activity in Istria has been found in the upper Bathonian deposits (MARINČIĆ & MATIČEC, 1991). However, Middle Jurassic outcrops can be found only in the coastal area north of Rovinj, and partly in the Limski kanal, excluding the possibility for regional correlation of their effects. These movements resulted with locally subaerially exposed deposits and a more variable relief of the depositional area due to the mild compression oriented 40–220° (in today geographic coordinates).

After a brief stratigraphic hiatus, marine deposition was reestablished until the beginning of the Kimmeridgian, when the area of Istria was subaerially exposed (through the most of the Kimmeridgian and early Tithonian). This regionally well-expressed unconformity (U1 on Fig. 3) was caused by the obduction of ophiolites along the northeastern Adriatic Microplate margin (SCHMID et al., 2008).

Tectonic activity during the Cretaceous was caused by mild synsedimentary tectonic movements occurring throughout the period, because the Western Istrian Anticline had already been formed in the Early Cretaceous (MATIČEC et al., 1996). The oldest evidence of its existence is from the Hauterivian, although elements of synsedimentary tectonics are visible in even older, Berriasian rocks (see Stop 4 in PALENIK et al., this Vol.). The hinge of this gentle brachyanticline structure probably remained subaerially exposed throughout the rest of the Cretaceous, i.e. until the Eocene transgression. There are several evidences supporting that, including:

- The exposed succession NE of the anticline axis towards the Pazin Flysch Basin is more or less continuous from the Bathonian to various levels of the Upper Cretaceous. There is no evidence for intense tectonic activity (the present structure is characterised by bed inclinations of only up to 8°), nor eroded material.
- Eocene foraminiferal limestone outcrops have been found directly overlying the Valanginian, Hauterivian, Barremian, Albian and Cenomanian deposits.

– Bauxites from the footwall of the Eocene limestones were deposited in the palaeorelief of both Cenomanian and Albian limestones. This fact, together with the transgressive Eocene foraminiferal limestones covering deposits of very different ages, excludes the possibility of continuous deposition of a complete Lower and Upper Cretaceous succession and its subsequent complete erosion prior to deposition of the Eocene foraminiferal limestones.
– The subaerially exposed area was inhabited by dinosaurs from Tithonian to the Maastrichtian as indicated by their footprints and bones. Such animals needed a large quantity of fresh potable water and terrestrial vegetation for their survival.
– Similar succession can be traced offshore towards the SW, along the extension of the core of the Western Istrian Anticline (VESELI, 1999).

All determined Cretaceous synsedimentary tectonic movements and the structures that they formed indicate the same orientation of stress, in accordance with the orientation of the Western Istrian Anticline. The hinge of the Western Istrian Anticline dips northeastward (towards 35°) and the greatest regional stress were oriented 125–305°.

By the end of the Cretaceous, almost the entire Adriatic Carbonate Platform, including the area of Istria, was subaerially exposed, with stratigraphic hiatus of very variable duration. The Late Cretaceous tectonic events initiated the disintegration of the former carbonate platform area, and marked the end of typical, productive platform carbonate sedimentation, since renewed marine conditions in the Eocene were mostly controlled by intense synsedimentary tectonics.

The Eocene transgression was a consequence of a new deformation. The intensity of these movements was witnessed by the formation of foreland basins filled by turbiditic deposits, as synorogenic deposits indicating the final uplift of the Dinarides. The result of this tectonic activity is visible along the Adriatic coastline by the so-called Dinaric strike of structures (NW–SE). However, as already mentioned and discussed in more details in description of Stop 2 of the Field trip A3 (PALENIK et al., this Vol.), only a minor part of Istria was included into the Dinarides, i.e. the areas of Western Istrian Anticline (including the Učka Anticline along the eastern margin of the Istrian peninsula) and Pazin Flysch Basin were not intensely tectonically deformed and they represent undeformed part of the Adriatic Microplate, i.e. Dinaric foreland. Deformation of the contact area with the Dinarides (Ćićarija Mt. and Učka Klippe), indicate several phases of deformation: (1) formation of the Pazin Flysch Basin, a typical foreland basin filled by turbiditic deposits; (2) formation of the Učka Anticline along the eastern margin of Istria with specific N–S strike and Eocene flysch deposits uplifted to 1,000 m asl; (3) refolding of the northern part of the Učka Anticline structures by typical Dinaric deformation (main stress oriented NE–SW) and formation of the complex Ćićarija Mt. fold and thrust structures; (4) thrusting of Učka Klippe (today an erosional remnant of the assumed much larger Učka Nappe structure) characterised by a very specific lithology on top of the Učka Anticline; (5) intense weathering of the Ćićarija Mt. area resulting in erosion of other parts of the assumed Učka Nappe.

Neotectonic deformation in the entire External Dinarides is a consequence of the N–S oriented greatest regional stress. Neotectonic activity comprised either the formation of new neotectonic structures, reactivation of inherited old brittle structures into the regional strike-slip faults, or rotation of already existing structures to adjust to the new stress orientation.

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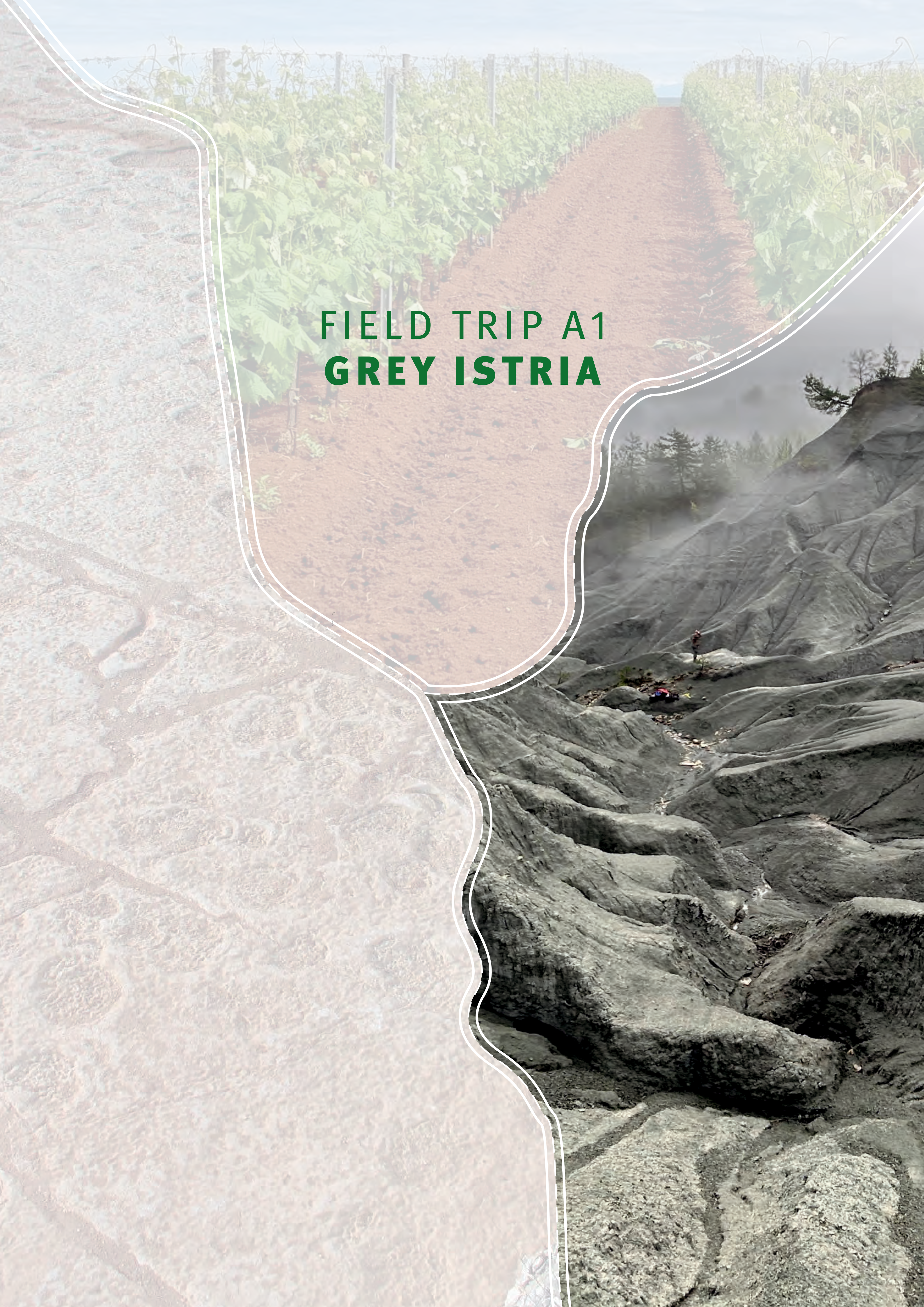
FIELD TRIP A1
GREY ISTRIA



FIELD TRIP A1 – GREY ISTRIA

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INTRODUCTION: EXPLORING THE DYNAMIC LANDSCAPE OF GREY ISTRIA

The scenic beauty of Central Istria is characterized by a balanced blend of grey and ochre colours, harmoniously set amidst lush greenery. The region's incised and hilly relief, ancient towns atop prominent hills, a diverse array of colours, and tranquil ambience owe their existence to its unique and dynamic geology. This field trip offers insights into the fusion of geological foundation and landform evolution, unravelling the interconnected narrative of geology and geomorphology. By studying this facet of Istria, we recognize that comprehending nature's extraordinary connections necessitates collaboration across diverse scientific disciplines. To gain an understanding of the region's geomorphic aspects, we delve into its geological foundation through the lenses of stratigraphy, lithology, mineralogy, and tectonics.

The main objective of this field trip is to explore Central Istria, renowned for its characteristic flysch deposits, which will be closely examined throughout the excursion, from two key perspectives. Firstly, we will investigate the stratigraphic evolution and the remarkable diversity of these flysch deposits, given their direct role in landscape formation. Secondly, we will study the ongoing geomorphological processes responsible for shaping the landscape formed by these deposits.

The field trip encompasses five stops, each carefully selected to showcase various sediment types. These stops demonstrate distinct geomorphological processes that have contributed to shaping the unique relief visible on the surface today, ultimately defining the terrain known as "Grey Istria" (Fig. 1).

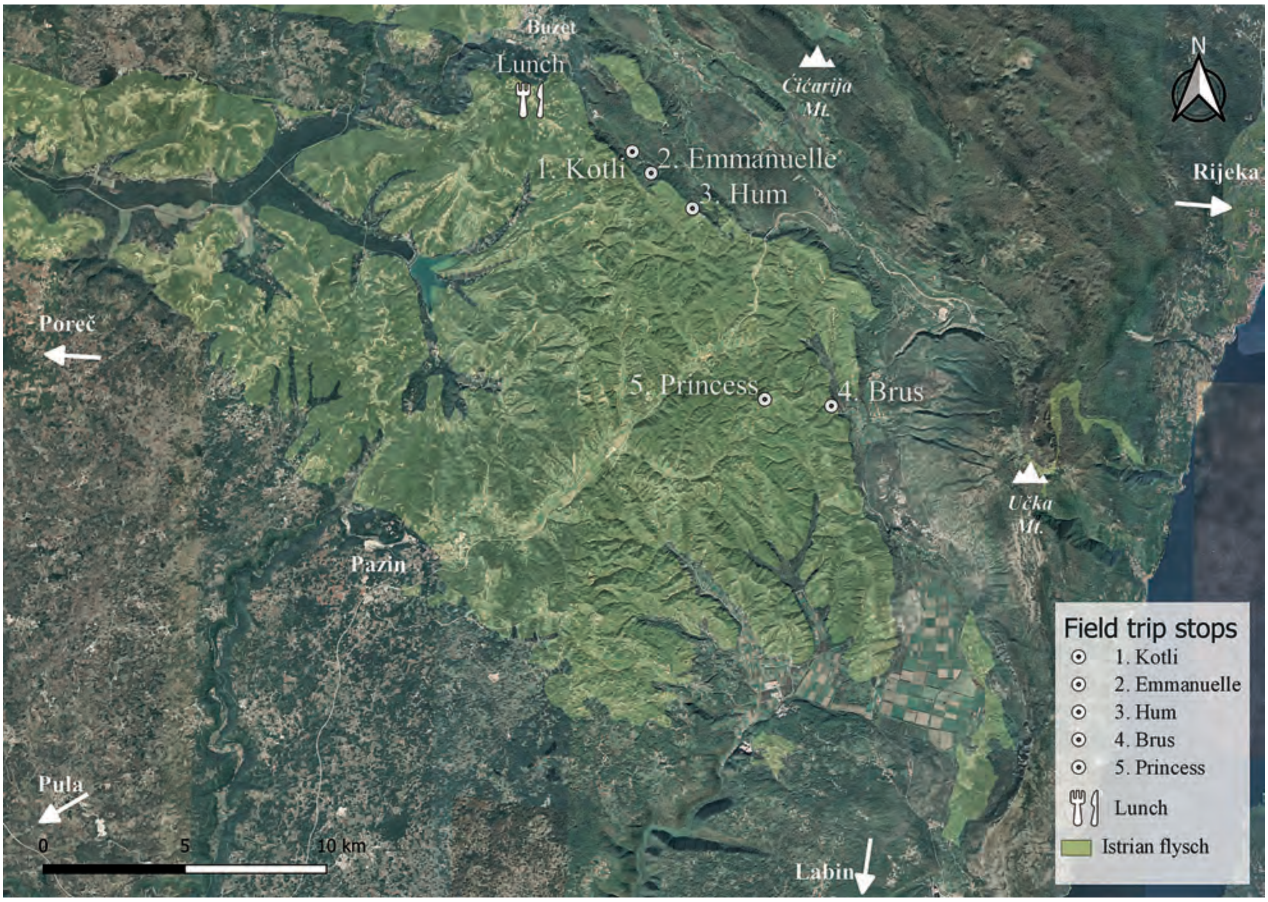

Fig. 1. The area of northwestern Istria and field trip stops

Paleogene Succession in the Istrian Peninsula: A Tale of Tectonic Processes and Sediment Deposition

The Istrian peninsula primarily belongs to the Adriatic foreland (SCHMID et al., 2008), with the northeastern mountainous region forming part of the External Dinarides' fold-and-thrust belt (Fig. 2). This mountain belt was formed during the Alpine orogenesis, a consequence of the collision between the Adriatic microplate (Adria) and the Eurasian plate (SCHMID et al., 2008; KORBAR, 2009).

Stratigraphically, Istria features a thick succession of Jurassic and Cretaceous limestones and a Paleogene foreland basin sedimentary sequence (Fig. 2). The Jurassic and Cretaceous limestones were deposited on a pre-orogenic Adriatic carbonate platform that existed long before the major geological event, the Alpine orogeny, took place (VLAHOVIĆ et al., 2005). They are separated from the Paleogene foreland basin deposits by a regional unconformity characterized by the presence of palaeokarst formed during the uplift of the Adriatic Carbonate Platform in the distant foreland, during the Late Cretaceous and Early Palaeogene (OTONIČAR, 2007). Additionally, a broad anticline in the Istrian region likely formed during the Late Cretaceous (Fig. 2), as suggested by MATIČEC et al. (1996), based on the observation that the deformed and palaeokarstified Cretaceous carbonates are overlain by the Palaeogene succession in a diachronous manner.

During the Paleogene, a foreland basin was formed in northwestern Istria as a result of tectonic activities caused by the Alpine orogeny and the formation of the External Dinarides mountain belt. This basin's sedimentary succession includes carbonate ramp deposits, referred to as Foraminiferal limestones and the Istrian flysch, originating from the foredeep part of the basin (DECELLES & GILES, 1996).

The lower Paleogene succession is distinguished by localized Kozina beds, either freshwater or brackish (ŠIKIĆ & POLŠAK, 1973; MARJANAC & ĆOSOVIĆ, 2000; ĆOSOVIĆ et al., 2008), followed by Foraminiferal limestones ranging from the Ypresian to Middle-Late Lutetian age (DROBNE, 1977; MARJANAC & ĆOSOVIĆ, 2000). These limestones were deposited on a fully marine carbonate ramp in the distant part of the Dinaric foreland basin (ĆOSOVIĆ et al., 2004; OTONIČAR, 2007). Four distinct members of Foraminiferal limestones are often identified as Miliolid-, Alveolinid-, Nummulitid-, and Discocyclina limestones (DROBNE, 1977; ŠIKIĆ & PLENIČAR, 1975; VELIĆ et al., 2003; VLAHOVIĆ & VELIĆ, 2009), stacked on top of each other, indicating progressive deepening of the carbonate ramp (ĆOSOVIĆ et al., 2004). Transitional beds above the Foraminiferal limestones mark the gradual shift from the neritic carbonate ramp to deeper-water distal foredeep environments (ŠIKIĆ & PLENIČAR, 1975; MARJANAC & ĆOSOVIĆ, 2000; TARLAO et al., 2005). Above, the Globigerina marl unit is present, a monotonous succession of grey marls up to 80 meters thick (ŠIKIĆ & PLENIČAR, 1975). This unit is considered Transitional deposits (MARJANAC & ĆOSOVIĆ, 2000; ĆOSOVIĆ et al., 2008), true deep-sea sediments (JURAČIĆ, 1979) or part of the flysch formation (MARINČIĆ et al., 1996; BERGANT et al., 2020).

The flysch formation is characterized by the alternation of hemipelagic marls and gravity-flow deposits, primarily turbidites. These turbidites consist of either mixed carbonate-siliciclastic detritus (hybrid) or carbonate detritus (MAGDALENIĆ, 1972; MARINČIĆ, 1981; BABIĆ & ZUPANIČ, 1996; MARINČIĆ et al., 1996; BERGANT et al., 2003). The mixed carbonate-siliciclastic turbidites indicate a paleotransport direction from the rising Dinarides and the Alps toward the east-southeast (MAGDALENIĆ, 1972; MARINČIĆ et al., 1996; BABIĆ & ZUPANIČ, 1996) and are considered as deposits of low-density turbidity currents (MARINČIĆ et al., 1996). Conversely, carbonate beds can range from thin turbidites to thick, bipartite beds composed of breccias, conglomerates, bioclastic arenites/siltites, and marls (BABIĆ & ZUPANIČ, 1996; BERGANT et al., 2003). These carbonate beds exhibit a paleotransport direction toward the north-northeast, generally perpendicular to the southeast direction of the mixed carbonate-siliciclastic turbidites (BABIĆ & ZUPANIČ, 1996). BABIĆ & ZUPANIČ (1996) propose that the carbonate detritus originated from the uplifted foreland (forebulge) located to the south, characterized by carbonate shoals and subaerial exposure.

According to the majority of published literature, the age of the Istrian flysch is considered to be Middle to Late Eocene, spanning from the Lutetian to the Early Priabonian. While there have been a few published (MIKES et al., 2008) and unpublished studies suggesting a late Oligocene or even Miocene age for the Istrian and Dalmatian flysch, the prevailing consensus among researchers supports the Middle Eocene age (KRAŠENINNIKOV et al., 1968; PICCOLI & PROTO DECIMA, 1969; AUBOUIN et al., 1970; MAGAŠ, 1973; POLŠAK & ŠIKIĆ, 1973; ŠIKIĆ & POLŠAK, 1973; ŠIKIĆ & PLENIČAR, 1975; BENIĆ, 1991; ŽIVKOVIĆ & BABIĆ, 2003; BABIĆ et al., 2007; ŽIVKOVIĆ & GLUMAC, 2007).

While the majority of Istrian flysch belongs to the tectonically undisturbed Adriatic foreland tectonic unit (Fig. 2), in the vicinity of the fold-and-thrust belt of the External Dinarides (Ćićarija Mt. and Učka Mt.), the flysch deposits can be slightly folded. This can be observed in the inclined turbidite beds at stops 1 – Kotli, 2 – Emmanuelle, and 3 – Hum.

Lithology: Decoding the Istrian Flysch – Marl-rich Landscape of Central Istria

In the stratigraphic succession and based on the deposits' composition and sedimentary facies, the Istrian flysch can be subdivided into distinct zones or units (MULD-

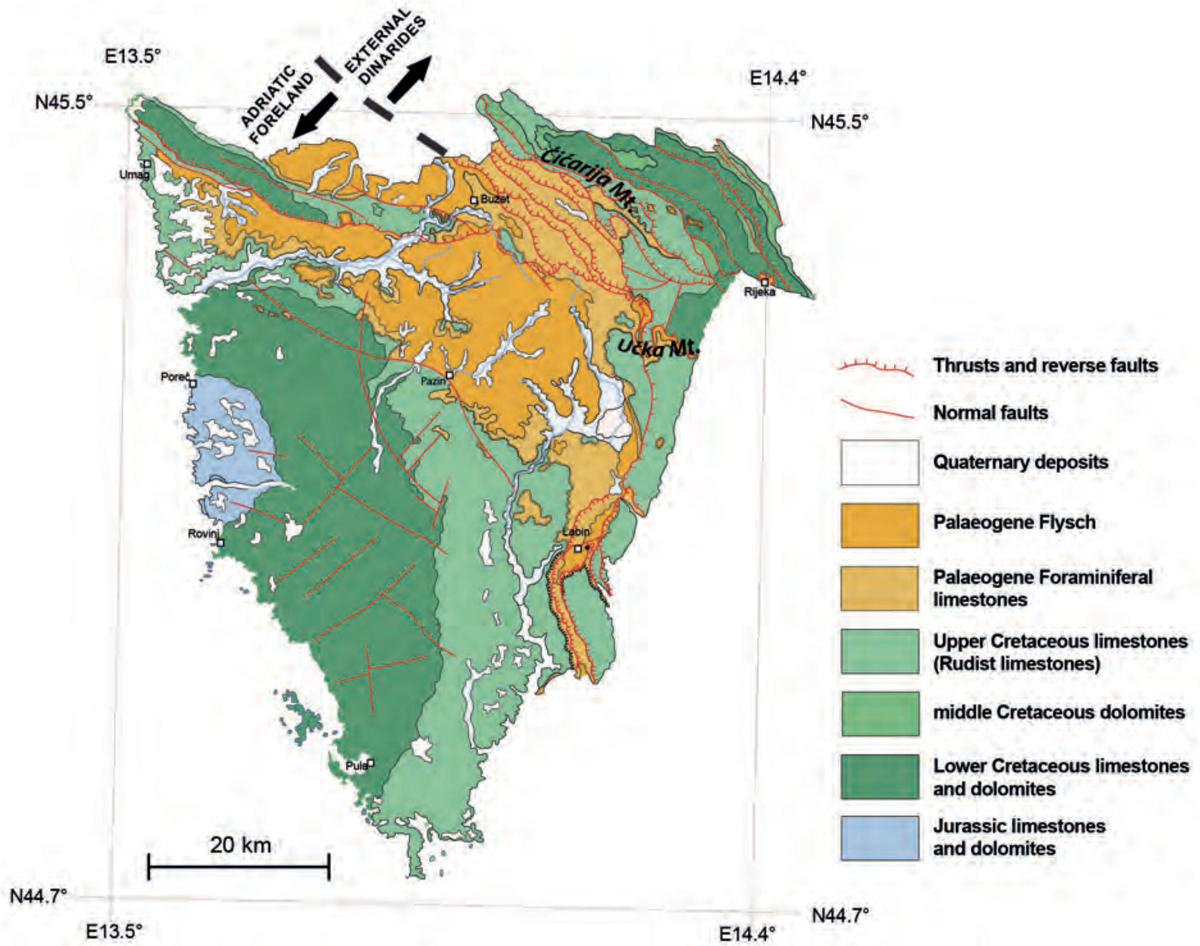

Fig. 2. Geological map of the Croatian part of the Istrian peninsula at a scale of 1:300 000 (CROATIAN GEOLOGICAL SURVEY, 2009)

INI-MAMUŽIĆ, 1964; MARINČIĆ et al., 1996; BABIĆ et al., 2007; PETRINJAK, 2021).

MARINČIĆ et al. (1996) identified three separate units within the Istrian flysch: a) Globigerina marls, b) the lower part of the Istrian flysch, and c) the upper part of the Istrian flysch. The transition from Globigerina marls to the lower part is characterized by the appearance of thicker arenite layers. The lower flysch unit mainly comprises carbonate turbidites composed of bioclastic material that varies in thickness, with some reaching up to 12 meters, and they may include breccias (debrites) in the lower part. Moving on to the upper part of the Istrian flysch, it is described as pockets, possibly due to limited distribution, formed by hybrid (mixed siliciclastic-carbonate) sandstone turbidites. These hybrid turbidites consist of a hybrid arenitic part, approximately 10 cm thick, along with a marl interval ranging from 10 to 40 cm in thickness.

According to BABIĆ et al. (2007), the flysch deposits are categorized as: a) Basal marls, b) the Middle unit, and c) the Upper unit. The Basal marls correspond to the lithostratigraphic unit of Globigerina marls and are characterized by massive marls indicating relatively pure pelagic sedimentation, with occasional calcarenite layers. The Middle unit consists of alternating intervals of marl and carbonate clastites. Based on planktonic foraminifera communities, this zone is classified in the upper part of the P12 biostratigraphic zone. The Upper unit is characterized by frequent alternations of turbidite sandstones and marl intervals, resulting in a sequence of thickening upward. The turbidites are composed of arenites with a higher siliciclastic component. The presence of sandstones with a higher siliciclastic component marks the introduction of extrabasinal siliciclastic (terrigenous) material into the basin.

In summary, the Istrian flysch can be classified into three units: a) Globigerina Marls (the Lower unit), b) the Middle unit, and c) the Upper unit (Fig. 3). The Lower unit is characterized by massive marls. The Middle unit displays alternating hemipelagic marl and occasional carbonate turbidites primarily composed of intrabasinal bioclastic material from carbonate ramps located on the shallower basin margins. The Upper unit consists of alternating mixed siliciclastic and carbonate turbidites and marls, indicating the influx of extrabasinal siliciclastic material into the basin.

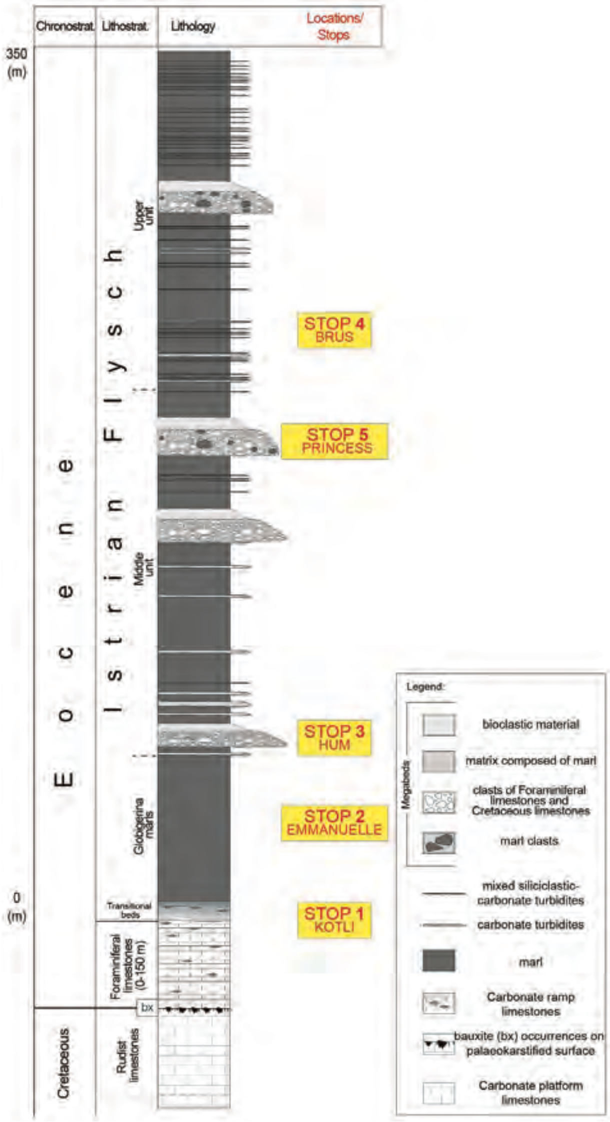

Fig. 3. Schematic geological column depicting the Middle Eocene Istrian flysch succession and the underlying Cretaceous and Paleogene carbonates, indicating the general relative stratigraphical positions of the excursion stops. Intervals of underlying Transitional beds, Foraminiferal limestones, and Cretaceous formations are not presented to scale in relation to the flysch. Modified after PETRINJAK et al. (2021).

The lithological characteristics discussed provide a profound answer to the question of why this Istrian region is named "Grey Istria". The distinctive grey colour that defines the landscape originates from the lithological composition itself, with the prevalence of marl throughout the entire flysch succession being the primary contributing factor.

Geomorphological Features of Grey Istria: Discovering Badlands

Grey Istria evidence highly dynamic exogenetic processes, resulting in a diverse relief characterized by a dense hydrographic network and numerous steep-sided valleys. The intensity of these processes is evident, and from a geological perspective exceptionally rapid, leading to significant changes on the terrain's surface within engineering timeframes. These intensive exogenetic processes contribute to the formation of spectacular geomorphological landforms, known in the literature as badlands (BRAYAN & YAIR, 1982). The badlands have a dual impact on society, manifesting in the following directions: Badlands as an Engineering Challenge and Badlands as Emerging Geotourism Destinations.

Badlands as an Engineering Challenge

In the mid-20th century, a clear societal need to address exogenetic processes affecting the area of Grey Istria led to the first research focused on torrential flows, co-funded by the UN-FAO organization (PETRAŠ et al., 2007). The impact of erosion, as the dominant exogenetic process in Grey Istria, was evident even before this research began. This influence encompassed a range of engineering challenges confronted by the community, among which the filling of the channels and artificial lakes with sediments, along with the degradation of agricultural land, proved to be the most problematic. The first research findings redirected the attention of both the professional and scientific community towards areas termed "badlands". The results revealed that badlands yield 8000 times more sediment than all other areas covered with some form of vegetation (JURAK & FABIĆ, 2000). Considering that badlands cover approximately 2.5 % of Grey Istria (BOSTJANČIĆ et al., 2023) and exert a pronounced impact on the dynamics of exogenetic processes, research efforts became specifically directed at these zones, which are also recognized as hotspots of excessive erosion (JURAK et al., 2002).

The initial research aimed, among other things, to determine sediment production across various surfaces covered with different types of vegetation. This approach utilized suspended sediment capture methods on test plots containing various forms of vegetation. However, this approach proved to be ineffective on badland surfaces due to the high erosion intensity, causing the deterioration of test plots shortly after the measurements commenced (PETRAŠ et al., 2008).

To overcome this problem, experts began exploring remote sensing techniques to determine sediment production on badland surfaces, particularly photogrammetric methods, which proved valuable in areas with low or no vegetation cover. One of the first studies was carried out at the Sveti Donat badland, revealing an annual sediment production of 20.7 dm^3/m^2, corresponding to an average annual surface retreat of 2.1 cm (JURAK et al., 2002). Encouraged by these promising results, along with technological advancements (digital cameras), other researchers have begun to apply remote sensing techniques to quantify badland surface retreat. Subsequent research within the area of Boljun badland demonstrated the advantage of digital technology, determining an average annual retreat of approximately 5 cm per year (GULAM et al., 2018).

Recently, the development of LiDAR sensors and unmanned aerial systems (UAS) has further revolutionized the approach to investigating excessive erosion in badland areas. This has led to the publication of the first works employing these innovative technologies on the Istrian badlands (FRANGEN et al., 2019), along with some preliminary results being presented within this guide. Scientists from the Croatian Geological Survey continue their research of excessive erosion in the Grey Istrian badlands, using state-of-the-art of RGB and LiDAR sensors and UAS technologies. The findings and results of these ongoing studies are expected to be published soon.

Badlands as Emerging Geotourism Destinations

Badland areas represent spectacular geomorphological phenomena that, due to their surreal appearance, have increasingly attracted a large number of tourists in recent times (MARTÍNEZ-MURILLO & NADAL-ROMERO, 2018; ZGLOBICKI et al., 2019). These visitors encompass nature enthusiasts or adventurers drawn to activities such as motocross, quad riding and ziplining. While the presence of geotourism activities in these locations undoubtedly affects ongoing research, it would be counterproductive to attempt to prevent them, as they undeniably contribute to the development of tourism in these areas. Nevertheless, due to the beauty of certain locations, it would be valuable to consider their preservation as geological heritage. This would serve not only their aesthetic value but also establish them as potential models for larger and more complex geomorphological systems (WAINWRIGHT & BRAZIER, 2011).

The Dance of Natural Forces: Exploring the Dynamics of Exogenetic Processes in Grey Istria

The question arises regarding the cause of the intense exogenetic processes in Grey Istria and the factors contributing to the complex relief characterized by deeply incised valleys and numerous gullies. The main exogenetic process is water erosion, involving concentrated surface water flow with varying erosive capacity (energy) across different temporal and spatial scales. This type of relief is conditioned by a specific geological composition, which in turn triggers a complex cycle of exogenetic processes (weathering – denudation /erosion and mass movements/ – deposition), especially within badland surfaces (GULAM et al., 2014). The paramount role in the development of the distinctive relief of Central Istria undoubtedly belongs to marls, a lithological unit that occupies a significant portion of the Istrian flysch succession as shown in Fig. 3. In this regard, the following physical-mechanical characteristics of marls should be highlighted:

- Low to moderate uniaxial compressive strength (according to most UCS classifications; BIENIAWSKI, 1984) – Fresh marl, which has not undergone processes of mechanical weathering, belongs to the group of poorly cemented rocks with uniaxial compressive strength values ranging from 10 to 30 MPa;
- Low durability – Marls are composed of 20 to 80 % of calcareous component and 20 to 80 % of clayey component. The clayey component renders them vulnerable to mechanical weathering, causing rapid disintegration under repeated wetting and drying, heating and cooling, as well as freezing and thawing (the main processes of mechanical weathering). In Central Istria, such processes occur on the terrain's surface leading to the relatively rapid formation of a surface regolith that almost entirely covers the flysch deposits;
- Low permeability – Due to the high proportion of clay minerals, marl is characterized by low permeability, enabling the rapid formation of surface flows.

Low resistance to mechanical weathering processes (Low durability) of marl deposits leads to the rapid formation of a surface regolith layer. Due to the low permeability of marl, the dominant member of the flysch complex in the Grey Istrian region, surface flows form relatively quickly after each rainfall event. These surface flows erode the regolith layer and incise deep valleys and gullies with very steep side slopes. The strength of fresh marl is high enough to maintain the steepness of side slopes. Simultaneously, these steep side slopes support the formation of highly erosive surface water flows capable of transporting the sediment produced by the mechanical weathering of marl. From this elaboration, it is evident that marls play a key role in sustaining this very intriguing and somewhat enigmatic chain of exogenetic processes, contributing to the formation of badland surfaces (stop 2 – Emmanuelle) and stop 5 – Princess) and, in some cases, supporting landslide formation (stop 4 – Brus landslide).

STOP 1 – KOTLI: THE BEGINNING OF DEEP-SEA SEDIMENTATION IN THE MIDDLE EOCENE BASIN OF ISTRIA

Kotli serves as the initial stop on this field trip, marking the stratigraphically lowest segment of the Istrian flysch succession (Fig. 3). Here, we can distinguish the Transitional beds that represent a gradual transition from neritic limestones formed on the Paleogene carbonate ramp (Foraminiferal limestones) to deposits formed within the deeper marine environment (Globigerina marls).

Throughout the history of research, the term "Transitional beds" was not always unambiguously used to describe the sediments we now associate with it. It previously referred solely to "Marls with crabs" (ŠIKIĆ et al., 1969) or even encompassed both "Marls with crabs" and "Globigerina marls" (MAGDALENIĆ, 1972). JURAČIĆ (1979) concluded that the term "Transitional beds" should include only "Marls with Crabs", excluding "Globigerina Marls", which represent typical deep-sea sediments. Based on the ratio between planktonic and benthic foraminifera across thin sections ranging from "Nummulitic limestones"

Fig. 4. Location of the Transitional beds at Kotli

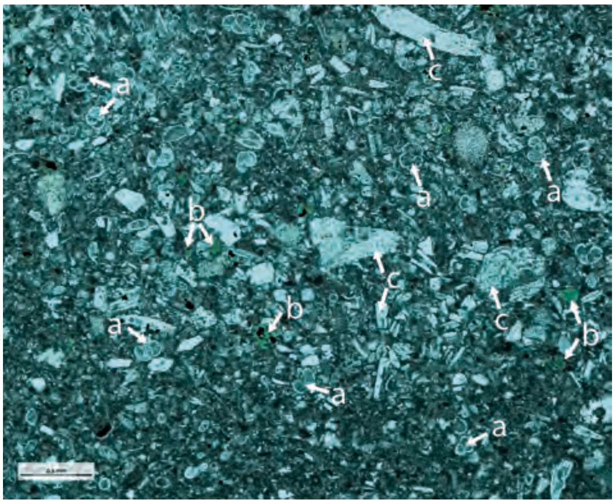

Fig. 5. Photomicrograph showing a bioclastic packstone with planktonic foraminifera (a), glauconite (b), and carbonate fragments of shallow-water fauna (c)

to the upper part of "Marls with Crabs", JURAČIĆ (1979) determined a continuous increase in sedimentation depth. The same author (JURAČIĆ, 1979) assumed that the upper part of "Nummulitic limestones" has been deposited in a shallow sea (20–60 m), the lower part of "Marls with Crabs" in depths ranging from 60 to 100 m, while the upper part of "Marls with Crabs" indicated an increase in depth to several hundred and even a thousand meters. Due to their locally high glauconite content, the "Marls with crabs" have also been called "Glauconitic limestones" (MAGDALENIĆ, 1972). Macrofossils are quite common and usually well-preserved, and in addition to decapod crustaceans include bivalves, gastropods, crinoid plates, shark teeth and vertebrae, nautiloids, corals, brachiopods, corals and nummulites (TARLAO et al., 2005).

In the Mirna riverbed (Fig. 4) we can observe the limestone beds with nummulites, orthophragminids, pelagic foraminifera, glauconite and rare bivalve macro-fossils. Toward the northeastern part of the riverbed, there is a gradual increase in the clay content accompanied by the appearance of beds with a bulbaceous habitus, characteristic of the Transition beds. Microfacies of the Transitional beds rich in planktonic microfauna are presented in Fig. 5. Transitional beds in the Kotli area gradually progress into a zone of massive marl, which can be seen next to the road traversing from the village of Kotli westward to Krušvari (Fig. 4). Along this road, at a hypsometrically higher position, the transition from massive marls to a zone of alternation of marl and turbidites (flysch) becomes evident.

STOP 2 – EMMANUELLE: THE MOST IMPRESSIVE BADLAND OF GREY ISTRIA

In the Palm of Your Hand: Globigerina Marls

At the second stop, the Emmanuelle badland, we can observe an exposed sequence of massive marl approximately 60 meters thick, belonging to the Globigerina marl unit. The Globigerina marl unit presents a monotonous succession of grey marls, occasionally with thin turbidite beds, which become more frequent towards the upper part, marking the transition to flysch (Fig. 3). According to HORVAT et al. (2023, accepted manuscript), the carbonate part of these marls originates from biochemical processes. However, it also contains fine carbonate debris that was re-deposited from a shallow carbonate ramp. On the other hand, the siliciclastic fraction of the marl is rich in clay, with around 10 % comprising silt and sand-sized particles. This detrital material primarily resulted from weathering in nearby continental source areas. The clay composition indicates the dominant influence of physical weathering, characterized by a considerable amount of illite and chlorite. Nevertheless, traces of climatic seasonality are evident in the presence of smectite-interlayered phases and occasional spikes in kaolinite content (HORVAT et al., 2023, accepted manuscript). Occasionally, on the weathered slopes of the marl outcrops, you can notice the repetitive alternations of two types of marls with varying resistance to weathering. Alternation of such intervals forms (parallel) lamination. LUŽAR-OBERITER et al (2010) ascribed these differences between the marl types to climatic variations, likely linked to Milankovitch oscillations. Considering the fossil content, marls are rich in planktonic foraminifera.

The Globigerina marls are considered Middle to Late Lutetian in age (BERGANT et al., 2003) and according to ŽIVKOVIĆ & BABIĆ (2003), these deposits in Pazin (Istrian) basin has been assigned to the lower part of the planktonic foraminiferal Zone P12 of BERGGREN et al. (1995), corresponding to Zone E10 of BERGGREN & PEARSON (2005). This age aligns with nannoplankton dating performed by BENIĆ (1996), who determined the NP15 – NP16 nannoplankton zone.

The Circular Chain of Exogenetic Processes Unfolding Before Our Eyes

Emmanuelle stands as the most stunning badland site in Grey Istria (Fig. 6). Apart from its exotic appearance, this location offers an exceptional opportunity to delve deeper into the circular chain of exogenetic processes. Due to sparse vegetation cover, it also serves as an incredibly suitable platform to test the applicability of remote sensing methods in assessing the rate of denudation processes, i.e., measuring the intensity of exogenetic processes on badland surfaces.

This stop's description will present the first findings of extensive research undertaken by scientists from the Croatian Geological Survey, which includes test plots scattered across nine badland sites in Grey Istria. The primary aim of these studies is to determine the denudation rate in the badlands of Grey Istria, and the preliminary results presented below vividly illustrate the rapid changes in the badlands' relief. This knowledge will undoubtedly shed light on the intensity of these processes, which are closely tied to specific geological compositions.

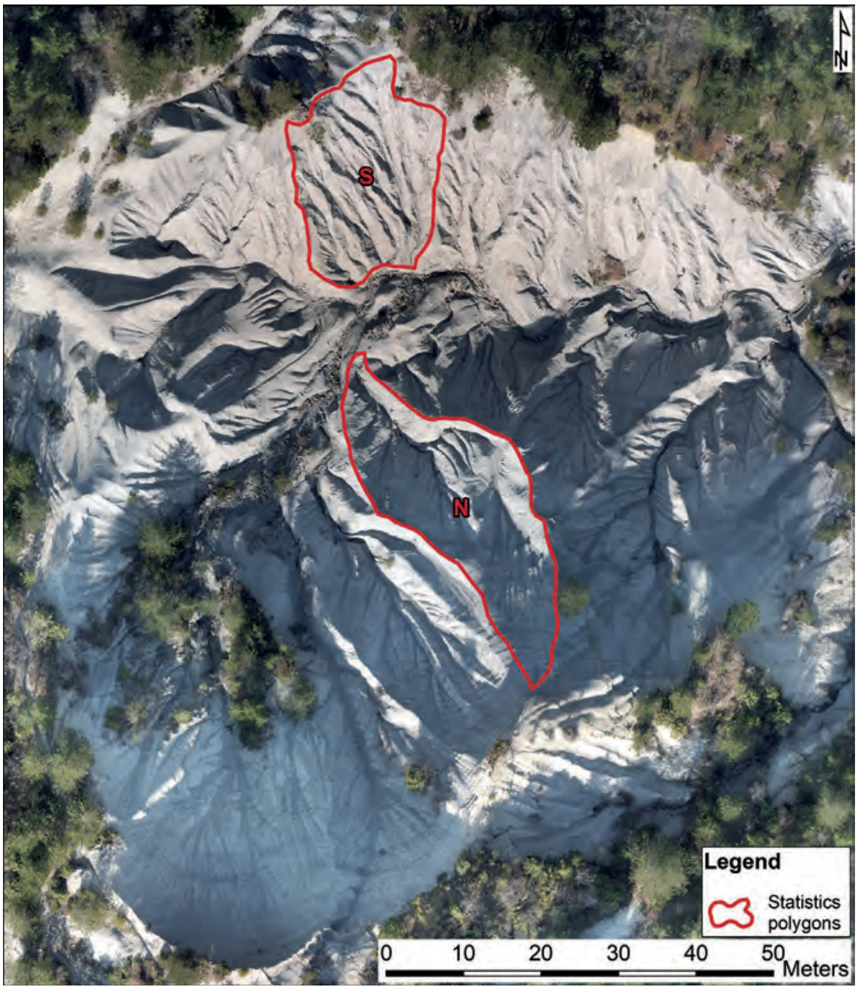

Fig. 6. The Emmanuelle badland site featuring outlined analysed polygons (S – southern aspect polygon; N – northern aspect polygon)

Unravelling denudation rate using remote sensing techniques

Data acquisition at Emmanuelle badland employed the unmanned aerial vehicle (UAV) DJI Matrice 300, equipped with an RGB sensor. The presented results arise from the analysis of two missions, the first in March 2022 and the second in July 2023. Denudation is determined by subtracting two digital surface models using the ArcGIS raster calculator tool and represents the denudation rate spanning around 16 months.

At this location, we focused on analysing polygons with both southern (marked by the letter 'S' in Fig. 6) and northern (marked by the letter 'N' in Fig. 6) aspects. These polygons were selected to minimize the potential influence of vegetation on exogenetic processes, and eventually, small vegetation-covered areas were excluded from the analysis.

From the data provided in Table 1, it can be noticed that the average denudation rate, represented as a vertical land retreat, amounts to 6.2 cm for polygon S and 4.2 cm for polygon N, which corresponds to remarkable annual rates of 4.65 cm and 3.15, respectively. These denudation values undoubtedly confirm the assertion found in various literature sources, strengthening the fact of Grey Istrian badlands as areas of excessive erosion. The spatial distribution of denudation rates on Emmanuelle badland is shown in Fig. 7.

Table 1. Average denudation rate on two analyzed polygons at Emmanuelle badland

Polygon	Polygon area [m²]	Average denudation rate [cm]	Standard deviation
S	420.2	6.2	6.5
N	518.1	4.2	4.3

Table 1 also reveals a discrepancy in denudation rates, indicating higher values in the polygon with a southern aspect, by approximately 2 cm. This intriguing anomaly can be explained by varying intensities of mechanical weathering along the opposing slope. Here is a hypothesis that supports the statement made in the previous sentence but needs to be tested at other badland locations.

Mechanical weathering, as the initial link in the circular chain of exogenetic processes, degrades the mechanical properties of the marl's surface, making it even more erodible (susceptible to erosion). Since we are analyzing the same material with similar morphometric parameters, it can be hypothesized that the intensity of mechanical weathering on the south-facing slopes is higher than on the north-facing slopes. A logical explanation would be that more frequent cycles of wetting and drying, freezing and thawing, as well as heating and cooling, contribute to more intense impacts of mechanical weathering on south-facing slopes. Such intensity of mechanical

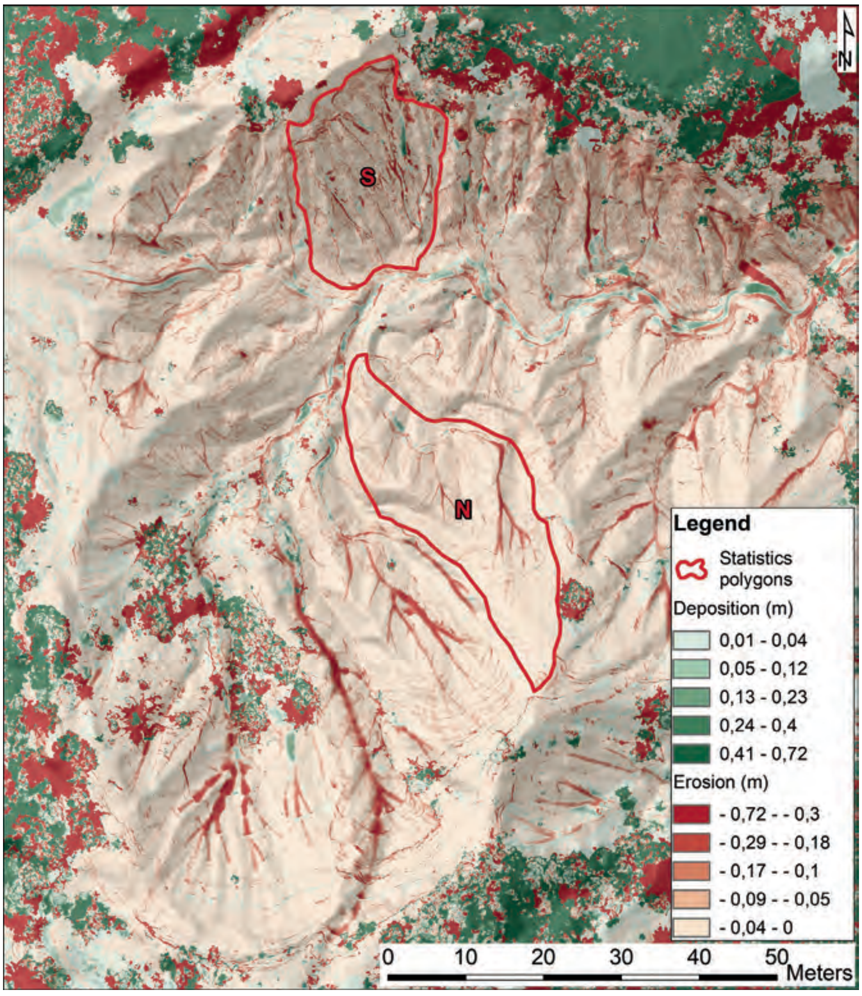

Fig. 7. Vertical differences for two Digital Surface Models with 16 months difference. Since DSMs contain vegetation, the biggest differences are contributed to it. The two polygons (N and S) are digitally "cleaned" from vegetation, so they are DTMs (Digital Terrain Model). Red represents the lowering of the surface, and green heightening (deposition).

weathering generates a more erodible surface zone, where water, as the erosive agent with similar erosive capacity, causes a greater denudation rate.

This brief but highly informative analysis demonstrates the complexity of the circular chain of exogenetic processes in Grey Istria. It highlights that each link within this chain (weathering – denudation /erosion and mass movements/ – deposition) presents a fascinating yet demanding scientific challenge.

STOP 3 – HUM: ISTRIAN FLYSCH AND MEGABEDS

The third stop of the field trip corresponds to a Middle flysch unit (Fig. 3) and is characterized by a sedimentary section approximately 40 meters thick, located near the town of Hum. This section consists of alternating layers of marl and carbonate beds. Among these beds, the most significant one is the megabed Hum, spanning 10 meters in thickness (Fig. 8). Due to the recent road expansion, new insights into the structure of this bed have come to light.

The lower part of the Hum megabed, considered as the debrite part, is composed of breccia/conglomerate, while the upper part, referred to as the high-density turbidite part, consists of normally graded calcirudite and calcarenite. The maximum thickness of the laterally thinning breccia/conglomerate interval is 10 m. The majority of visible clasts vary in size from a few cm to 0.75 m in diameter (Fig. 8). Boulders/clasts are predominantly (sub) rounded, although angular clasts are rarely present (PETRINJAK et al., 2021). The conglomerate/breccia is mostly clast-supported, although we find an interval with a larger amount of matrix and large marl clasts. The matrix is composed of marl, small lithoclast fragments and foraminiferal debris (nummulitid and orthophragminid tests). The major lithological components are Cretaceous limestones (88 %) while Foraminiferal limestones (9 %) and other constituents (3 %) are less abundant. We can rarely find clasts of chert and reddish coarse-crystalline limestone. Also, the Lower Cretaceous limestone clasts were only observed in this megabed. The turbidite part consists of normally graded calcirudite and calcarenite, with the following composition: orthophragminids (32–54 %), nummulitids (6–26 %), and red algae (9–13 %). Similarly, the carbonate turbidites are also composed of normally graded calcirudite and calcarenite, sharing the same main constituents but in varying proportions. The consistent composition suggests that the carbonate detritus source for both the turbidites and the megabed was the same (PETRINJAK et al., 2021).

The carbonate turbidites are classified as high-density turbidites. They are normally graded and composed of bioclastic calcirudite/calcarenite, with bed thicknesses ranging from 0.6 to 3 meters. The lower bedding plane of each turbidite is sharp and erosional, implying that the gravity flow eroded previously deposited basinal sediments. Flute casts observed on the lower bedding plane of one of the turbidite indicates a paleotransport direction towards 165° (SE), while ripple marks observed on the upper bedding plane of the Hum megabed suggest a direction towards 195° (SW). Also, the ripple marks found on the uppermost turbidite show a paleotransport direction towards 230° (SW).

According to PETRINJAK et al. (2021), the composition of megabeds of Istrian flysch suggests that the detritus originated from different parts of the Cretaceous to Palaeogene neritic carbonate succession and from the flysch itself, with marl clasts found in some megabeds (Fig. 9). It is likely that synsedimentary faults, forming along carbonate ramps in the fast-evolving distal foredeep, triggered submarine failures. These faults exposed older carbonates along the fault scarps, leading to erosion along steep slopes. Subsequent slope collapses along the fault scarps transported large quantities of carbonate material, including bioclastic detritus from synchronous carbonate ramps and clasts from older neritic limestones, and also reworked the basinal marls. Earthquakes related to the tectonic evolution of the approaching Dinaric orogen could have been the main triggering mechanism for the submarine collapses. The megabeds are randomly distributed in the flysch stratigraphy, without any specific facies associations, suggesting a seismically triggered slope collapse. Unlike in the Middle Dalmatian basin, where megabeds are associated with accelerated sea-level rise and interbedded with fan deltas, the Istrian megabeds do

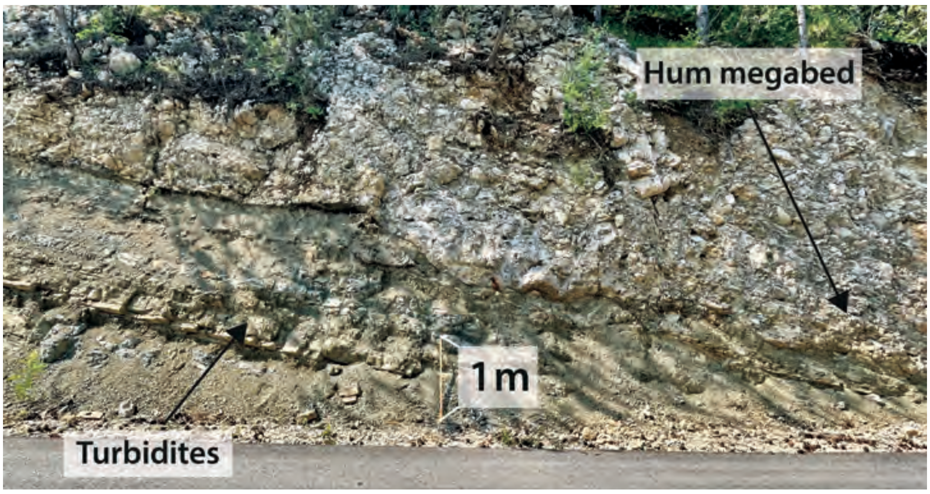

Fig. 8. Outcrop showing carbonate turbidites and the debrite part of the Hum megabed

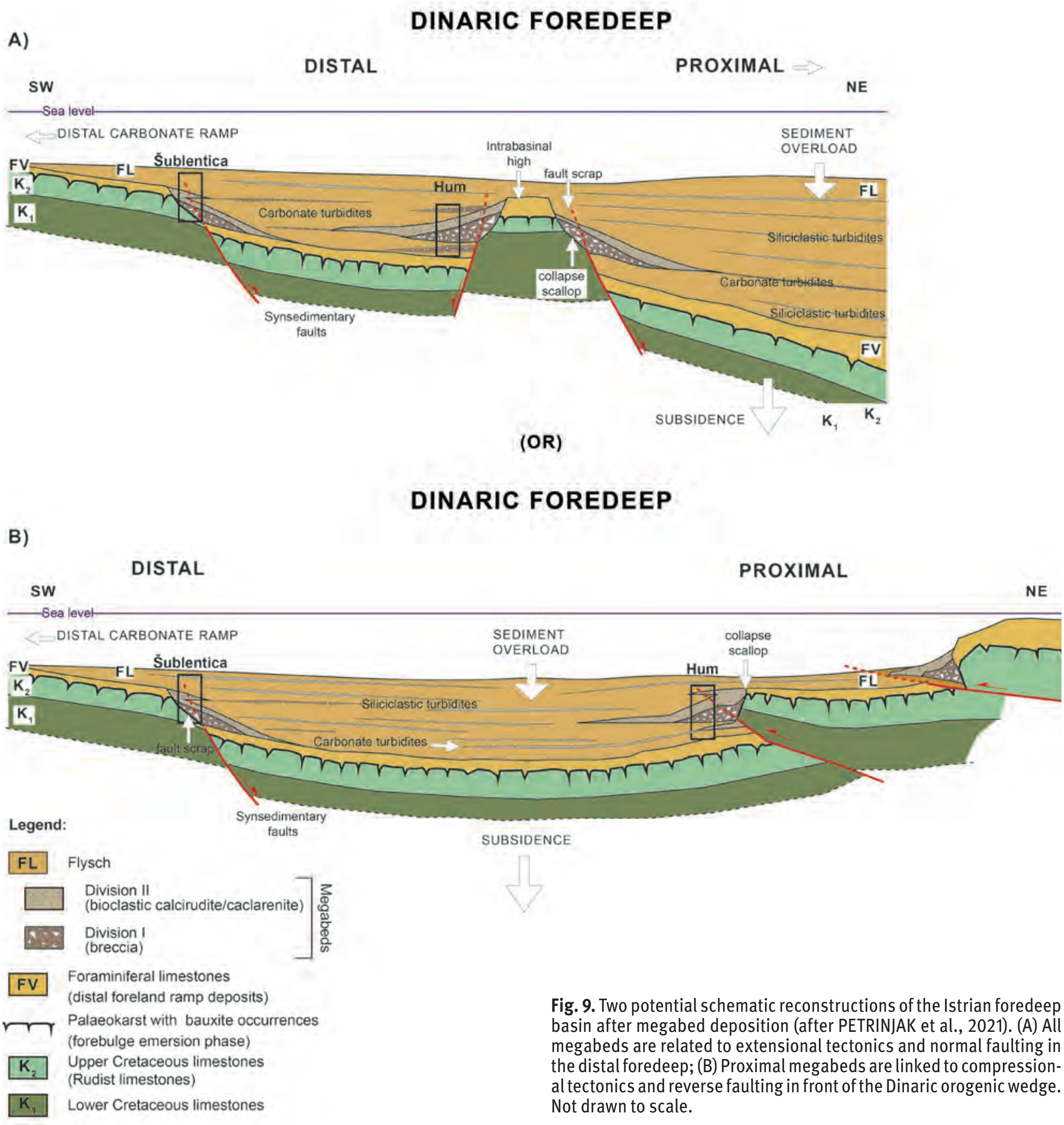

Fig. 9. Two potential schematic reconstructions of the Istrian foredeep basin after megabed deposition (after PETRINJAK et al., 2021). (A) All megabeds are related to extensional tectonics and normal faulting in the distal foredeep; (B) Proximal megabeds are linked to compressional tectonics and reverse faulting in front of the Dinaric orogenic wedge. Not drawn to scale.

not show such patterns. Although the link between global sea level changes and the deposition of Istrian megabeds is not yet established, a possible sea level drop during the Middle Eocene could have impacted the re-sedimented carbonates in Istria (PETRINJAK et al., 2021).

STOP 4 – BRUS: CIRCULAR CHAIN OF EXOGENETIC PROCESSES AND UNIQUE GEOLOGICAL SETUP UNFOLDS THE LANDSLIDE

The Grey Istria, among its prevalent badlands, is also marked by the occurrence of landslides. Various landslide types are evenly distributed throughout the entire area (DUGONJIĆ JOVANČEVIĆ, 2013), where abundant rainfall and human activities present their main triggers. The fourth stop of this excursion is situated at the landslide location, which was activated in the vicinity of the village of Brus, on the road between Boljunsko Polje and Paz (Fig. 10).The landslide was activated on April 9, 2005, after a heavy rainfall event that was identified as the main triggering factor. The translational sliding block defines the movement pattern of the Brus landslide, which is approximately 150 meters long and 33 meters wide (DUGONJIĆ JOVANČEVIĆ & ARBANAS, 2012). Although rain is considered the main trigger for the activation of this landslide, it is important to emphasize that the landslide was brought to a state of unstable equilibrium due to the interaction of an unfavorable geological setup and the circular chain of exogenetic processes.

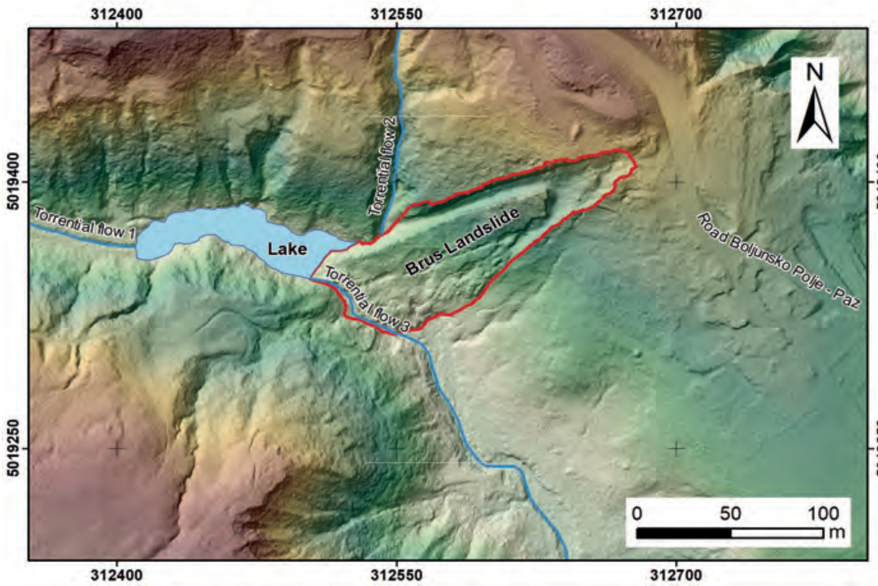

Fig. 10. Brus landslide polygon mapped on LiDAR DTM

The Brus Landslide Activation Reveals a Fresh Outcrop in the Upper Istrian Flysch Succession

Activation of the Brus landslide has exposed a geological outcrop, revealing the typical geological features of the upper part of the Istrian flysch succession (Fig. 3). At this location, we can observe the alternating beds of hemipelagic marl and turbidites (Fig. 11). According to the low ratio of arenite to marl thickness, one can assume that they were deposited in a distal setting. Most of the turbidite beds consist of laminated and cross-rippled sandstones, interpreted as the uncomplete Bouma sequences (Tb-e, Tc-e & Td-e), which supports the distal character of described sediments. Also, some beds appear homogeneous and have no visible structures. The largest turbidite detritus is the size of coarse sand. Flute casts, groove casts and various ichnofossils are visible on the lower bedding planes. These turbidites are regarded as turbidites deposited by low-density turbidity currents (LOWE, 1982; MARINČIĆ et al., 1996). Submarine channel sediments have not been described, so deposits of these facies are considered sediments of the basin plain or distal parts of turbidite fans – according to MUTTI & RICCI LUCHI (1972).

Petrographically, the turbidite sandstones have various carbonate-siliciclastic ratios and can be described as calcarenaceous sandstones (with > 50 % extrabasinal siliciclastic components), bioclacarenites (with more than 50 % fossil carbonate detritus), and calclithite (more than 50 % of carbonate grains derived from older carbonates; PETTIJOHN et al., 1972). Among the siliciclastic component subangular quartz grains, lithoclasts of chert and quartzite predominate, while grains of feldspar, micas and heavy minerals are less frequent. Resistant minerals are often well-rounded, indicating relatively long transport (BERGANT et al., 2003). According to MAGDALENIĆ (1972), among the accessory heavy minerals in flysch sandstones, garnet, zircon and tourmaline dominate, rutile and chromite (chrome spinels) are secondary, while epidote, staurolite, titanite, glaucophane, chloritoid, coisite and brookite

Fig. 11. The Brus landslide. Apart from the landslide, the outcrop shows an alternation of sandstones and marly intervals. A traffic sign marked with a grey circle can be used for the scale.

occur in small quantities. Based on this composition, MAGDALENIĆ (1972) concludes that the detrital material of the flysch series originated from older sedimentary rocks (Mesozoic volcanogenic-sedimentary complex, Paleozoic and Triassic clasts) and ultrabasic and basic eruptive and rocks of low and high metamorphism. The carbonate component is present as fragments of micritic limestones, crystalline calcite and as a fine-grained matrix. Fossil fragments are represented with deep-water fossils (planktonic foraminifera) or sporadically redeposited carbonate ramp fauna (fragmented benthic foraminifera corallinaceans, gastropods, corals, bryozoans, echinoderms, etc.). Mixed carbonate-siliciclastic turbidites show longitudinal paleotransport directions toward the ESE which led to the conclusion that the material was supplied from the rising Dinarides and the Alps (MAGDALENIĆ, 1972; BABIĆ & ZUPANIČ, 1996; MARINČIĆ et al., 1996).

According to MARINČIĆ et al. (1996), the ichnofossil assamblages within the turbidite beds are extremely diverse and are comparable to the Nereites ichnofacies, indicating an ecologically stable and well-oxygenated, moderately oligotrophic environment. The Nereites community is characteristic of deep-sea environments.

Diving into the Formation of the Brus Landslide

To explain the activation of the Brus landslide, it is necessary to highlight two highly unfavorable conditions within the landslide zone, whose mutual interaction brought the slope into a state of unstable equilibrium: the unfavorable layer's orientation (bedding planes (sub)parallel to the slope) and a highly erosive stream in the toe of the slope (Fig. 12).

To illustrate the impact of these two unfavourable factors more vividly, a schematic profile of the Brus landslide is created, simplifying the representation of the action-reaction principle. It demonstrates the increasing consumption of reactive forces as the gully deepens at the base of the slope. In this greatly simplified concept, it should be emphasized that the reactive forces actually arise from the strength of the rock mass, while the action forces result from the weight of the engaged volume of the rock mass.

Over time, erosion by the torrential stream at the base of the slope, caused the gully to deepen, engaging a larger and larger volume of rock mass in increasing the action force, which in turn consumes more and more of the reactive force, i.e., the strength of the rock mass. This denudation process of water erosion, thus brought the slope from a state of stable equilibrium to a state of unstable equilibrium, provoking the activation of another denudation process – the landslide (mass movement process).

It is important to note that this process explains the transition of the slope from a state of stable equilibrium into a state of unstable equilibrium, where the action forces and the reactive forces are approximately equal in magnitude. The transition from a state of unstable equilibrium to instability was likely triggered by heavy rainfall. This rainwater infiltrated the rock mass, increasing its water saturation and subsequently adding to the weight of the already engaged slope volume. As a result, the action force increased. Simultaneously, the increased pore pressure reduced the effective strength of the rock mass, leading to

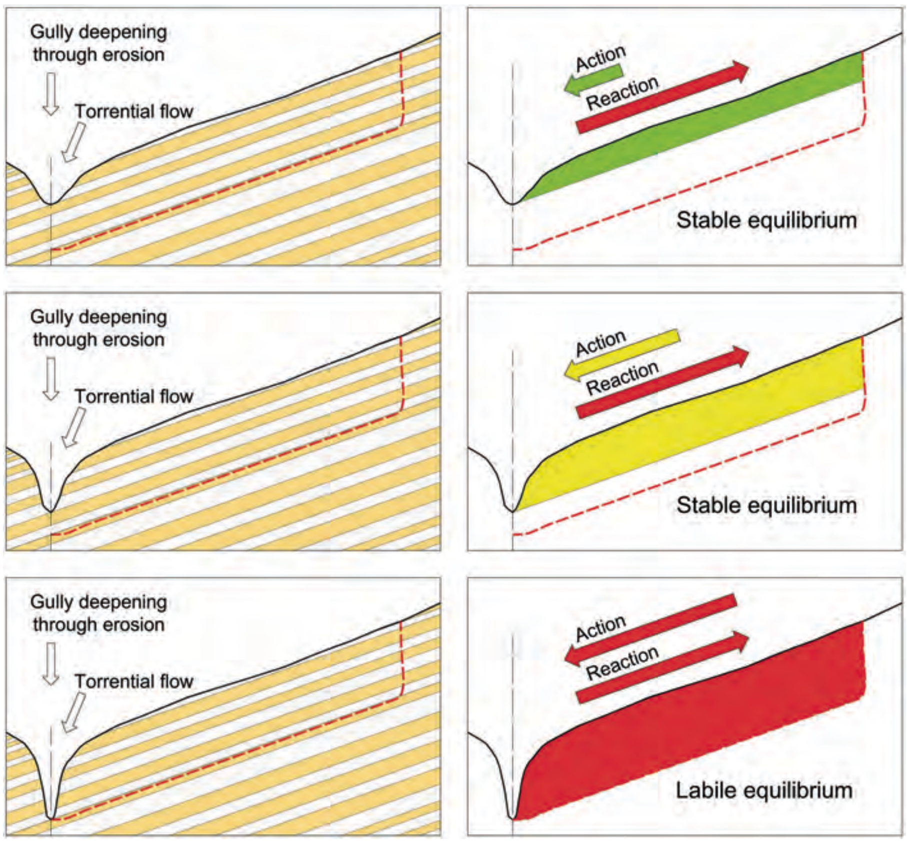

Fig. 12. Conceptual insight into the Brus landslide activation

a decrease in the reactive force. These combined factors contributed to the shift from unstable equilibrium to instability.

From this example, it is possible to conclude that factors leading a slope into a state of unstable equilibrium belong to the group of exogenetic processes, which are, however, somewhat more long-lasting. These factors create conditions in which triggers such as earthquakes, extreme climatic events, and human activities shift unstable slopes to instability.

STOP 5 – PRINCESS'S MEGABED: A GEOLOGICAL ODYSSEY

The fifth stop is located west of the village of Paz and is represented by the Princess badland, situated within a marly interval (Fig. 13) placed above the five-meter-thick mega-bed (Fig. 14), illustrating a situation typical of the Middle or Upper unit of the Istrian flysch succession (Fig. 3). The thick marl interval within the flysch formation could also be a result from the deposition of a big amount of fine-grained material transported by turbidity currents. Indeed, situated at the bottom of this 20 m thick marl interval is a substantial, more than 5-meter-thick megabed (Fig. 14). It is considered that this marl was partly deposited by the tail end of the turbidity current responsible for the deposition of the megabed. This substantial megabed holds great significance in the geomorphology of Grey Istria. Furthermore, its existence leads to the creation of two important geomorphological situations responsible for the development of the Central Istrian relief (GULAM, 2012).

Situation 1 – Badland Formation

Our previous discussion touched upon marl's significance in the formation of Grey Istrian badlands. At this particular stop, a thick layer of marl (approximately 20 m) contributes to the formation of a badland called Princess. At this stop, it is essential to emphasize a hypothesis that, while not yet verified, rests on strong logical foundations: the consistent and thick marly interval within the Globigerina marls, or occurring within the turbidites or megabeds, establishes a geomorphic setting in which the continuity of the badland landscape can be more effectively maintained over both space and time. This stands in contrast to scenarios where alternating layers of marl and sandstones, each measuring only a few decimetres thick, come into play.

The comprehensive explanation of this logical assumption encompasses a series of exogenetic processes and geomorphological situations, thereby exceeding the scope of this guide. However, this assumption can be sim-

Fig. 13. The Princess badland

Fig. 14. Megabed near Paz showing a distinct uneven and erosional lower bedding plane. The megabed has a thickness of up to 10 meters.

plified by elaborating on just one, which is also the most crucial, denudation process: erosion.

Specifically, thick layers of marl, acting as a homogeneous rock mass, support the development of weathering zones with boundaries approximately parallel to the exposed surface of the rock. The establishment of such regular weathering zones facilitates the onset of sheet erosion, characterized by the even removal of surface material (GOVERS, 2006), which is also the weakest material. This type of erosion persists over engineering time, as the characteristics of the eroded surface do not significantly change (Fig. 15). Put simply, erosion on such surfaces continues until the geometrical characteristics of the marl layer significantly change (for example until the layer tilts). To illustrate the intensity of the erosional process on similar surfaces, it can be emphasized that vertical denudation, quantified through remote sensing methods on the test plot within the Princess badland, reaches up to 1 cm per year, as indicated by preliminary findings.

Fig. 15. The marl surface that supports sheet erosion

Fig. 16. Terrains composed of thin layers of marl and sandstone with pronounced surface roughness

On the other hand, erosion occurring on surfaces composed of rhythmic alternations of thinner layers of marl and sandstone can be associated with the erosion type known as rill erosion (FAVIS-MORTLOCK, 2006). Through the erosion of the marl and sandstone rock, plate-like fragments of sandstone remain on the surface. Due to the pronounced roughness of such a surface, the development of sheet erosion is not possible, and the aforementioned sandstone fragments encourage the development of rill erosion. Although this type of erosion can cause the erosion of larger volumes of soil, it is not feasible here, as the sandstone fragments eventually become natural erosion protection. In this way, the gentler slope formed by slightly thinner layers of marl and sandstone quickly “seals off” the fresh rock mass (Fig. 16), effectively preventing the development of prolonged, sometimes excessive erosion, i.e., the development of badlands.

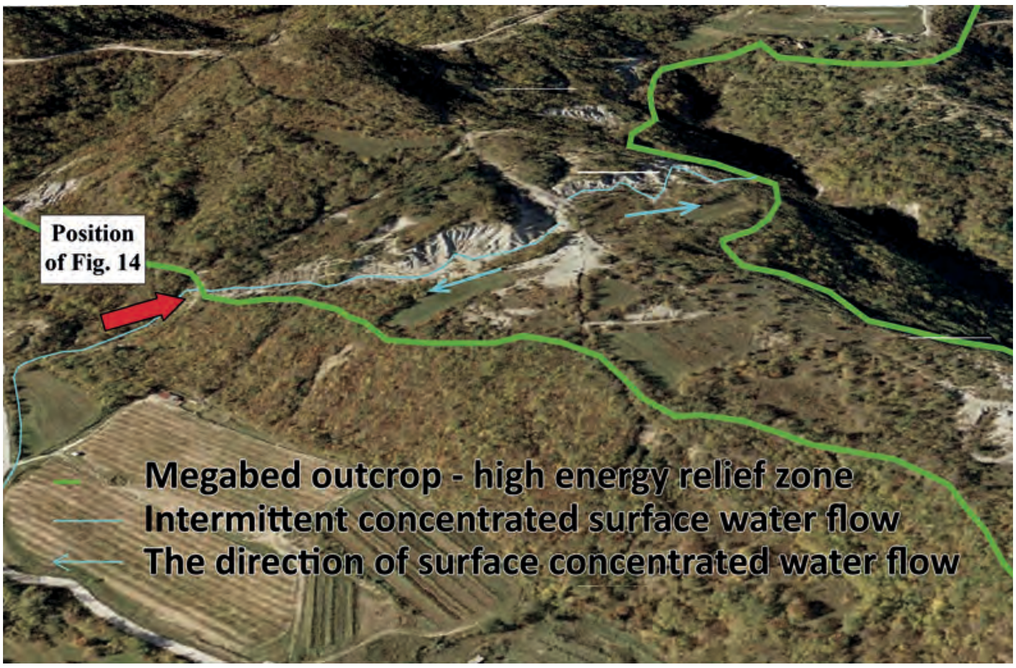

Fig. 17. High-energy relief zone marked by the outcrop of the mega-bed

Situation 2 – Caprock: Nature's Energy Staircase Shaping High-Energy Relief Zones

The significant contrast in the physical-mechanical characteristics of the megabeds and the surrounding rock mass (alternating thin beds of marl and sandstones) leads to the formation of zones with high energy differences. These zones typically mark the boundary between differently inclined terrains; the higher terrain is usually gently sloped, while the lower terrain is steeper (Fig. 17).

When such a zone intersects with the surface-concentrated water flow, it leads to the formation of an excessive erosion point, resulting in the detachment of substantial blocks of megabed material, reaching volumes of several cubic meters as indicated in Fig. 14 and Fig. 18.

Fig. 18. High-energy relief zone marked by the high-volume blocks of megabeds

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FIELD TRIP A2
RED ISTRIA

FIELD TRIP A2 – RED ISTRIA
WESTERN ISTRIAN ANTICLINE AS AN IDEAL NATURAL LABORATORY FOR THE STUDY OF THE REGIONAL UNCONFORMITIES IN CARBONATE ROCKS

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INTRODUCTION

Erosional unconformities typically represent long periods of subaerial erosion associated with loss of rock records, vadose diagenetic and/or pedogenetic alteration, and the formation of surface and subsurface karst (ALONSO-ZARZA & WRIGHT, 2010). Depending on their origin (e.g., tectonically controlled uplift and/or associated relative sea-level fall), climate, and duration of subaerial exposure, unconformities in shallow-water carbonate successions can be associated with different phenomena and/or materials. Due to the formation of surface and subsurface karst, unconformities may be characterised by various karst features (e.g., cavities, dolines, sinkholes, caves, canyons). In addition to various deposits (e.g., laminar rootcrete, breccias, carbonate lithoclasts), unconformities in shallow-water carbonate successions can be accompanied by various palaeosols, soils, soil-derived sediments, and/or pedo-sedimentary complexes. According to MINDSZENTY (2004), these materials may be associated with long-lasting (tectonically controlled) subaerial exposures at major regional unconformities and/or to cyclically organised sequences where marine sedimentation is repeatedly interrupted by shorter or longer episodes of nondeposition and subaerial exposure (controlled by intrinsic or extrinsic factors).

The Adriatic Carbonate Platform (AdCP – see VLAHOVIĆ et al., 2005 and references therein) is one of the best preserved Peri-Mediterranean platforms, located on a stable part of the Adria Microplate characterised by almost continuous carbonate deposition since the Middle Permian. This long-lasting shallow marine carbonate deposition is the result of a combination of (i) stable subsidence, (ii) relative isolation from continental influences, and (iii) palaeogeographic position between the equator and 30° N latitude until the end of the Cretaceous, i.e., within a climatic belt ideal for massive carbonate production. Unlike other parts of the AdCP, the area of the Western Istrian Anticline, the adjacent Pazin Flysch Basin and the Savudrija–Buzet Anticline (which together make up most of Istria, with the exception of the Ćićarija Mt. in its northernmost part and the Učka Mt. along its northeastern and eastern margins) does not belong to the Dinaric mountain belt, but is part of the undeformed foreland of the Adria Microplate (SCHMID et al., 2008; HANDY et al., 2010), i.e. its tectonically much less disturbed part. During the Jurassic and Cretaceous periods, Istria formed the northwestern part of the AdCP and was subsequently covered by Eocene deposits, especially in the northern and northeastern parts, followed by a thin layer of Quaternary deposits throughout the region.

Due to the particular palaeogeographic position of present-day Istria on the northwestern margin of the Adriatic Carbonate Platform, which was characterised by decreased subsidence rate, the Jurassic, Cretaceous, and Palaeogene Istrian successions generally have lower deposition rates, thinner deposits, and more pronounced unconformities compared to contemporaneous deposits in other parts of the AdCP. The particular geotectonic position of the Western Istrian Anticline, which includes most of Istria as an undeformed part of the Adria Microplate, resulted in relatively gentle tectonic deformation and thus very good preservation of original conditions, which allowed almost complete record of unconformities along the limbs of the anticline.

Succession of Istrian deposits can be divided into four megasequences, mostly composed of carbonate deposits (mostly limestones, interbedded with dolomites in some layers; VELIĆ et al., 1995, 2003): 1st Megasequence (lower Bathonian–lower Kimmeridgian); 2nd Megasequence (upper Tithonian–lower/upper Aptian); 3rd Megasequence (lower/upper Albian–upper Santonian); and 4th Megasequence (Eocene) (Fig. 1). The aforementioned megasequences were bound by regional unconformities characterised by significant stratigraphic hiatuses, leading to the formation of terrestrial deposits under different conditions. It is important to emphasise the influence of synsedimentary tectonics, which in the studied area started already in the Early Cretaceous (MATIČEC et al., 1996), resulting in the uplift of the Western Istrian Anticline core and the erosion and/or non-deposition of younger deposits, which significantly reduced the total cover of the underlying deposits.

Well-defined stratigraphic hiatuses of long duration and very gentle post-depositional tectonics allowed the formation of an ideal natural laboratory for the study of major long-lasting unconformities in carbonate rocks. Regional unconformities are characterised by different terrestrial materials (bauxites, palaeosols, terra rossa, pedo-sedimentary complexes, loess-palaeosol sequences) formed in specific palaeoenvironments.

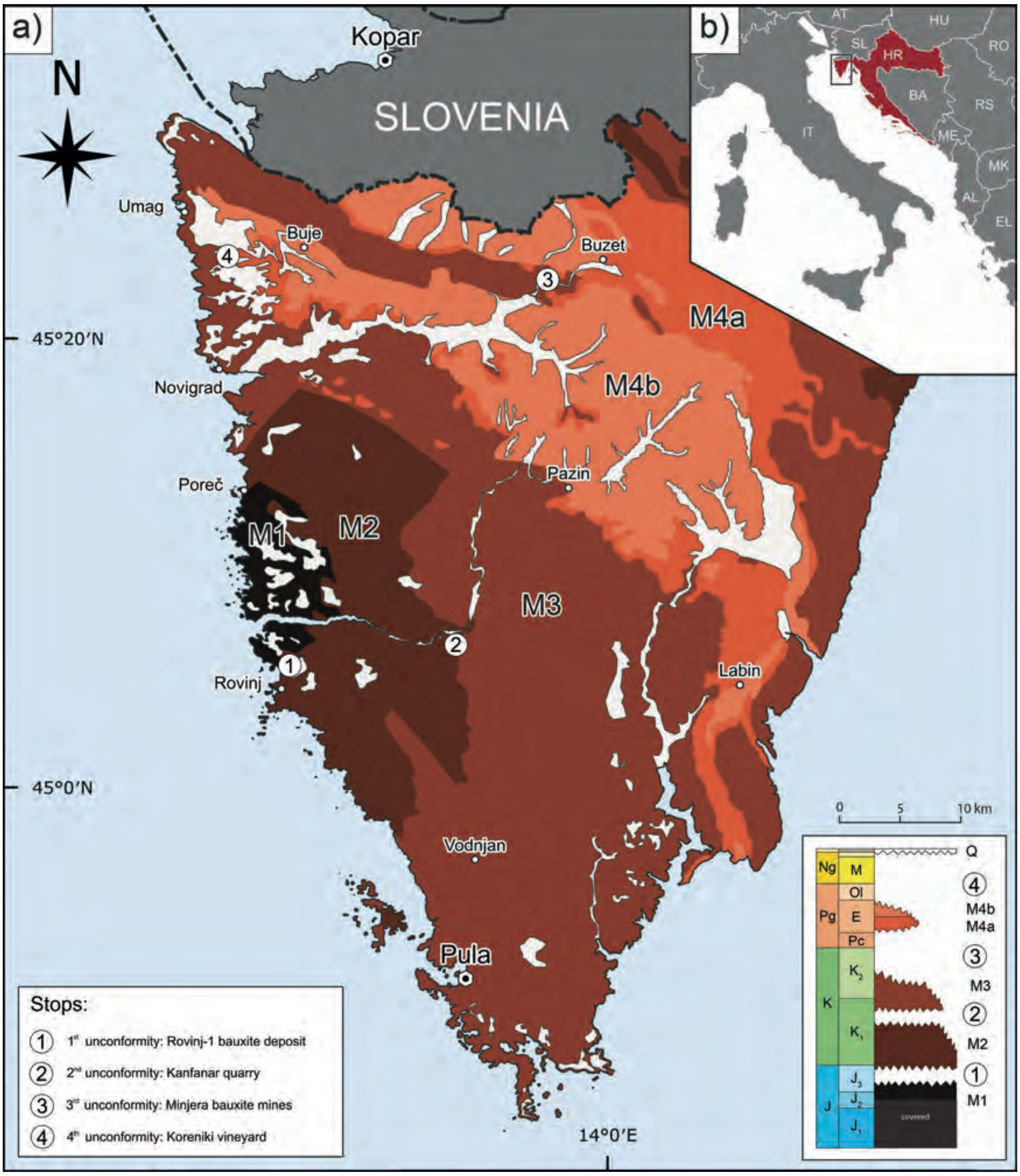

Fig. 1. (a) Map of the Istrian peninsula showing large-scale megasequences separated by regional unconformities, modified after Velić et al. (1995). Legend: M1 – 1st Megasequence (lower Bathonian–lower Kimmeridgian); M2 – 2nd Megasequence (upper Tithonian–lower/upper Aptian); M3 – 3rd Megasequence (lower/upper Albian–upper Santonian); M4a – Carbonate deposits of the 4th Megasequence (lower– middle Eocene); M4b – Clastic deposits of the 4th Megasequence (middle–upper Eocene); Q – Quaternary deposits. The excursion stops are marked with numbers 1 to 4. (b) Location map of Istria.

(A) Lowermost Kimmeridgian–upper Tithonian unconformity was of a relatively long duration (at least 6 Ma – VELIĆ et al., 2003) and resulted in a strongly differentiated relief. This unconformity is accompanied by bauxites. Bauxites are humid-tropical weathering products, similar to ferrallitic soils, which require long-lasting (> 1 Ma) subaerial exposure to form. The occurrence of bauxites is therefore generally considered to indicate a long-lasting exposure and a hot, humid climate. RETALLACK (2010) pointed out that the abundance distribution of bauxites over geologic time shows a pronounced positive correlation with "greenhouse periods" in Earth history. According to D'ARGENIO & MINDSZENTY (1995), subaerial exposure favouring bauxitization on carbonate successions (karst bauxites) is almost always the result of tectonically controlled uplift and associated relative sea-level fall (which may or may not coincide with a lower-order eustatic event). Using studies of Cretaceous karst bauxites, MINDSZENTY (2016) showed that bauxite peaks coincide with globally high temperatures, concurrent eustatic sea-level highs, positive anomalies of global magmatic activity, and abundant oceanic anoxia. Although the likely palaeogeographic position of Istria during the Late Jurassic was within the intertropical belt (after CHANNELL, 1996; STAMPFLI & MOSAR, 1999), the mechanism that led to the sufficient duration of exposure deserves attention because it apparently counteracted the uniform thermal subsidence that is considered characteristic of most Jurassic peri-Adriatic carbonate platforms (e.g., BERNOULLI, 2001). It is likely that the main cause of such an event was large-scale compressional deformation caused by the obduction of ophiolites along the northeastern AdCP margin (SCHMID et al., 2008). The uppermost part of the bauxite is highly altered: its colour is greenish-grey to yellowish-white with vertical to subvertical extensions penetrating the underlying deep red bauxite (DURN et al., 2003). The nature of Fe mineral alteration is clearly redox-based and is closely related to the environmental changes that accompanied overburden deposition.

(B) Upper Aptian–upper Albian unconformity in Istria was of variable duration (11–19 Ma; VLAHOVIĆ et al., 2005) and is accompanied by greenish-grey clay occurring mainly in palaeokarst pits and coarse brecciated regolith (DURN et al., 2003). The thickness of the clays associated with this unconformity ranges from a few centimetres to 1 metre. Transition zones between the shallow-water carbonates and emerged parts of the platform are characterised by clay and marl deposition or by the formation of extensive coastal marshes with reductive conditions and deposition of black sediments (black pebbles). A weakly developed soil structure, the presence of root remains, burrows and channels, now filled mainly with pyrite framboids, nests containing faecal products of soil-dwelling fauna, and nodular pedofeatures indicate that they have been pedogenetically altered. Therefore, the colour of the palaeosols, the presence of root remains only in the upper part of the profile, and the large amount of pyrite framboids may indicate that they were probably seasonally marshy soils or permanently waterlogged soils (DURN et al., 2003).

(C) Upper Cenomanian/upper Santonian–lower Eocene unconformity was of a very long duration (from about 25 Ma in southern Istria and the Učka Mt. to 40 Ma in northern Istria). However, in western Istria, erosional remains of middle Eocene foraminiferal limestones were found at several sites overlying Lower Cretaceous limestones, indicating a possible stratigraphic hiatus of up to about 87 Ma (MATIČEC et al., 1996). During such a long hiatus, part of the succession could have been chemically and physically weathered, so the above duration of the hiatus is probably somewhat overestimated. Although this stratigraphic hiatus was extremely long, relatively thin deposits, mostly bauxite, formed in the lowermost parts of the palaeorelief. Their occurrence coincides with one of the most widespread bauxite events in the Peri-Mediterranean area. Bauxites of Palaeogene age are abundant from Hungary through Slovenia, Croatia, and Bosnia and Herzegovina to Albania and Greece, all belonging to the Apulian Promontory (Adriatic Microplate) palaeogeographic domain (BÁRDOSSY & DERCOURT, 1990). As with the Jurassic bauxites, the cover sequence also normally begins with a palustrine/lacustrine transitional facies. In association with the initial transgression, the introduction of stagnant pore water into the soil-derived sediment occurs at a stage when diffuse porosity throughout the deposit is not yet restricted and the iron oxide phases are still partially mineralized, so that the bauxite may well react with reducing fluids, resulting in large-scale alteration. This is the case with most of the bauxite deposits in the Minjera area. The famous pyritic bauxites, where whole bauxite bodies are thoroughly grey due to finely disseminated eogenetic pyrite, clearly show that in some deposits the above conditions were met when marine pore water came into contact with the unconsolidated bauxite (ŠINKOVEC et al., 1994).

(D) Upper Eocene–Recent unconformity was of a very long duration, even in the areas with the longest deposition – up to the flysch in the latest Eocene (resulting in a stratigraphic hiatus of about 35 Ma). However, on the limbs of the Western Istrian Anticline, the hiatus was longer due to synsedimentary tectonics, as there were probably no younger deposits than the Middle Eocene foraminifera limestones (40–45 Ma), while in the apical part there may have been no carbonate deposition since the Early Cretaceous time, so the duration of the stratigraphic hiatus could be more than 100 Ma. This unconformity is accompanied by different deposits and soils/palaeosols among which terra rossa, loess-palaeosol sequences and pedo-sedimentary complexes are predominant.

Field trip participants will visit four sites with different terrestrial materials associated with the four regional unconformities in the shallow-marine carbonate successions within the Western Istrian Anticline and the Savudrija–Buzet Anticline, serving as indicators of palaeoenvironment, palaeoclimate, and provenance: (1) Rovinj-1 bauxite open pit (the lowermost Kimmeridgian–upper Tithonian regional unconformity) with red, grey and white bauxite; (2) Kanfanar quarry (the upper Aptian–upper Albian regional unconformity) with greenish-grey clay as palaeosol;

(3) Minjera historic bauxite mine (the upper Cenomanian/upper Santonian–lower Eocene regional unconformity) with pyritic bauxite; (4) Koreniki vineyard (the upper Eocene–Recent regional unconformity) with terra rossa soil profile classified as Rhodic Lixisol according to the WRB system. Special attention will be given to the sedimentology, mineralogy, geochemistry and micromorphology of these materials. This excursion was prepared on the basis of investigations carried out within the WIANLab project. Some of the results that we will present during this excursion have been published this year. This refers mainly to site 4, where the results published in the article by DURN et al. (2023) will be presented.

STOP 1 – ROVINJ-1 BAUXITE DEPOSIT: LOWERMOST KIMMERIDGIAN–UPPER TITHONIAN UNCONFORMITY

The 1st stop represents one of the best outcrops of the 1st unconformity. The unconformity between the Oxfordian to lower Kimmeridgian and upper Tithonian carbonate deposits is marked with the formation of a large bauxite body up to 20 metres thick – the Rovinj-1 bauxite deposit (Fig. 2), which is currently the only operating bauxite mine in Croatia. However, the bauxite ore is not used for aluminium production due to its high silica content but is currently used as a secondary ore in the production of mineral wool. With estimated reserves of around 15 million tonnes, it is also one of the largest bauxite deposits in Croatia.

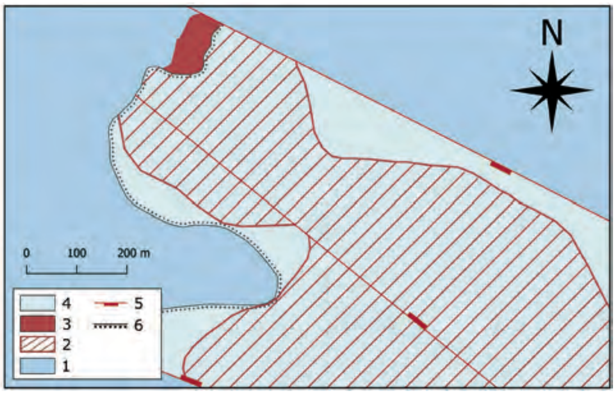

Fig. 2. Geological map of the Rovinj-1 deposit, based on the data provided by the GEO-5 company. Legend: 1 – Muča unit, 2 – extent of the bauxite body below the Kirmenjak unit, 3 –bauxite, 4 – Kirmenjak unit, 5 – normal faults, 6 – unconformity.

Regional and geotectonic setting

Rovinj bauxite represents one example of bauxite formation in the northern part of the Adriatic Carbonate Platform, since many other bauxite occurrences of the same age are documented in this area (TROJANOVIĆ, 1973; ŠINKOVEC, 1974; VELIĆ & TIŠLJAR, 1988). It is also a part of a much more widespread bauxite belt which formed during the Late Jurassic in the Tethyan realm. Karst bauxites of Upper Jurassic age can be found in Spain (MOLINA et al., 1991), Austria (STEINER et al., 2021), Slovenia (DOZET et al., 1993) and all the way to Montenegro (RADUSINOVIĆ et al., 2017; RADUSINOVIĆ & PAPADOPOULOS, 2021) and Greece (LASKOU & ECONOMOU-ELIOPOULOS, 2007; GAMALETSOS et al., 2017). The timing of this regional bauxitisation event coincides with higher temperatures recorded in the Late Jurassic (FRAKES et al., 1992; BRIGAUD et al., 2008; HAQ, 2018), as most of the Jurassic was characterised by coldhouse conditions (FRAKES et al., 1992). Since the sea level during the Late Jurassic was the highest during all of Jurassic (HAQ & AL-QAHTANI, 2005; HAQ, 2018), it is likely that the prolonged subaerial exposure required for bauxite formation was not a consequence of eustatic sea-level changes, but was of a tectonic nature. On the Adriatic Carbonate Platform this uplift was triggered in response to the overburden pressure generated by the oceanic crust during the obduction of the Vardar Ocean ophiolites in the Late Jurassic (SCHMID et al., 2008, 2020; VAN HINSBERGEN et al., 2020), which generated a flexural forebulge in the front of the obduction zone. As such, Rovinj-1 bauxite and accompanying bauxite bodies belong to the 2nd type "bauxites formed in passive plate interior under intraplate stress" sensu D'ARGENIO & MINDSZENTY (1995).

Geological setting

The Rovinj-1 bauxite formed during the early Kimmeridgian to late Tithonian subaerial exposure phase, which separates the first, early Bathonian to early Kimmeridgian megasequence and the second, late Tithonian to early/late Aptian megasequence. More precisely, it is situated between the Oxfordian to lower Kimmeridgian Muča and Lim units of the first megasequence, and lower Tithonian Kirmenjak unit of the second megasequence. The Muča unit mainly comprises ooid grainstones and ooid rudstones deposited in high-energy tidal bar facies, and peloidal to skeletal wackestones deposited in low-energy conditions in the shallow subtidal. The Muča unit appears as lenses in the beds of the Lim unit, which is composed of peloidal packestones deposited in a low-energy lagoonal environment within the shallow subtidal (VELIĆ & TIŠLJAR, 1988, and references therein). The fossil assemblage of both the Muča and Lim units indicates their Oxfordian to lower Kimmeridgian age. The formation of Rovinj breccias (VELIĆ & TIŠLJAR, 1988) followed the deposition of these two units, which have formed during the regression that preceded the subaerial exposure. They show a gradual transition from Lim and Muča units and are composed from fragments of these two units. The subaerial exposure phase that followed is marked with the formation of bauxites (ŠINKOVEC, 1974; VELIĆ & TIŠLJAR, 1988) showing an erosional contact with the Lim and Muča unit, as well as the Rovinj breccias (VELIĆ & TIŠLJAR, 1988). Besides the formation of bauxites, the unconformity is also recorded as simple erosional gaps between the Muča and Lim units and the Kirmenjak unit on some localities (VLAHOVIĆ et al., 2003) as well as the deposition of palaeosols (VELIĆ & TIŠLJAR, 1988; VLAHOVIĆ et al., 2003). The end of the subaerial exposure phase is followed by the oscillating transgression, after which the deposition of the Kirmenjak unit had begun. This unit is composed of the cyclical alternation of mudstones, mudstones with indications of subaerial exposure and finally the lenses or intercalations of black pebble breccias (TIŠLJAR, 1986; VELIĆ & TIŠLJAR, 1988). In the Rovinj-1 deposit, the immediate cover consists of a cyclic alternation of clays, brackish to marine limestones and black pebble breccias, which gradually alternate in the lower part of the Kirmenjak unit.

Provenance of the Rovinj-1 bauxite

The source material from which the Rovinj bauxite was formed, was derived from several sources, mainly from the wind-blown material and the insoluble residue, as proposed by ŠINKOVEC (1974). Volcanic material likely represented the majority of the aeolian input, which is also evident from the presence of zircon and apatite grains in the bauxite (ŠINKOVEC, 1974). The presence of multiple levels of tuffs and bentonites of Kimmeridgian age in the Trento plateau, Northeastern Italy (PELLENARD et al., 2013) and of Kimmeridgian–Tithonian age in Gorski Kotar area, central Croatia (ŠĆAVNIČAR & NIKLER, 1976; VELIĆ et al., 2002), is indicative of deposition of such materials on the area of the Adriatic Carbonate Platform during the duration of the subaerial exposure in Istria. Limestones of the Muča unit which represent the bedrock of Rovinj-1 bauxite deposit contain a relatively high content of insoluble residue (2.19%; DURN et al., 1999). During the karstification of these limestones, this material was also accumulated in karstic depressions, and subjected to bauxitisation together with wind-blown volcanic material.

Petrology, mineralogy and geochemistry of the Rovinj-1 bauxite

In the Rovinj-1 deposit, the bauxite consists mainly of red bauxite (Fig. 3a–d), while white and grey bauxite occur in the uppermost section of the deposit (Fig. 3b, c).

Red bauxite consists mainly of boehmite, haematite and anatase, with minor amounts of chlorite, rutile, and locally gibbsite (Fig. 4a). The bauxite shows a uniform mineralogical composition throughout the profile (Fig. 4a), which is in accordance with major oxide data (Fig. 4b) as they are also uniform, and do not show any significant trends throughout the profile. Grey bauxite formed during the initial flooding of the bauxite, which led to the formation of a swampy environment on top of the bauxite. The microbial activity in the swamp led to the depletion of oxygen and production of hydrogen sulphide, which in turn caused the solubilisation of iron oxides and pyritization of the topmost section of the underlying bauxite. Precipitated

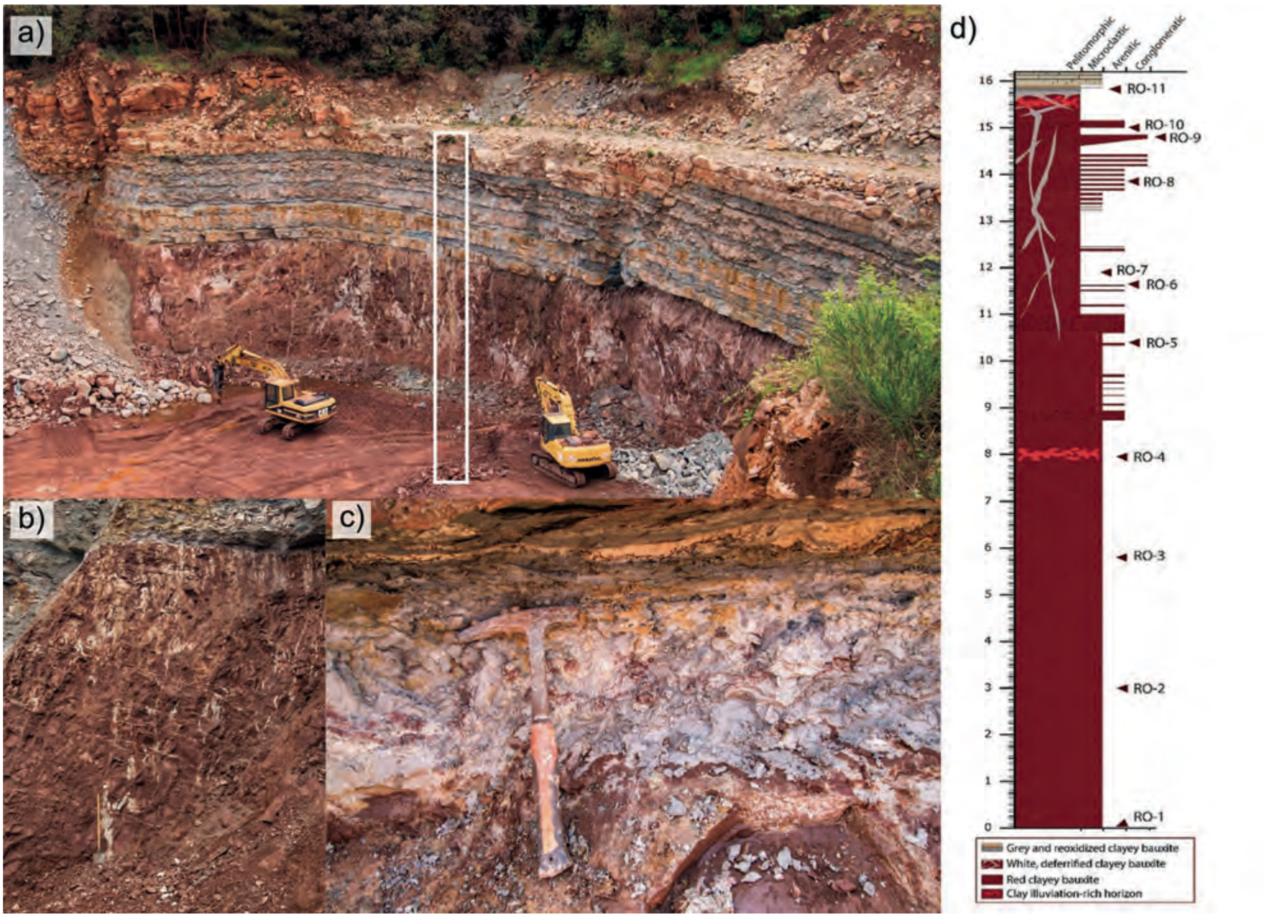

Fig. 3. Field photographs from the Rovinj-1 deposit and the analysed bauxite profile. (a) Rovinj-1 deposit with indicated position of the analysed profile. (b) Red bauxite cross-cut with veins of white bauxite. (c) Topmost part of the bauxite, with visible grey bauxite, pyritised roots and kaolinitic marl (white intercalations in the clay at the top of the photograph) and red bauxite with occurrences of white bauxite. (d) Reconstruction of the sampled bauxite profile (RO – sampling site).

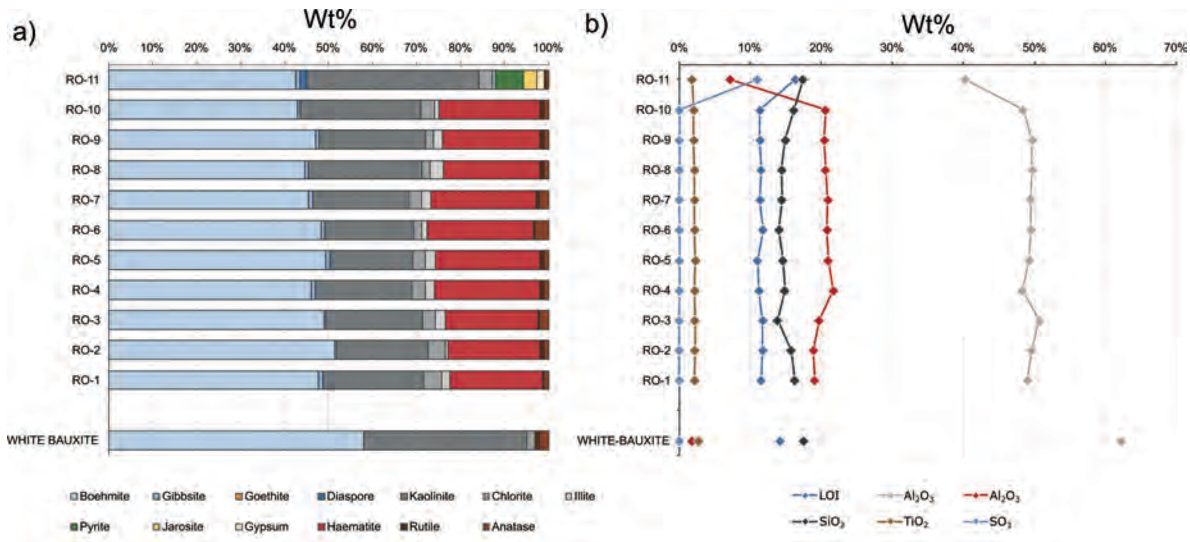

Fig. 4. Plots displaying mineralogical and chemical composition of the analysed bauxite. (a) Bulk mineralogical composition throughout the Rovinj-1 bauxite deposit. (b) Distribution of major oxides along the studied bauxite profile.

iron sulphides are present as framboids and crystals dispersed in the bauxite matrix and in the in situ pyritised root remains. As such, the grey bauxite has a similar composition as the red bauxite, but with the substitution of haematite with iron sulphides, coupled with different sulphate minerals which formed during pyrite decomposition. The major oxide content of the grey bauxite is in accordance with mineralogical content of the bauxite (Fig. 4a, b), as the presence of pyrite and sulphate minerals is also confirmed with elevated SO_3 values (Fig. 4b). White bauxite appears directly under the grey bauxite, as a metre-thick zone crosscut by the veins of the white bauxite (Fig. 3c). On closer inspection, these white bauxite veins reveal roots that have penetrated the upper part of the bauxite, which is also evident from the sporadic presence of pyritised root remains within the white bauxite veins (Fig. 3c). White bauxite is also present below this zone, extending along fractures and faults up to ten metres into the red bauxite (Fig. 3b). The white bauxite in this zone probably originated from reducing and acidic pore water that seeped into the bauxite along the fractures. Since the white bauxite is simply the deferrified red bauxite, it has the same composition but does not contain iron minerals (Fig. 4a). The concentrations of major oxides (Fig. 4b) in the white bauxite agree well with the mineralogical content (Fig. 4a), as the values for Al_2O_3, SiO_2 and TiO_2 are elevated compared to the red bauxite, while Fe_2O_3 is almost zero.

The bauxite is structurally heterogeneous, as parts of the analysed bauxite profile are mainly pelitomorphic to microclastic (Fig. 5a), with intercalations of arenitic (Fig. 5b) and conglomeratic bauxite (Fig. 5c). On the studied profile, a coarsening upward trend can be observed, as the conglomeratic and arenitic bauxite are progressively more abundant in the upper part of the profile (Fig. 3d). This may be associated with the progressive aridification of the climate in the middle Tithonian (WIGNALL & RUFFELL, 1990; RUFFELL & RAWSON, 1994; BRIGAUD et al., 2008; HESSELBO et al., 2009) as this probably led to a decrease in vegetation cover over the deposit, allowing greater erosion of the bauxite during the formation of its uppermost part.

The bauxite cover sequence

The cover sequence of the Rovinj-1 bauxite begins with the deposition of grey kaolinitic clay (Fig. 3c), which is then covered with a cyclically alternating sequence of clays, marls, limestones and black pebble breccias (Fig. 6a, b).

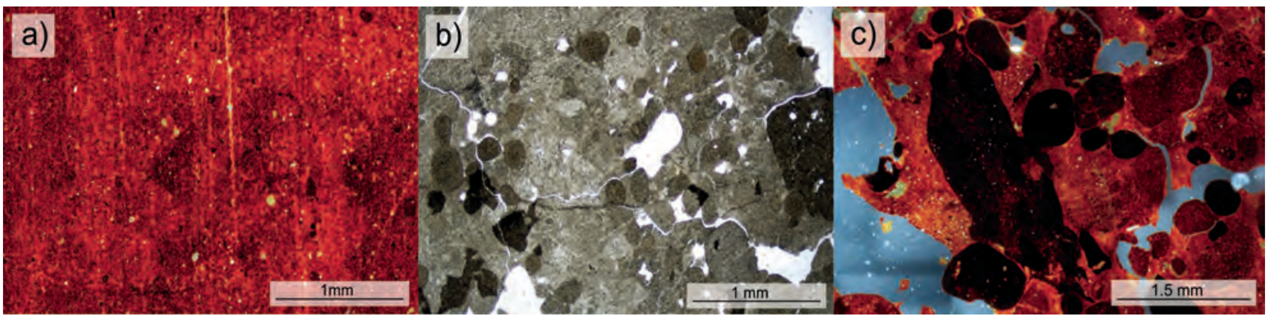

Fig. 5. Photomicrographs of different bauxite samples. (a) Pelitomorphic to microclastic red bauxite sample (RO-6), taken under conoscopic illumination. (b) White bauxite sample (RO-45) with visible arenitic structure, PPL. (c) Red bauxite sample (RO-9) with visible conglomeratic structure and bauxite clasts enriched in iron oxides, taken under conoscopic illumination.

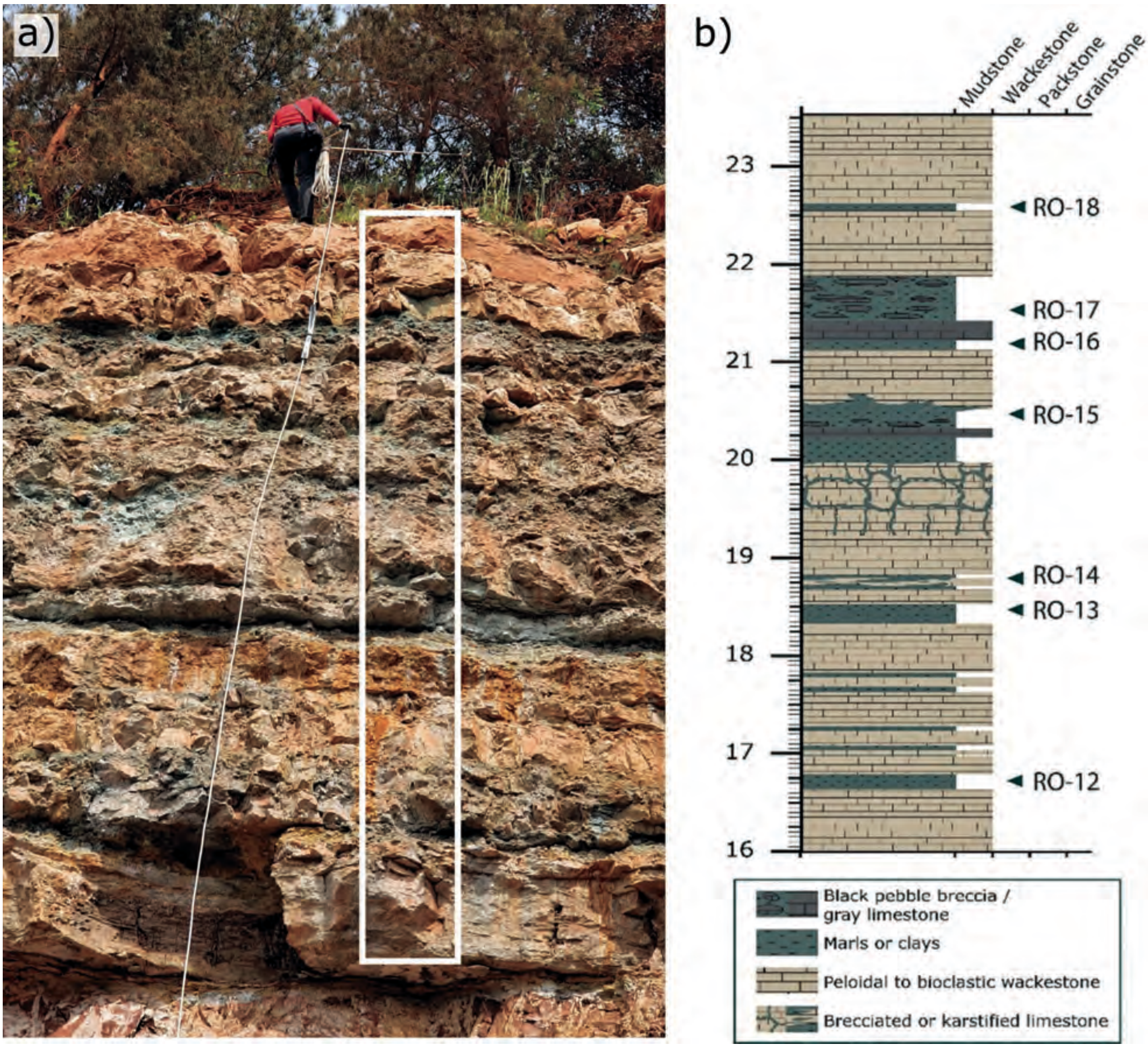

Fig. 6. Studied profile of the bauxite cover. (a) Field photograph of the profile, with the studied section indicated with the white rectangle. (b) Reconstruction of the studied cover section. RO – sampling site.

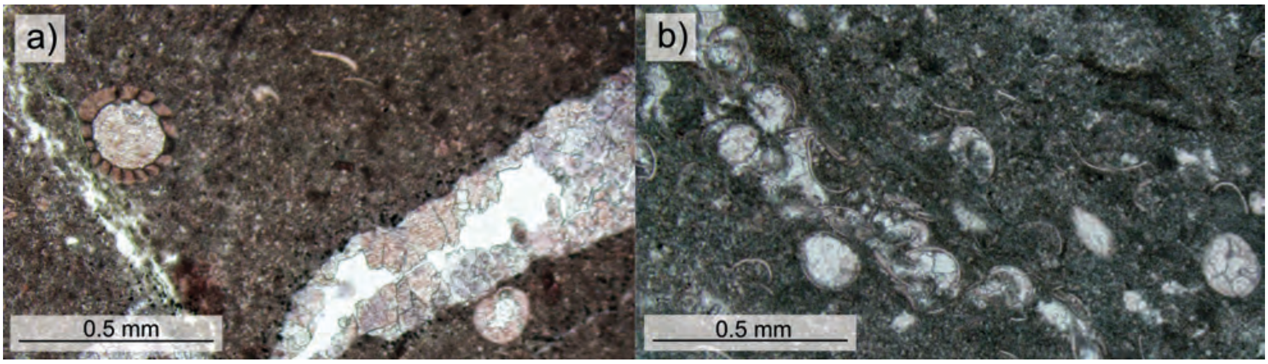

Fig. 7. Photomicrographs from the limestones in the cover sequence. (a) Charophyta oogonium, PPL. (b) Accumulation of ostracod shells, PPL.

The limestones from the cover sequence formed in a restricted environment with a very scarce fossil record, consisting exclusively of *Charophyta* oogoniums (Fig. 7a) and ostracods (Fig. 7b). Based on the aforementioned faunal and floral record, these limestones formed in conditions of fluctuating salinity (schizohaline), before they were covered with fully marine limestones of the Kirmenjak unit.

Such variations in salinity and the very restricted fauna dominated by ostracods indicate that these limestones were deposited in an isolated body of water that probably developed over the bauxite and other karst depressions

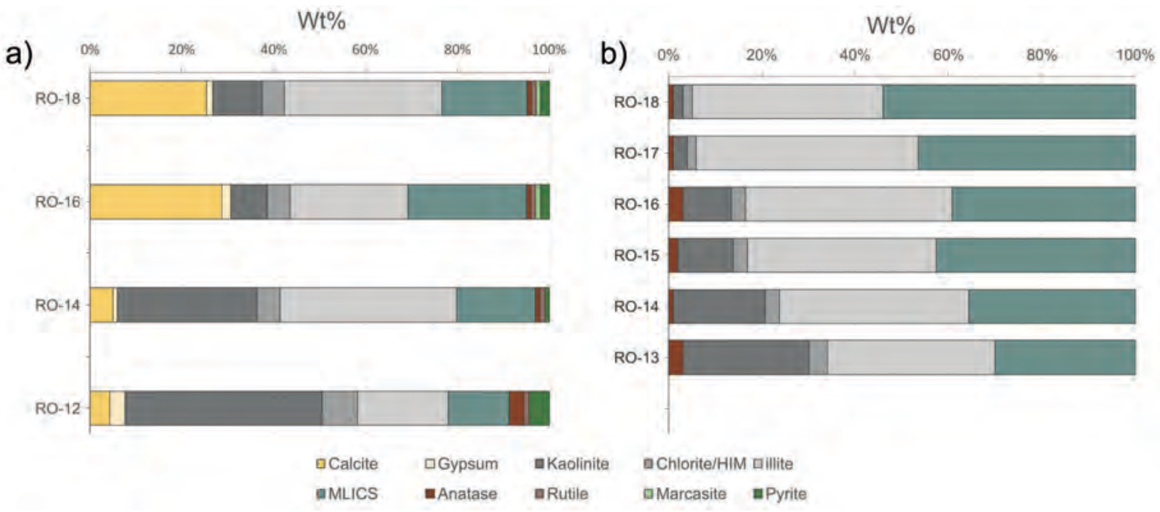

Fig. 8. Mineralogical composition of the studied clay/marl samples throughout the studied section of the Rovinj-1 bauxite deposit. (a) Bulk mineralogical composition throughout the analysed profile. (b) Mineralogical composition of the clay fraction throughout the analysed profile.

during the rise of the water table before the emerged terrain was completely flooded and the fully marine Kirmenjak unit was deposited. Such a deposition sequence is commonly referred to as a "blue hole sequence" according to RASMUSSEN & NEUMANN (1988), and is indicative of internal transgression.

Mineralogy of the cover sequence

These materials consist mainly of clay minerals (kaolinite, chlorite/hydroxyl interlayered mineral (HIM), illite, mixed layered illite–chlorite–smectite (MLICS)), calcite, and gypsum, with minor amounts of titanium oxides (rutile and anatase) and iron sulphides (pyrite and marcasite, Fig. 8a). The clay fraction has a similar composition to the bulk samples and consists of MLICS, chlorite/HIM, illite, kaolinite and minor amount of anatase (Fig. 8b). There is a gradual increase in MLICS and illite and a gradual decrease in kaolinite throughout the studied profile, best visible in the composition of the clay fraction.

This trend probably reflects aridification at the end of the Tithonian, which is also recorded in the clay record (WIGNALL & RUFFELL, 1990; HESSELBO et al., 2009) and by stable isotope data from oyster shells (BRIGAUD et al., 2008). Because the areas around the "blue hole" above the bauxite were still emerged during the initial stages of transgression, pedogenesis was not yet complete and the newly formed soils may have responded to climate change. These newly formed soils, rich in illite and mixed layered clays, gradually replaced the previously formed ferralitic soils, as indicated by the changes in clay mineralogy in the studied profile.

STOP 2 – KANFANAR QUARRY: UPPER APTIAN–UPPER ALBIAN UNCONFORMITY

The upper Aptian–upper Albian unconformity marks the division between the second, late Tithonian–early/late Aptian megasequence, characterised by facies heterogeneity and considerable thickness, and the third, late Albian–late Santonian megasequence, which is more than 1000 m thick with different facies successions (Fig. 1). Deposition of the second megasequence started at the end of the Tithonian with an oscillating transgression resulting mainly in peritidal shallowing-upward cycles. The megasequence is capped by the upper Aptian deposits, which experienced a relatively rapid shallowing, resulting in subaerial exposure (VELIĆ et al., 1989, 2003). This regional event was a consequence of a relative sea-level fall caused by the interaction of eustatic changes and synsedimentary tectonics in the Istrian part of the Adriatic Carbonate Platform (AdCP), resulting in variable duration of shallow-water conditions in different parts of the platform, as well as in varying intensity of erosion of Aptian and Barremian deposits (VELIĆ et al., 1989, 2003). The end of this megasequence was marked by deposition of breccias and conglomerates with common blackened pebbles, representing swamp conditions, interbedded with clay layers.

Terrestrial deposits associated with this regional unconformity can be found at various places (Fig. 9) such as Tri jezerca quarry, Selina 4 quarry, Balle quarry, Kanfanar quarry, Lakovići quarry, Goda quarry as well as at the road cut of 'Istrian Y' highway (OTTNER, 1999; DURN et al., 2003, 2006; MILEUSNIĆ, 2007). The duration of the stratigraphic hiatus varied between 11 and 19 million years, depending on the palaeogeographical position of individual localities (VELIĆ et al., 1989). The evidence of exposure is characterised by greenish-grey clay, found mainly in paleokarst pits, varying in thickness from a few centimetres to a meter, and coarse brecciated regolith.

Following an extensive subaerial exposure during the late Aptian and early Albian, a gradual transgression took place during the middle Albian, which eventually transitioned into a complete transgression. It marks the beginning of the third megasequence, that comprises several lithostratigraphic units (VELIĆ et al., 1995), including records of a peritidal and foreshore sedimentary system

Fig. 9. Outcrops of terrestrial deposits associated with second regional unconformity. (a) Selina quarry; (b) Kanfanar quarry; (c) Lakovići quarry; (d) Goda quarry; (e) Tri jezerca quarry; (f) Road cut of 'Istrian Y' highway near Kanfanar.

in the middle and late Albian, differentiation of sedimentary systems during the latest Albian and Cenomanian, drowned platform systems during the latest Cenomanian and Turonian (JENKYNS, 1991; GUŠIĆ & JELASKA, 1993; VLAHOVIĆ et al., 2005), and re-establishment of the shallow-water sedimentary system during the late Turonian, Coniacian and Santonian (VELIĆ et al., 2003; VLAHOVIĆ et al., 2005).

Location

For presentation of the upper Aptian–upper Albian unconformity, the studied geological profile located at the southeastern part of the Kanfanar quarry (Fig. 9c) is selected.

Kanfanar quarry is located 2 km west of Kanfanar on both sides of the road connecting Kanfanar and Rovinj. It is renowned for containing the high-quality natural stone, a light brown lower Aptian oncoid limestone, the most famous architectural and building stone from Istria, known on the market under the names *Kanfanar*, *Bale* (*Valle*), *Rose Karst* and under the general name *Istrian yellow* (*Giallo d'Istria*). The first notes on the exploitation of this stone date back to the 15th century, when it was transported from Brijuni archipelago (island of St. Jerolim) to Ancona on the order of the famous sculptor and architect Juraj Dalmatinac. The Kanfanar quarry was opened for the needs of the Austro-Hungarian railways. Since 1961, the quarry has been part of the company Kamen d. d. Pazin, and it is still active today. The extraction of the stone from the quarry employs both surface and underground methods. Some of the most famous architectural structures made of this stone are interior of the Austrian Parliament building in Vienna (Austria), Park at Krasnodar Stadium (Russia), Europa-park Colosseo in Freiburg (Germany) and Szent Istvan Szobor square in Budapest (Hungary).

Lithology, sedimentology and micropaleontology of carbonates

A total thickness of 20.85 m (9.60 m below and 11.25 m above the unconformity; Fig. 10) of predominantly shallow-water platform carbonates comprising the second unconformity was examined to: (1) reconstruct environmental changes around the unconformity in more details, (2) refine/define lithostratigraphic subdivision of the Aptian and Albian deposits and (3) define extensive subaerial exposure phase – regional unconformity during the late Aptian and early Albian. The focus was on the shallow-water carbonates deposited in inner-platform peritidal environments overlain by carbonate breccia with clayey calcareous matrix and greenish-grey clay, both marking a regional unconformity, and on the succession on its top (Fig. 10). The lithology, sedimentology and microfossil content of the studied section were analysed and described on site at cm-scale.

Fig. 10. Kanfanar quarry. (a) Kanfanar section below unconformity level; (b) and (c) Section including regional unconformity and succession above it.

A total of 71 samples were collected for sedimentary (petrographic) and micropalaeontological analyses. Benthic foraminifera together with other biogenic and lithologic constituents are used to reconstruct sedimentary environments and conditions that existed during depositional history of the investigated Lower Cretaceous succession in western Istria (CVETKO TEŠOVIĆ et al., 2011).

Based on lithological features and micro- and macrofossil assemblages, seven lithofacies types (LF1–LF7) were determined (Fig. 11, Tab. 1).

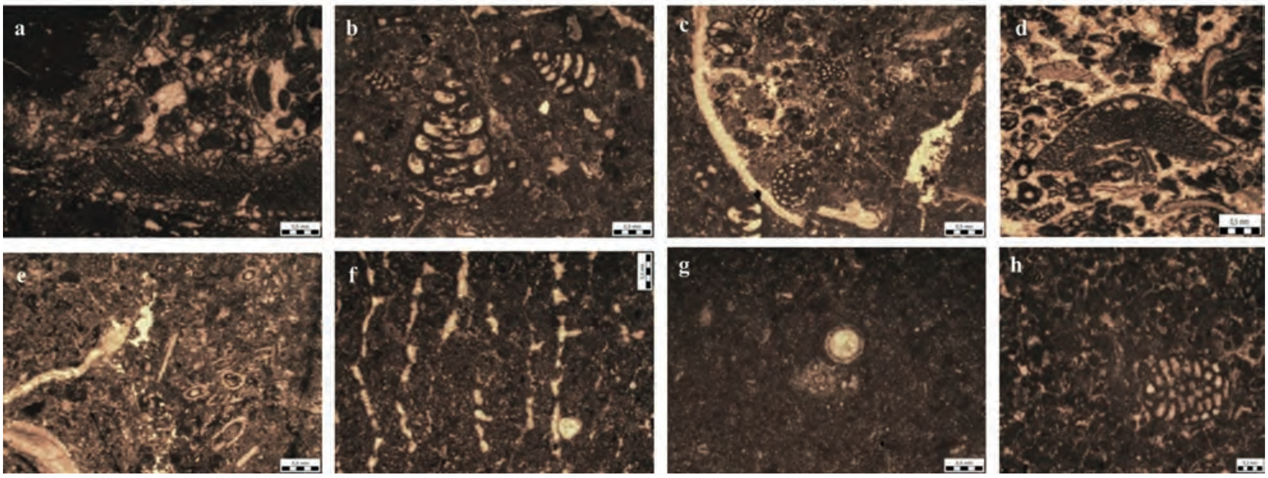

Fig. 11. (a) Bacinella wackestone–floatstone: *Palorbitolina lenticularis* (BLUMENBACH), *Bacinella irregularis* RADOIČIĆ (LF1, KA-1). (b) Foraminiferal-peloidal wackestone–packstone: *Praechrysalidina infracretacea* LUPERTO SINNI (LF2, KA-8). (c) Foraminiferal-peloidal packstone: *Voloshinoides murgensis* LUPERTO SINNI and MASSE, *Archaealveolina reicheli* (DE CASTRO), (LF2, KA-8). (d) Peloidal-bioclastic packstone–grainstone: *Palorbitolina lenticularis* (BLUMENBACH) (LF3, KA-24). (e) *Decastronema*-algal packstone: *Salpingoporella dinarica* RADOIČIĆ (LF4, KA-28;). (f) Fenestral-peloidal wackestone–packstone (LF5, KAK-10A). (g) Characean-ostracod wackestone (LF6, KAK-24;). (h) Peloidal-miliolidal packstone: "*Valdanchella*" *dercourti* DECROUEZ and MOULADE (LF7, KAK-27).

Tab. 1. Kanfanar lithofacies types (LF) with texture, sedimentary structures, skeletal and not-skeletal particles, and depositional environment.

Kanfanar lithofacies (LF)		Texture, sedimentary structures, skeletal and non-skeletal particles	Depositional environment
LF1	**Micritic limestones with *Bacinella* oncoids and low diversity biota**	Wackestone–floatstone with *Bacinella* oncolites and common bioturbated features. Benthic foraminifera (common **orbitolinids** – *Palorbitolina lenticularis*), **bioclasts** of bivalve fragments (mostly requiniid rudists –shells of *Toucasia* sp.)	Lagoons and low-energy shallows with a very low sedimentation rate, common bioturbated levels, and low biotic diversity
LF2	**Finer-grained micritic limestones**	Mostly wackestone–packstone with intraclasts, bioclasts, peloids, benthic foraminifera and bivalve shells.	Subtidal environment of elevated water energy to moderate energy
LF3	**Grainy limestones with diverse biota**	Foraminiferal-peloidal packstone–grainstone with more diverse biota: common benthic foraminifera (*Palorbitolina lenticularis*), decastronemas, bivalve and echinoderm fragments and occasionally dasycladal algae (predominate *Salpingoporella dinarica*).	Deepening of the sedimentary environment and deposition in somewhat deeper subtidal environments characterised by higher biotic diversity
LF4	***Decastronema*-algal limestones**	Wackestone and wackestone–packstone. Abundant dasycladal algae – *Salpingoprella dinarica, rare Decastronema*? and benthic foraminifera and thaumatoporellaceans. Less common fragments of rudists and other bivalve shells. Common small peloids.	Restricted to open shallow subtidal carbonate platform with moderate water energy
LF5	**Fenestral micritic limestones occasionaly laminated with very low diversity biota**	Mudstone–wackestone, locally packstone, with peloids, ostracods shells, algal filaments, pedoturbations, desiccation cracks and circumgranular cracks. Rare *Decastronema*?, thaumatoporellaceans, nubeculariids and sporadically common miliolids.	Restricted and protected shallow subtidal and intertidal inner carbonate platform with low water energy
LF6	**Micrites with ostracods and characean remains**	Wackestone with numerous ostracod shells, rare characean oogonia and algal filaments, gastropods and thaumatoporellaceans. Locally laminated.	Shallow subtidal carbonate platform with low water energy (with occasional fresh water influence)
LF7	**Fine-grained limestones with abundant miliolids**	Wackestone–packstone and packstone–grainstone with abundant miliolid and other benthic foraminifera, micritized gastropod fragments. Less common intraclasts, peloids and rare aggregated grains.	Shallow subtidal carbonate platform with moderate water energy

Mineralogy and micromorphology of greenish-grey clay

In the Kanfanar area, the subaerial exposure phase which commenced during the late Aptian left clay deposit up to 40 cm thick. Within the transgressive upper Albian deposits which overlain clay deposit, features suggesting multiple brief subaerial exposures events are recorded, primarily manifested through coarse brecciated zones filled with greenish-grey and greenish-yellow clay.

Four samples of the 40-cm-thick greenish-grey clay (5GY6/1 to 5GY7/1 after Munsell) in the central part of the quarry outcrop were collected along the profile (Fig. 12) for various analyses, including mineralogical, chemical and micromorphological analyses. Samples of carbonates situated immediately below the clay were also collected and analysed.

The greenish-grey clay does not exhibit significant variation in colour, microstructure and fabric along the profile. The samples are grey with areas of yellowish and greenish hues, and the microstructure is predominantly granular, angular blocky, and planar (Fig. 13a, b, c). The yellowish patches are partially opaque and likely composed of iron oxides (Fig. 13e). Pyrite sporadically appears in aggregates of typical nodules (Fig. 13g) and framboids.

The birefringent fabric of the micromass (b-fabric) also shows little variation, with random streaking, stipple-speckled b-fabric (Fig. 13d, f), cross-streaking, and

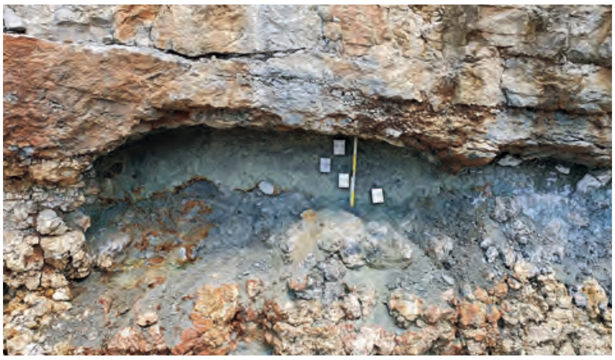

Fig. 12. Profile of greenish-gray clay in the Kanfanar quarry with Kubiena boxes for micromorphological analyses

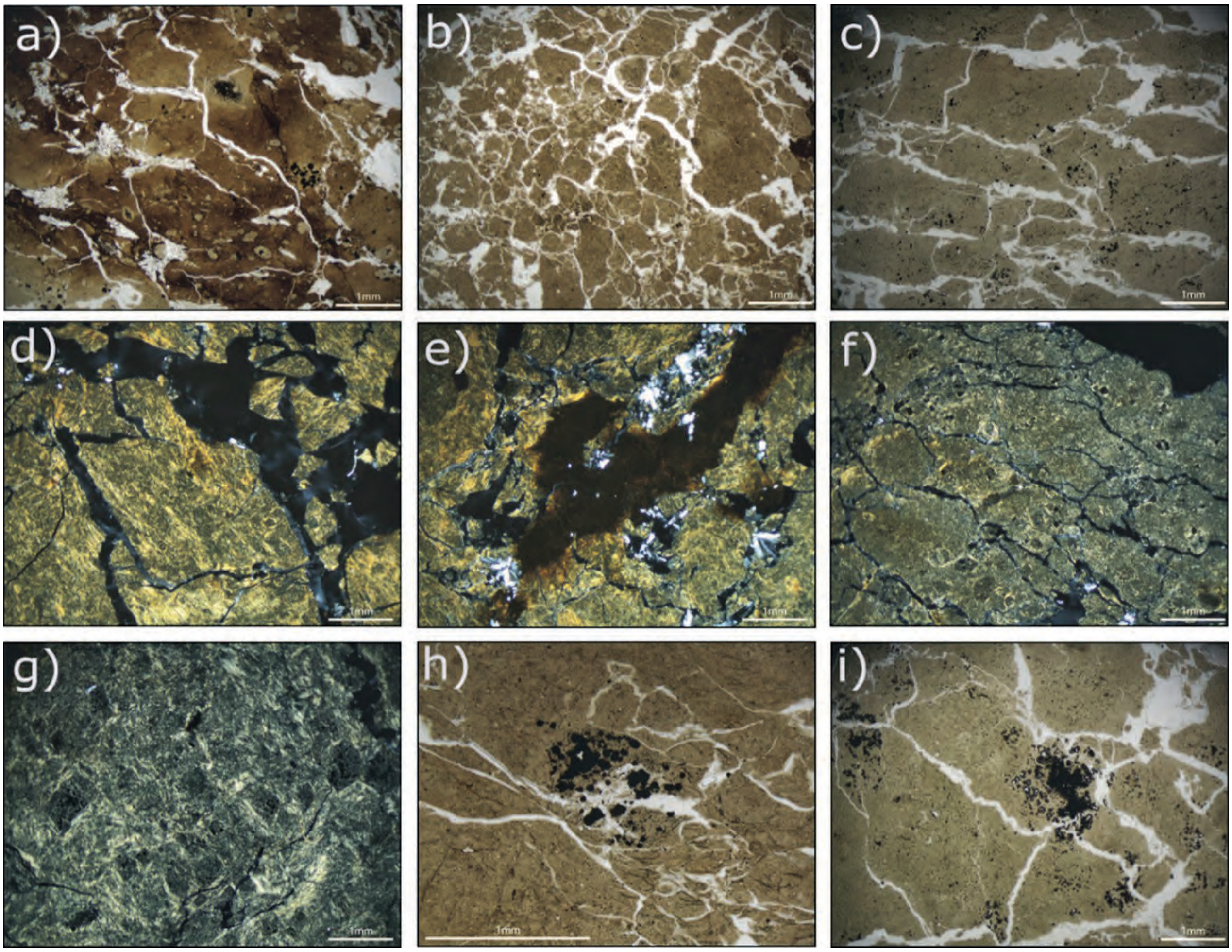

Fig. 13. Photomicrographs of greenish-grey clay from the Kanfanar profile. (a) Subangular blocky microstructure with visible gypsum rosettes and impregnation of peds with iron oxides – K-1. (b) Complex microstructure consisting of granular, channel, and subangular blocky microstructure – K-1M2. (c) Subangular blocky to planar microstructure – K-2. (d) Parallel striation and stipple-speckled b-fabric (photograph taken under crossed polarised light) – K-4. (e) Impregnation of iron oxides with the appearance of gypsum rosettes (photograph taken under crossed polarised light) – K-4. (f) Clearly visible granostriation and stipple-speckled b-fabric (photograph taken under crossed polarised light) – K-1M2. (g) Granostriation around clasts rich in typical orthic nodules of pyrite and clearly visible cross-striation of b-fabric (photograph taken under crossed polarised light) – K-3. (h) Clast rich in typical orthic nodules of pyrite – K-3. (i) Aggregated typical orthic nodules of pyrite – K-2.

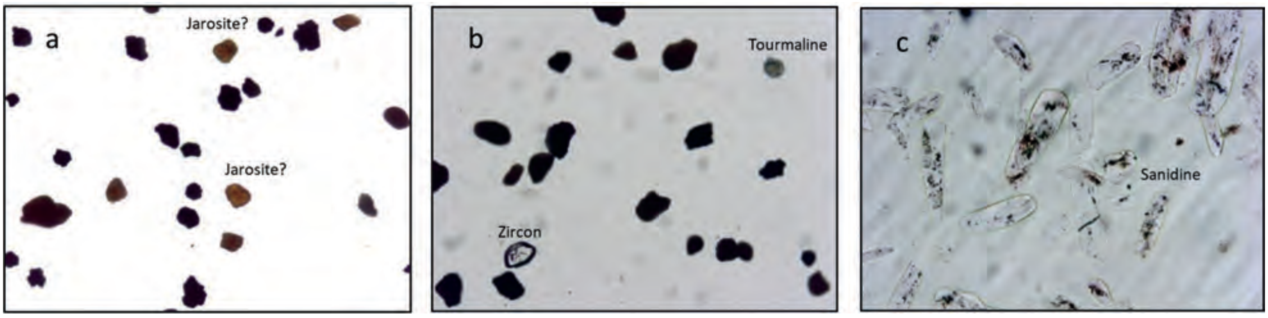

Fig. 14. Photomicrographs of greenish-grey clays – heavy (a and b) and light mineral fraction (c) from the upper part (0–20 cm) of the Kanfanar greenish-grey clay profile

granostriated b-fabric (Fig. 13h, i) alternating throughout the sample.

Based on X-ray diffraction, the greenish-grey clays consist mainly of phyllosilicates (60–90 wt%), followed by K-feldspar (5–10 wt%), gypsum (5–10 wt%), quartz (up to 5 wt%) and a mineral from the Ti-oxide group (anatase, up to 2 wt%). Iron sulphide (pyrite, 2–4 wt%) and sulphate (jarosite, 2–8 wt%) are also observed. It is possible that altered plagioclase – albite – may be present in the analysed samples, but this could not be confirmed with certainty. Phyllosilicates are represented by micaceous (illitic) material (35–40 wt%), followed by illite–smectite (15–45 wt%) of variable composition which occurs in several phases (determined by the ratio of its illitic and smectitic components) and discrete smectite from the dioctahedral montmorillonite–beidellite series (5–7 wt%). Quantification did not account for the amorphous component, the content of which, based on the shape of the central portion of the diffraction curves, is estimated to be between 5–10 wt%. Given the low carbon content, it is most likely associated with poorly crystalline oxy-hydroxides or potentially volcanic glass. The distribution of clay minerals does not exhibit a clear trend along the profile. The insoluble residue of limestone beneath the greenish-grey clay consists mainly of smectite, and also contains illite and kaolinite but no mixed-layered minerals. Of the non-clay minerals, it contains a small amount of goethite (up to 5 wt%).

Heavy mineral fraction (size fraction 32–125 microns) is dominated by opaque minerals, most likely pyrite and jarosite, but also contain zircon and tourmaline (Fig. 14a, b). Light mineral fraction consists almost exclusively of bright anhedral sanidine (Fig. 14c).

Interpretation

During the early Aptian in the area of the Adriatic Carbonate Platform spacious lagoons and low-energy shallows were formed characterised by a very low sedimentation rate. Consequence is the thin succession deposited in a relatively long early Aptian period with common bioturbated levels. During this transgressive phase, the deepening of the sedimentary environment and the deposition in somewhat deeper subtidal environments was characterised by higher biotic diversity (dominated by orbitolinid foraminifera) as a consequence of the relative sea-level rise and ecological disturbances, probably eutrophication caused by an oceanic anoxic event. The lower Aptian deposits represent a regionally recognisable event related to the partial drowning of the carbonate platform, which correlates well with the early Aptian oceanic anoxic event (OAE 1a; e.g., JENKYNS, 1980; MENAGATTI et al., 1998; JONES & JENKYNS, 2001; VLAHOVIĆ et al., 2005; HUCK et al., 2010; CVETKO TEŠOVIĆ et al., 2011). The first 3–5 m of the lower Aptian succession are commonly characterised by the presence of bivalves, manly requieniid shells of *Toucasia* sp. and benthic foraminifera, as well as oncoids of *Bacinella irregularis* RADOIČIĆ. The overlying massive limestones, known as the architectural-building stone "Istrian Yellow" (for its characteristic yellowish colour), are composed of cyclical alternations of mudstones/wackestones and *Bacinella* oncoid floatstones (TIŠLJAR, 1978). The lower Aptian oncoid limestones are very variable in thickness (max. 19 m) as a consequence of different duration of the regional Aptian–Albian subaerial exposure phase in the Istrian part of the AdCP. This unconformity was caused by variable amounts of synsedimentary tectonics and erosion during the late Aptian and early Albian (VELIĆ et al., 1989; MATIČEC et al., 1996) and it is clearly visible with breccias, some conglomerates, marl and clay (Fig. 15). The Kanfanar section indicates a contact between the lower and upper Aptian deposits. Lower Aptian age is confirmed by the presence of the characteristic early Aptian taxa, including *Palorbitolina lenticularis*, whereas the latest Aptian age is suggested by common dasycladacean *Salpingoporella dinarica* and foraminifera *Praechrysalidina infracretacea* deposited in protected to agitated shoals of the carbonate platform. The shallowing tendency is indicated by laminated mudstones/wackestones with fenestral fabric overlain by breccia/conglomerate beds of the first subaerial exposure surface. The end of the deposition in western Istria is marked mainly by shallowing and a long-lasting subaerial exposure (Fig. 15).

Greenish-grey clay found in the palaeokarst depressions of the Kanfanar quarry are remnants of ancient soils that ranged from seasonally marshy to permanently waterlogged conditions. These soils were formed through the erosion and accumulation of surficial soils and sediments during a phase of oscillating marine transgression that

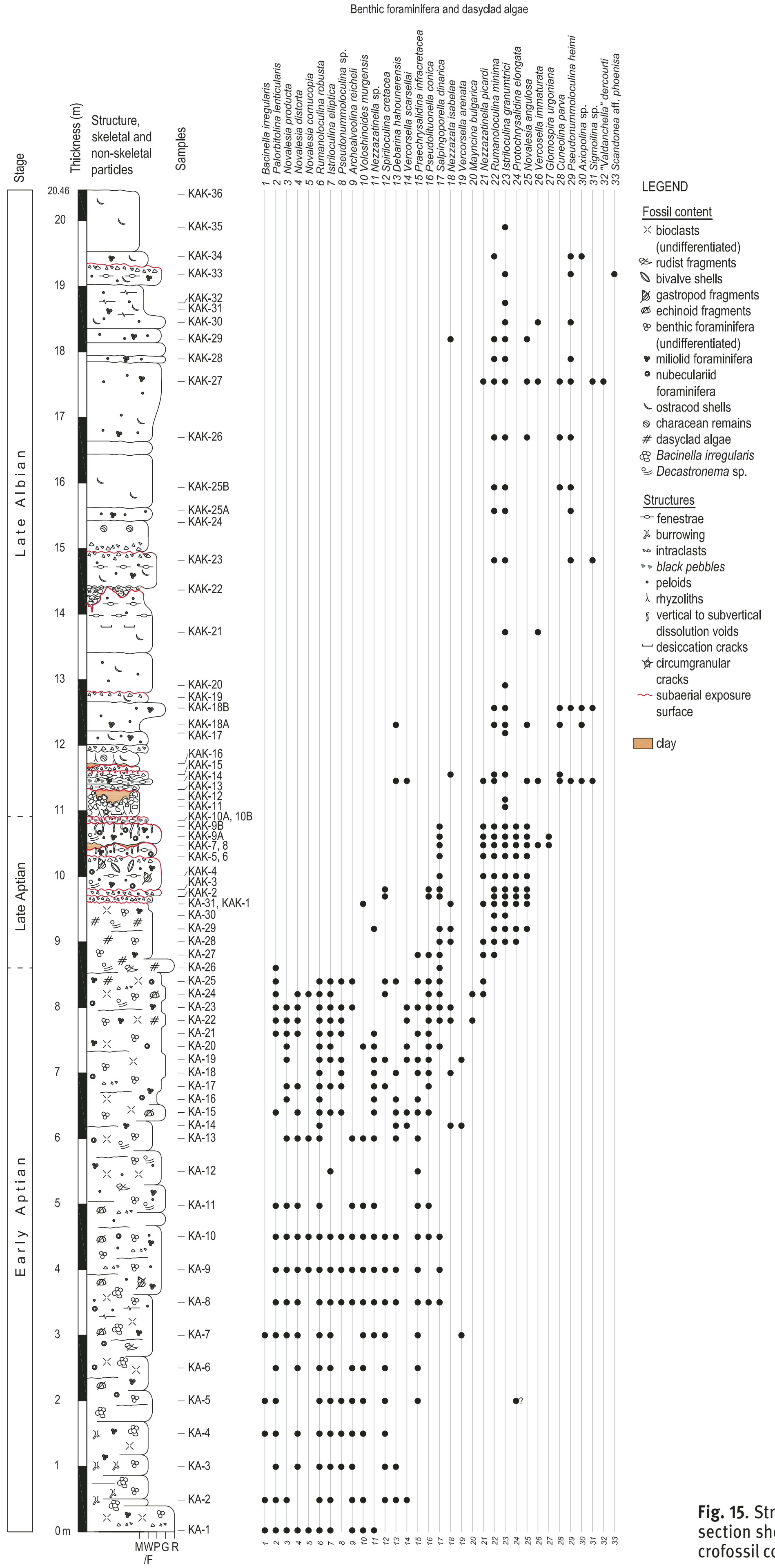

Fig. 15. Stratigraphic column of the Kanfanar section showing structures, textures and microfossil content.

marked the end of the subaerial exposure period. This interpretation has been supported by results of micromorphological analysis that showed granular, subangular blocky, and channel-like microstructures which clearly point to the reworked and bioturbated sediment. This indicates that prior to burial, the sediment was subjected to pedogenic processes over an extended period under conditions of fluctuating high-water levels. Furthermore, the granostriation around clasts (which are weakly rounded, indicating bioturbation processes or very short transportation) also effectively points to pedogenesis. The birefringent fabric of the micromass (b-fabric) shows patterns which are likely the result of various processes such as sedimentation, soil formation, and diagenesis.

Clay mineralogy do not show clear vertical trend along the Kanfanar profile, as it was found in Tri jezerca quarry where there is indication of smectite through the mixed-layered illite–smectite to illite transformation by fixation of potassium from plants, marine waters, volcanic dust and other sources through wetting and drying cycles which was supported by experiments. Deposits that bear strong sedimentological and palaeogeographical resemblance can be found in Jurassic–Cretaceous marine carbonate sequences of the Swiss and French Jura Mountains. These formations are referred to as Purbeckian sediments and have been described by DECONINCK et al. (1988).

Considering the isolation of the Adriatic Carbonate Platform during the Aptian–Albian, probably three potential source materials contributed to the formation of the greenish-grey clay related to the upper Aptian–upper Albian unconformity: (1) insoluble limestone residue; (2) aeolian dust; and (3) volcanic material. The limited presence of insoluble limestone residue, signs of vadose zone (geopetal structures, crystal silt), and shallow palaeokarstification in limestone suggests its minor contribution. Significant atmospheric circulation and accumulation of a large amount of aeolian material during Aptian and Albian due to arid climate on land and presence of metamorphic minerals (such as kyanite, garnet, staurolite, epidote and clinozoisite found in the Tri jezerca clay) indicate minor contribution of aeolian material derived from the surrounding continental blocks. Indicators such as immobile chemical element ratios, a negative europium anomaly, altered volcanic glass, and especially the abundance of bright anhedral sanidine suggest that the parent material from which the greenish-grey clays formed is largely volcanic in origin.

The late Albian microfossil assemblages from western Istria consist predominantly of the characteristic late Albian forms (“*Valdanchella*” *dercourti*, *Pseudonummoloculina heimi*, *Cuneolina parva* etc.). The low diversity assemblages without typical index fossils is probable of the late Albian age containing common miliolid forms *Istriloculina granumtrici*, *Rumanoloculina minima*, *Sigmilina* sp., *Axiopolina* sp. and less common *Scandonea* aff. *phoenissa* and *Glomospira urgoniana* (CVETKO TEŠOVIĆ et al., 2011). Depositional environments with the determined upper Albian deposits exhibit features of the so-called oscillating transgression (TIŠLJAR et al., 1995) characterised by peritidal limestones, high-energy breccias/conglomerates (e.g., the Kanfanar quarry), common laminated deposits with well-developed fenestrae and stylolites.

STOP 3 – HISTORIC BAUXITE MINE, MINJERA (THE PALAEOGENE BAUXITE-BEARING AREA OF MINJERA): UPPER CENOMANIAN/UPPER SANTONIAN–LOWER EOCENE UNCONFORMITY

The 3rd stop comprises a group of smaller Palaeogene bauxite bodies located in the valley of the Mirna river in Istria, called Minjera bauxites. Compared to other Palaeogene bauxites from Istria, they are characterised by their complete or partial reduction and subsequent pyritization. Such pyritized bauxites have been documented throughout the Jurassic and Palaeogene in the Mediterranean bauxite belt, with the best examples found in Montenegro (DRAGOVIĆ, 1989; RADUSINOVIĆ & PAPADOPOULOS, 2021), Hungary (BARDOSSY, 1982), and Greece (ECONOMOU-ELIOPOULOS et al., 2022; LASKOU & ECONOMOU-ELIOPOULOS, 2007, 2013). This pyritization phase is epigenetic in all these deposits as well as in the Minjera bauxites and is related to the transgression that succeeded the bauxitization phase. This unique grey bauxite from the Minjera locality was mined in the past for the production of alum and vitriol, which were obtained by processing pyritized bauxite. Only the grey bauxite was used as ore, while the mined non-pyritized red bauxite, which occurred in some Minjera bauxites, was left on tailing heaps in the area. Mining in Minjera probably dates back to the 16th century, but was historically documented from 1784 to 1824 (D’AMBROSI, 1926), when the mine was in operation. The Minjera bauxite deposits are of great importance as they are the first locality where bauxite was mined in the world. There are visible differences in the type of mining between the individual bauxite bodies, as many larger bauxite bodies contain adits indicating underground mining, while some of the smaller bauxites do not contain adits and were probably surface mined. Most of the deposits are present as large and sub-vertical canyon-type ore bodies (20–30 m thick), while the rest are smaller (< 5 m thick) and have mainly sinkhole morphology. The main part of the canyon-type deposits is located on the northern side of the Mirna River in this area, while the sinkhole type and smaller canyon-type deposits are found on the southern side, indicating the variability of the palaeotopography during the formation of the Minjera bauxites (Fig. 16).

Regional and geotectonic setting

Minjera bauxites are an example of the Palaeogene bauxites found throughout Istria, and their formation coincides with the widespread bauxitization event in the Perimediterranean area, during which a variety of bauxites developed, extending from Hungary through Slovenia, Croatia, Bosnia and Herzegovina, Albania and Greece. This widespread bauxitization event coincides with the Late Cretaceous closure of the Vardar Ocean (SCHMID et al., 2008, 2020; VAN HINSBERGEN et al., 2020), which

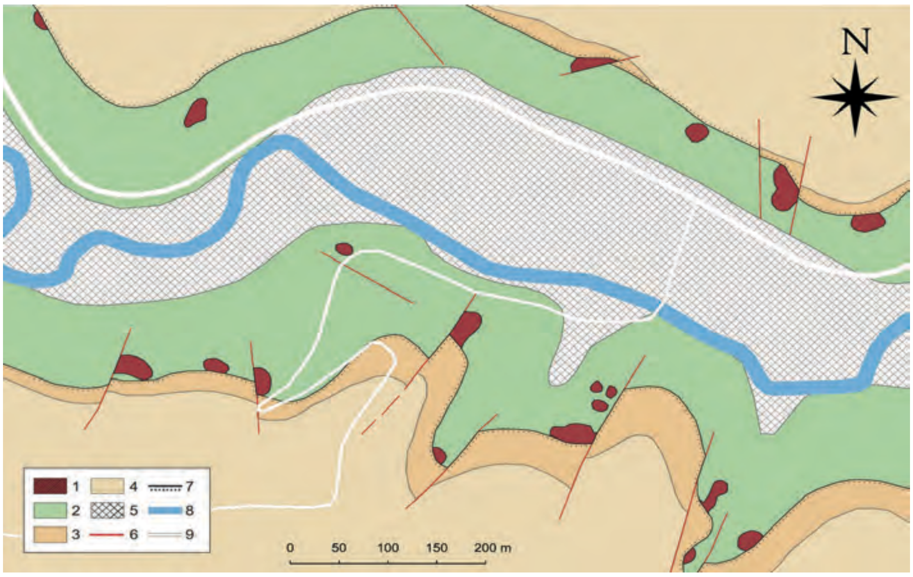

Fig. 16. Geological map of the Minjera bauxites area, modified after ŠINKOVEC et al. (1994). Legend: 1 – Bauxite; 2 – Rudist limestones; 3 – Kozina beds; 4 – Foraminiferal limestones; 5 – Quarternary aluvial deposits; 6 – Faults; 7 – Unconformity; 8 – Mirna River; 9 – Roads.

caused the formation of a foreland basin and deformation that led to the uplift of the advancing carbonate terrains and bauxitization in the emerged areas. In the area of the Adriatic Carbonate Platform the flexural foredeep formed, externally accompanied by the development of the flexural forebulge in response to the overburden pressure generated by the advancing nappes. The emerged areas in the developed flexural forebulge likely served as the environment in which the Paleogene bauxites of Istria and Dalmatia formed. As such, these bauxites belong to the 1st type – "bauxites in collisional settings" – sensu D'ARGENIO & MINDSZENTY (1995). The duration of this subaerial exposure phase was between 25 Ma (southern Istria and Mt. Učka) and 40 Ma (northern Istria). More than 10,000 bauxite occurrences and deposits are known in Istria, as an evidence of a widespread and intensive bauxitization during this period. As deformation continued, rapid subsidence and disintegration of the former Adriatic Carbonate Platform began, resulting in the temporary restoration of carbonate production in the form of foraminiferal limestones, followed by transitional beds and turbidite deposition, accumulating over the former fore-bulge and bauxites (VLAHOVIĆ et al., 2005).

Geological setting

Structurally, the Minjera bauxites are located in the northern Istrian Buzet–Savudrija brachianticline. The bauxites are located on karstified bedrock consisting of Cenomanian rudist limestones, while their hanging wall consists of brackish to freshwater limestones of Eocene age, also called Liburnian deposits or Kozina beds (ŠINKOVEC et al., 1994). During the early Cenomanian period, there were notable changes in the depositional environment in the study area. Prograding sand bar bodies with rudist and chondrodont bioclasts surrounded by peritidal lagoons with rudist biostromes were formed. Rising sea-level led to the destruction and redeposition of calcareous sand bars. Subsequently, lagoonal limestones with rudist and chondrodont coquinas and pelagic influence developed. Increased pelagic influence was observed in younger limestones, featuring calcisphaeres, silicisponge spicules, algae, and planktonic foraminifera (VLAHOVIĆ et al., 1994). Shallowing-upward sequence marks the final deposits in the middle Cenomanian, which consisted of rudist mounds and biostromes together with associated low-energy shoals. These limestones are karstified in the Mirna river valley, and overlain with numerous bauxite bodies and transgressive Palaeogene beds (VLAHOVIĆ et al., 1994). Karstification proceeded along joints intersections, developing numerous vertical sinkholes and decameter-sized karst canyons that served as sediment traps and places where bauxite may have formed and accumulated. This indicates that this carbonate terrain must have been uplifted at least several dozen metres above the sea level. The red colour and oolitic nature of the Palaeogene bauxites that developed in northern Istria are consistent with the highly elevated and intensely karstified carbonate bedrock, as such bauxites belong to the vadose lithofacies (D'ARGENIO & MINDSZENTY, 1995) typical of such terrains. Some bauxite occurrences in northern Istria, such as the Minjera bauxites, were exposed to reduced pore waters during the transgression that followed their deposition, resulting in the reduction of haematite and precipitation of iron sulphides. In some cases, entire bauxite bodies were exposed to this process, which completely transformed them into grey and pyritized bauxite, as is the case with the Minjera bauxites. The transgression was marked with the deposition of the Kozina beds, which are restricted to the immediate hanging wall of the Minjera bauxites, as they indicate deposition in freshwater and brackish ponds which developed in palaeodepressions related to the bauxite-filled sinkholes and vadose canyons. The Kozina beds in this area consist of a sequence of bituminous, gastropod-rich limestones and coal seams. Charophyta oogonies found in most layers point to the freshwater to brackish-water depositional environment, but they alternate with layers abundant in milliolids, indicating a periodic marine influence during their deposition (ŠINKOVEC et al., 1994). The Kozina beds in this area are gradually replaced with foraminiferal limestones, indicating a gradual transition from a freshwater–brackish water environment to an open marine environment.

Mineralogy and geochemistry of the Minjera bauxites

ŠINKOVEC et al. (1994) studied the D-15 bauxite deposit in detail, while recent studies reexamined the D-15 deposit, together with the previously unanalysed D-1 deposit (Fig. 17) and red bauxite left on tailing heaps during the mining of pyritized bauxite from these deposits.

Recent studies have revealed the petrographic, mineralogical, and geochemical differences between the two bauxite bodies. Both the bauxites from D-1 and D-15 deposits originally show an oolitic structure with abundant bauxite clasts (Figs. 18a, b, c, d, e), which is not surprising since the same structure was observed in the nonpyritized red bauxite sample (Fig. 18f).

The D-15 bauxite deposit contains mainly grey bauxite composed primarily of kaolinite and moderate to high amounts of diaspore, with small to moderate amounts of boehmite and iron sulphides (Fig. 19a). Diaspore occurs as a replacement for boehmite and the clay matrix as clusters of 20–50 μm prismatic crystals as well as in veins where it is associated with iron sulphides, clearly post-dating the development of structures and minerals formed during bauxitization. The D-1 deposit consists mainly of pyrite-bearing bauxite (Fig. 18a, c, d), which is predominantly boehmite and iron sulphides (pyrite and marcasite) and contains little or no kaolinite and diaspore (Fig. 19a). Red bauxite is mineralogically most similar to the D-1 bauxites in that they both contain very little kaolinite and a high amount of boehmite, but differ in iron phases (Fig. 19a), which is related to epigenetic pyritization in the D-1 deposit. The major oxide content agrees well with the mineralogical data, as the increased kaolinite content in the D-15 deposit corresponds to higher SiO_2 values, while the increased Fe_2O_3, SO_3, and Al_2O_3 values in the D-1 deposit correspond to higher boehmite and iron sulphide content than in the D-15 deposit. During subrecent weathering of grey bauxite, pyrite was oxidised, producing sulphuric acid that reacted with the phases present in the bauxite and led to the formation of sulphate minerals such as jarosite, pickeringite, and gypsum (Fig. 19a).

The differences in bauxite quality between deposits are probably due to differences in palaeotopography and morphology of the deposits. The bauxite from deposit D-1 is mainly boehmitic and is of higher quality, which is related to the steeper morphology and larger size (> 20 m thick) of the bauxite deposit compared to deposit D-15, which is much smaller and less steep (< 5 m thick). Most of the larger canyon-type bauxite bodies such as the D-1

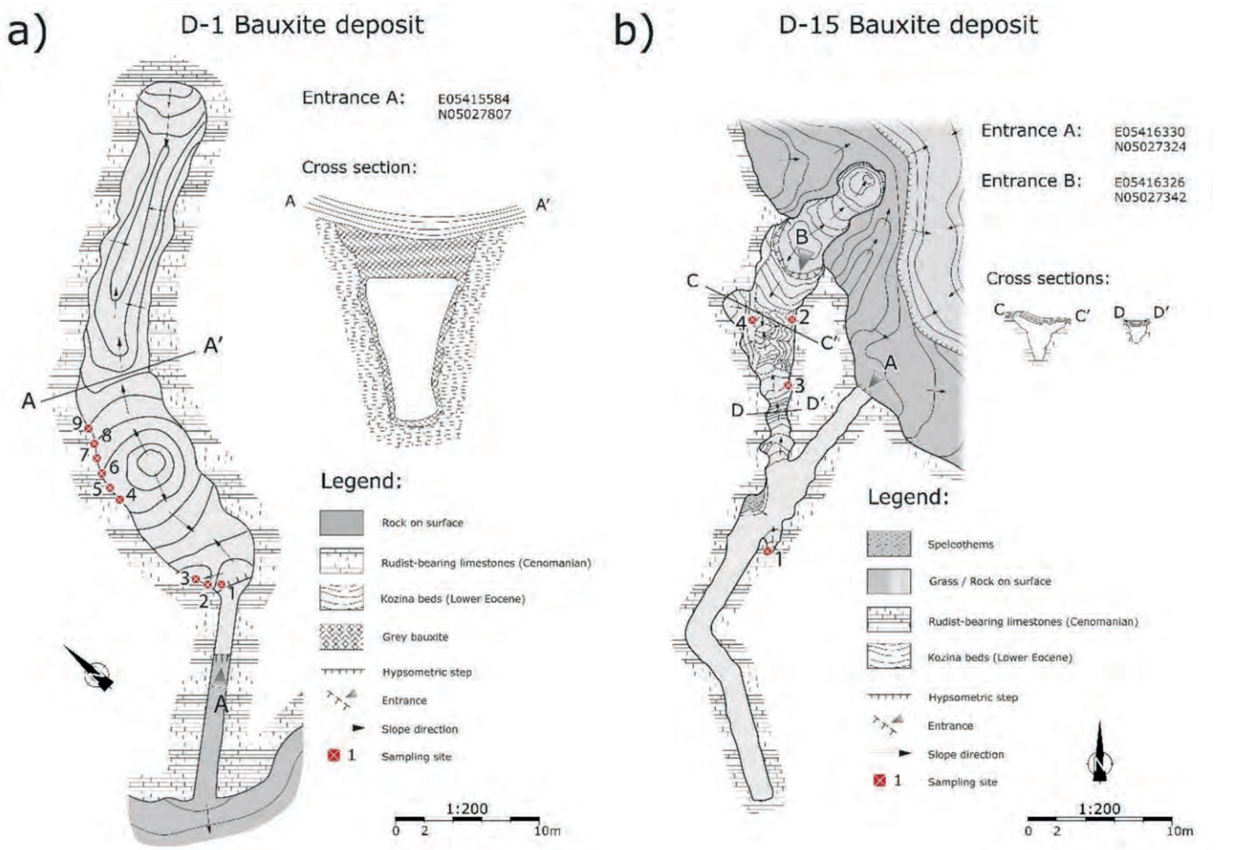

Fig. 17. Plans of the investigated Minjera bauxite deposits and the sampling sites. (a) Plan of the bauxite deposit D-15. (b) Plan of the bauxite deposit D-1.

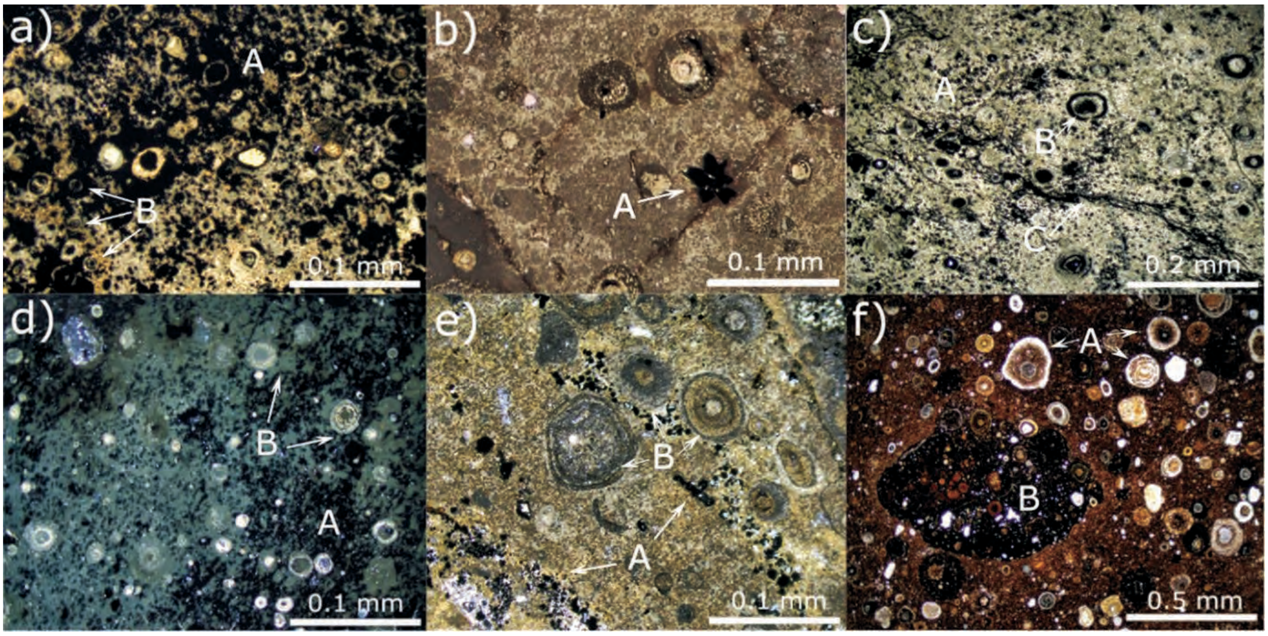

Fig. 18. Photomicrographs of different samples from the Minjera bauxites. (a) First generation of iron sulphides visible in the bauxite matrix (A) together with iron sulphides replacing the iron oxide rich lamellae in the ooids (B), D-1 deposit, PPL. (b) Completely deferrified section of the sample, with visible opaque marcasite rosette (A), D-15 deposit, PPL. (c) Clusters of iron sulphide framboids (A) in the bauxite matrix, iron sulphide rich lamellae in the ooids (B) and iron sulphide veins (C), D-1 deposit, PP. (d) Iron sulphides in the matrix of the bauxite (A), surrounding the diasporitic ooids (B), D-1 deposit, XPL. (e) Iron sulphide framboids (A) together with deferrified bauxite matrix and boehmitic ooids (B), D-15 deposit, PPL. (f) Ooids showing iron oxide rich and deferrified lamellae (A) together with iron oxide rich pebbles (B), red bauxite (MN-25), PPL.

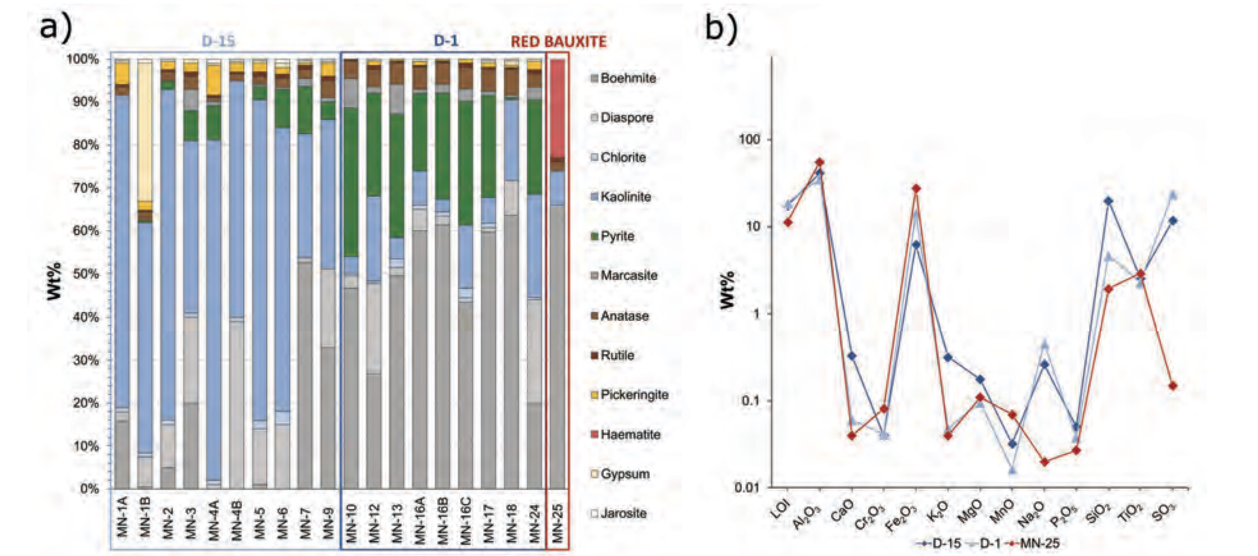

Fig. 19. Plots displaying the mineralogical composition and major oxide composition of the examined samples from the Minjera bauxite deposits D-15 and D-1 as well as the red bauxite sample. (a) Plot displaying the mineralogical composition of the examined samples. (b) Plot displaying the average major oxide composition of the samples from the deposits D-15 and D-1 as well as the values of the red bauxite sample.

deposit are located on the northern bank of the Mirna river, while the smaller canyon-type deposits and sinkhole-type deposits are found on the southern bank of the Mirna river, which probably indicates a higher palaeotopographical position of the bauxite bodies on the northern bank, allowing the formation of deeper karst structures with a stronger fluid flow and leading to more intense leaching during bauxitization.

The morphology of the iron sulphides is particularly diverse in the D-1 deposit and can be divided into several types. The first type of iron sulphides occurs as groups of crystal rosettes and anhedral aggregates that replace the bauxite matrix between the ooids and bauxite pebbles (Fig. 18a–e). The second type involves the replacement of iron oxide-rich lamellae in the ooids by iron sulphides (Fig. 18a, c). Iron sulphides replacing iron oxides in embedded bauxite pebbles represent the third type, while the last type includes iron sulphide veins cutting through previously formed structures (Fig. 18c), textures, and iron sulphide types. This is also visible on a large scale within the D-1 deposit, as centimetre-thick bands of iron sulphides parallel to the bedrock frequently occur in the bauxite of this deposit, indicating structural control during its formation. All of these textures are visible in the D-15 deposit, although less common due to the lower iron sulphide content, and correspond to those described in the D-15 deposit by ŠINKOVEC et al. (1994). The required reducing environment and hydrogen sulphide were produced by reducing pore waters formed in the swamp that formed in the cover of the bauxite. Part of the sulphur probably came from decomposition of organic matter, while most of it was probably derived from microbial reduction of marine-derived sulphate as the marine influence is evident in certain sections of the overlying Kozina beds (ŠINKOVEC et al., 1994). When the bauxite was mostly consolidated after burial, overburden pressure caused fracturing of the bauxite, with iron sulphides crystallising along the newly formed fractures by reaction of undissolved iron oxides with the remaining pore water.

STOP 4 – TERRA ROSSA SOIL PROFILE IN THE KORONIKI VINEYARD: UPPER EOCENE–RECENT UNCONFORMITY

Terra rossa, the term introduced by TUĆAN (1912), is a generic term that refers to a soil formed in a Mediterranean climate that is red in colour, well-structured, high in Fe oxides strongly associated with clay minerals and overlying carbonate rocks (PRIORI et al., 2008). The term is commonly used by pedologists, geologists, archaeologists, geomorphologists, and sedimentologists in the Mediterranean region and in areas with continuously humid climates in the Western (e.g., Caribbean, Indiana, Wisconsin) and Eastern Hemispheres (e.g., southern Australia). In the WRB classification system, terra rossa soils can be classified as Cambisols, Luvisols, Nitisols (PRIORI et al., 2008), Leptosols (e.g., MUHS et al., 2010), and Lixisols (DURN et al., 2023). However, terra rossa is also referred to as a relict soil, polygenetic soil, palaeosol, pedosediment vetusol, lithified palaeosol, pedosedimentary complex, soil sediment, and sediment by various researchers (e.g., CREMASCHI, 1987; MIRABELLA et al., 1992; ALTAY, 1997; BENAC & DURN, 1997; BRONGER & BRUHN-LOBIN, 1997; DURN et al., 1999, 2007, 2014, 2018, 2023; DURN, 2003: FEDOROFF & COURTY, 2013; ZHANG et al., 2018; JONES, 2021; TRAVÉ et al., 2021).

The terra rossa soils on the Istrian peninsula, an archetypal example of a non-isolated karst area influenced by karst processes, (neo-)tectonic activity and input of external materials, are extensively studied polygenetic relict soils or palaeosols with a high content of crystalline iron oxides, composed mainly of haematite and sporadically of goethite, with kaolinite and illitic material representing the main clay mineral phases and nanosized mineral fraction dominated by authigenic kaolinite (DURN et al., 1999, 2001, 2007, 2013, 2014, 2019, 2021, 2023). They overlie and mark the youngest subaerial unconformity in the northwestern part of the Adriatic Carbonate Platform and are susceptible to erosional and redepositional processes. Terra rossa in Istria and generally in the northern Adriatic, fills cracks in karstified limestones (Fig. 20a) and forms a discontinuous surface layer up to several meters thick, which in favorable locations has been used for grape and olive cultivation since Roman times (Fig. 20b). It is also recognised as red polygenetic soil (Fig. 20c), red pedosedimentary complex (Fig. 20d, e), red palaeosol (Fig. 20f, g), and red lithified palaeosol (Fig. 20h). Terra rossa soils represent a valuable archive of information that can be used to understand present and past soil formation processes related to climate variability and landscape dynamics, especially in low-lying parts of the karst landscape where thicker deposits are typically found.

The provenance analysis revealed that the terra rossa allochthonous material originated mainly from the submerged alluvial plain/emerged Adriatic shelf, with two distinct signatures, i.e. Alpine/Apenninic area and Eocene flysch as dominant sources (RAZUM et al., 2023). Other less distinctive source materials include insoluble carbonate

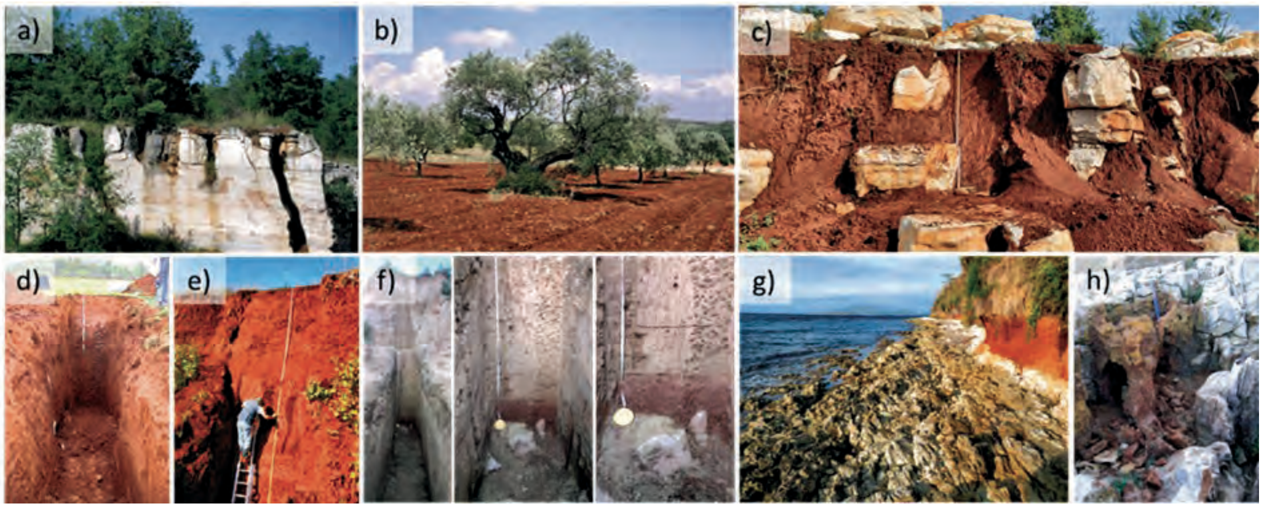

Fig. 20. The appearance of terra rosa soil in Istria and in the northern Adriatic. (a) Terra rossa filling of karstified cracks, Seline. (b) Terra rossa as a substrate for olive cultivation, Novigrad. (c) Red polygenetic soil, Kanfanar. (d) Red pedosedimentary complex, Koreniki. (e) Red pedosedimentary complex, Rovinj. (f) Red palaeosol at the base of an eight-meter-thick loess-palaeosol sequence, Savudrija. (g) Red palaeosol, Susak island and (h) Red lithified palaeosol, Susak island.

residue, tephra, and bauxite material. It is concluded that on a global scale terra rossa is most pronounced on the eastern Adriatic coast, because of the availability of siliciclastic material that was eroded in the Adriatic basin and regularly blown off the emerged shelf surface during periods of low sea levels (Fig. 21). The formation of terra rossa soils in the eastern Adriatic is a recurrent process that has started at least during the Miocene, while the aeolian transport from the emerged shelf may have begun even earlier, during the Oligocene (RAZUM et al., 2023).

Location

The studied soil profile is located on the southwestern Istrian planation surface, called "Red Istria" due to the widespread terra rossa soils, in a vineyard near the village of Koreniki (Fig. 22). On the highly generalized FAO-UNESCO Soil Map of Croatia (scale 1:1,000,000 – BOGUNOVIĆ et al., 1998), "Red Istria" includes almost exclusively Chromic Luvisols and Chromic Cambisols (in places associated with Mollic Leptosols, Dystric/Eutric Cambisols, Albic Lixisols and Anthrosols). The climate of the study region is often described simply as Mediterranean. However, since there are no significant differences in precipitation between seasons, it is actually humid subtropical climate (Cfa) according to the Köppen classification. The soil profile was studied on the site of a grass-covered vineyard in a soil pit excavated to a depth of three meters to the contact with the lower Eocene limestone (Fig. 22a, b). The surface of the limestone, at the point of contact with the soil, appears soft and moist. Occasional weathered limestone boulders and clasts are found in the soil at the contact with the limestone (Fig. 22b). The soil was described and sampled in accordance with FAO (2006) and IUSS Working Group WRB (2022). Details of sampling and analytical methods performed can be found in DURN et al. (2023).

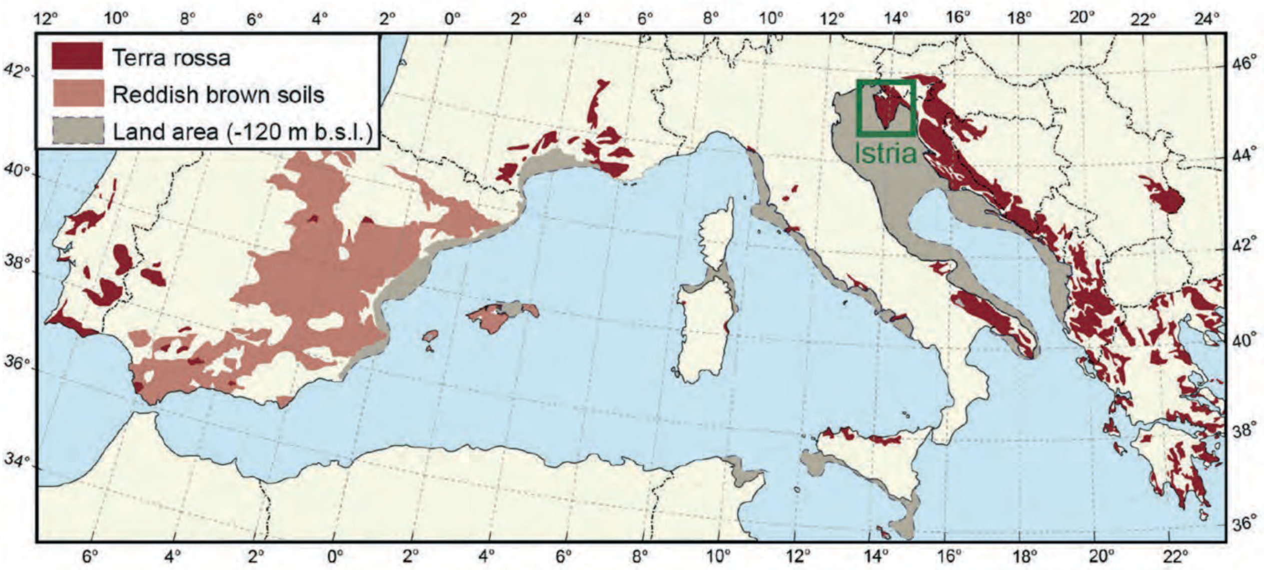

Fig. 21. Distribution map of terra rossa soils and surface shelf during the low sea levels. Large occurrences of terra rossa in the eastern Adriatic are associated with the large shelf area that, when emerged, provided siliciclastic material for terra rossa formation. Although this map shows the most recent palaeogeographic situation, large shallow shelf of the eastern Adriatic coast was probably formed in Oligocene (POPOV et al., 2004), thus from that age, recurrent aeolian transport of emerged shelf material onto the carbonate platform is probable (from RAZUM et al., 2023).

Fig. 22. (a) Terra rossa soil profile located in a vineyard at Koreniki (northwest coast of the Istrian peninsula, Croatia). (b) Cleaned terra rossa soil profile overlying a lower Eocene foraminifera limestone (marked by white arrow) with Ap, Ap/Bt1, Bt1, Bt2, 2Btb1, 2Btb2, 3Btb3 and 3Btb4 horizons.

Pedology and micromorphology

The studied profile has eight soil horizons (Fig. 22b) overlying lower Eocene limestone. All horizons are non-calcareous. The overall depth and general morphology of this profile suggest intense weathering and pedogenesis. The boundaries between soil horizons appear mostly smooth and gradual, except between the Ap and Ap/Bt horizons, where the Ap horizon abruptly and, at places, discontinuously overlies the Ap/Bt horizon. This is due to the soil disturbance by tillage (to a depth of 30–40 cm) and vine roots. In general, the soil becomes redder with increasing depth, indicating an increase in iron content. The soil texture is predominantly clay and generally becomes finer with increasing soil depth. Despite the high clay content, the soil morphology does not indicate significant water stagnation in this profile. The entire profile can be considered porous. Examination of the profile for redoximorphic soil features (mottling) revealed very few to few concentrations of Fe (and possibly Fe/Mn) (Fig. 22b). Due to intense weathering and pedogenesis, pH and CEC values are very low throughout the profile (Fig. 23a). Discontinuities in clay distribution at 170 cm and 230 cm depth (Fig. 23b) suggest that at least two major erosion and sedimentation cycles were involved in the formation of this soil profile, such that a more recent soil (0–170 cm) now overlies the remains of two older, i.e., more relict, soils (2Btb horizon from 170 to 230 cm and 3Btb horizon from 230 to 300 cm depth). A detailed description of the soil pedology and micromorphology can be found in DURN et al. (2023).

The groundmass is generally yellowish red or reddish in the topsoil and becomes distinctly red with increasing depth (in PPL). Microaggregation observed in the thin sections increases with soil depth. The redoximorphic features of the soil in the studied thin sections are usually few and very fine and consist of iron hypocoatings and infillings, depletion hypocoatings, more or less rounded, typic, orthic, and moderately Fe-impregnated nodules, and concentric, strongly impregnated, disorthic and anorthic Fe nodules with sharp boundaries (Fig. 24a). Their abundance does not follow a clear, gradual trend in the profile. In relation to the increasing clay content, the ratio of coarse to fine material of the groundmass (c/f ratio) is about 1:3 in the upper horizons (Fig. 24a) and decreases significantly in the lower horizons (Fig. 24e, f). The coarse fraction consists mainly of coarse silt-sized minerals (mainly quartz, followed by micas and rare quartzite grains). Bauxite clasts and quartz rosette grains were observed in a few samples, indicating colluvial transport from the surrounding area.

The clasts of red soil material (pedorelics) which are often incorporated within the generally less red groundmass suggests that colluvial and aeolian contribution of allochthonous soil material probably played a crucial role in the formation of this profile. The discontinuity between the more recent soil horizons (from Ap to Bt2) and the underlying older horizons (the pedosedimentary layers 2B and 3B), initially identified by the distribution of clay content, is clearly visible in the abrupt change of the groundmass at the Bt2/2Btb1 interface (Fig. 24b). Apart from the Ap horizon, more than 1% textural pedofeatures (mainly clay coatings and infillings) were detected in thin sections, confirming the argic nature of these soil horizons. In gen-

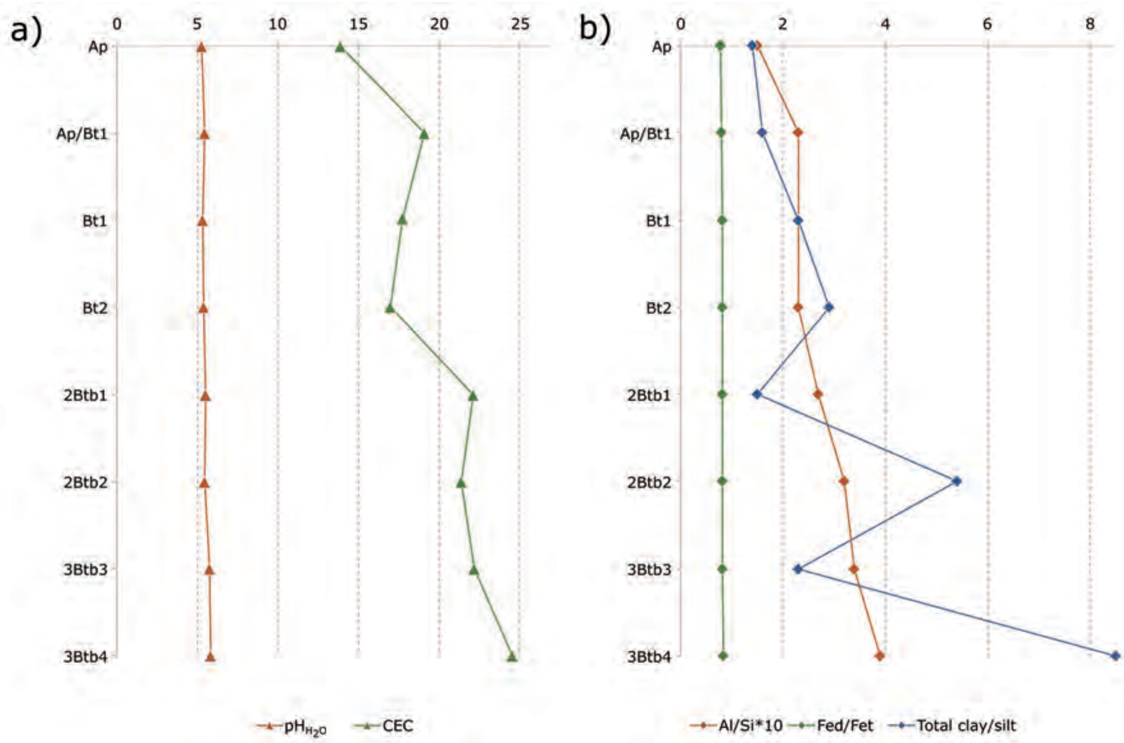

Fig. 23. (a) Distribution of pH_{H_2O} and CEC (cation exchange capacity of the bulk soil) at Koreniki profile. (b) Distribution of Al/Si(*10) ratio, Fe_d/Fe_t ratio (Fe_d – Na-dithionite-citrate-bicarbonate-extractable iron, Fe_t – total iron) and total clay/silt ratio at Koreniki profile.

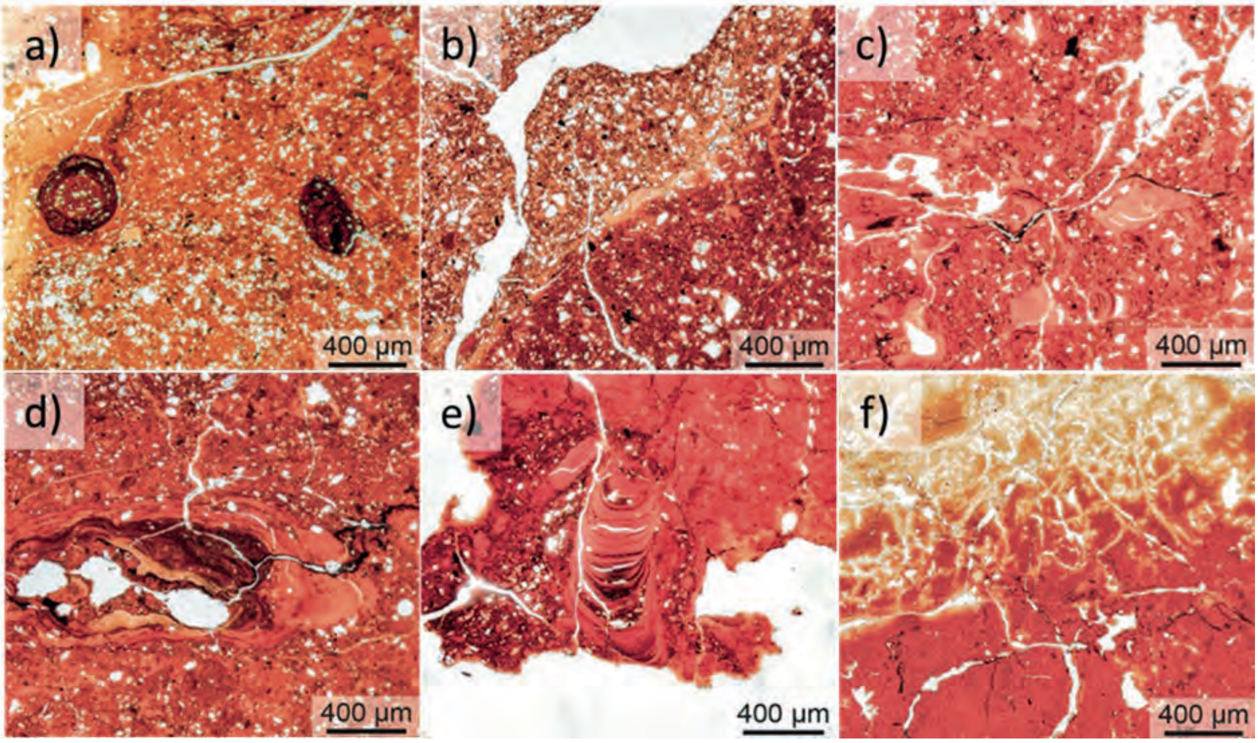

Fig. 24. Photomicrographs of the Koreniki soil profile. (a) Photomicrograph of the Bt2 horizon (PPL). The groundmass is dense and yellowish red. Dominant pedofeatures are nodules strongly impregnated by iron. (b) Photomicrograph (PPL) of the Bt2 horizon (boundary with the 2Btb1 horizon, PPL). Blocky microstructure with accommodated and partially accommodated subangular peds is well developed. Partially relict, dusty clay coatings/infillings are visible along the plane void that forms the abrupt boundary between the Bt2 and 2Btb1 horizons. (c) Photomicrograph of the 2Btb1 horizon (PPL). Microstructure is complex but dominated by planes along which clear clay coatings and infillings developed together with ferromanganese coatings and infillings. (d) Photomicrograph of the 2Btb2 horizon (PPL). The main feature is the alternation of limpid, dusty, and impure clay coatings, with depletion hypocoating overprinting the older illuviated material altogether truncated by younger illuviated material. (e) Photomicrograph of the 3Btb4 horizon (PPL). The groundmass in the upper right quadrant of the photo is denser (finer) than the remaining less dense, coarser groundmass; however, the coarser groundmass incorporates clasts of the material comprising the finer groundmass. In the center of the photo a large, crescent, complete infilling of limpid, dusty, and impure clay coatings is visible. (f) Photomicrograph of the 3Btb4 horizon (XPL). Complex microstructure mostly comprises accommodated angular peds separated by planes and occasionally channels forming a pore network coated with pure clay.

eral, the abundance of illuvial clay increases with the depth of the studied profile. The corresponding features include both pure illuvial clay coatings/infillings (Fig. 24c) and impure coatings of clay and silt (Figs. 24d, e). Both recent and relict illuvial pedofeatures were recognised, with the latter generally impure, deformed, and/or isolated within the former voids (Fig. 24e).

Geochemistry and mineralogy

The Al/Si ratio increases with depth from the Ap horizon to the 3Btb4 horizon and generally agrees with the increase in clay content (Fig. 23b), indicating that lessivage is a very important pedogenic process in the studied soil. Both Fe_t and Fe_d follow the distribution of Al/Si along the profile and show a continuous increase from the Ap horizon to the 3Btb4 horizon (DURN et al., 2023). The Fe_d/Fe_t ratio ranges from 0.79 in the Ap horizon to 0.85 in the 3Btb4 horizon, reflecting a very high degree of weathering of the Fe-bearing primary silicates (Fig. 23b). Due to the very high Fe_d/Fe_t ratio, the investigated soil is one of the most weathered terra rossa in Istria studied so far.

The main mineral phases in the < 2 mm fraction of all horizons are phyllosilicates, followed by quartz and haematite, while plagioclase, K-feldspar, and anatase are the minor mineral phases (Fig. 25a). The predominant mineral phase in the phyllosilicate group is kaolinite, followed by muscovite/illite, chlorite is a minor constituent, while paragonite occurs in trace amounts in the upper three horizons (Fig. 25a). While kaolinite and haematite show a continuous increase from the Ap horizon to the 3Btb4 horizon, quartz and chlorite show an opposite trend (Fig. 25a). Muscovite/illite, plagioclase, and anatase are uniformly distributed with depth (Fig. 25a). The distribution of kaolinite and haematite agrees with the distribution of Al/Si ratio, Fe_t and Fe_d, clearly indicating an increasing degree of weathering with depth and the process of lessivage.

The predominant mineral phase in the clay fraction is kaolinite, followed by illite, and hydroxy-interlayered mineral (HIM) is present in trace amounts (Fig. 25b, HIM is not shown). The kaolinite content shows a clear trend and increases from the Ap horizon to the 3Btb4 horizon (from 76 to 92 %). Illite content shows an opposite trend and decreases with depth (Fig. 25b). The presence of kaolinite and haematite in the studied profile and the absence of chlorite in the clay fraction may indicate that chlorite (together with feldspars, mica, and illite) was one of the main precursors for the formation of authigenic kaolinite and the

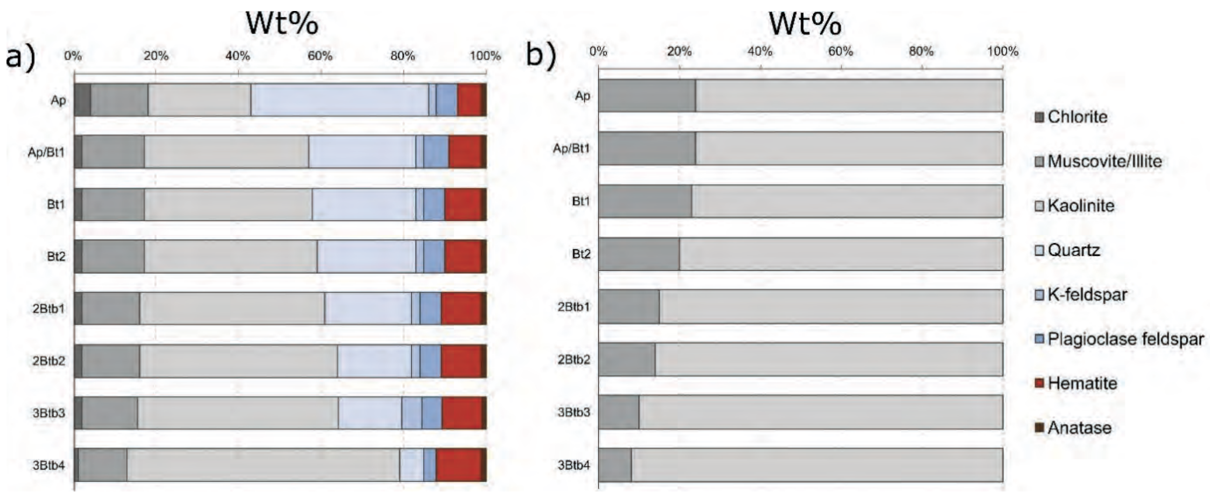

Fig. 25. (a) Mineral composition of the < 2 mm fraction at Koreniki profile (in mass %). (b) Mineral composition of the < 2 μm fraction at Koreniki profile (in mass %).

main source for the formation of iron oxides (haematite). The predominance of kaolinite in the clay fraction and clear evidence for the in situ growth of pedogenic kaolinite nanoparticles (DURN et al., 2023) suggest that kaolinisation, together with the formation of iron oxide (high Fe_d/Fe_t ratio), is a dominant process in the studied profile. Both processes complement ferralitisation, i.e., ferrallitic weathering. A detailed description of the soil geochemistry and mineralogy can be found in DURN et al. (2023).

Soil classification

According to the IUSS Working Group WRB (2022), the studied profile is classified as Rhodic Lixisol (Clayic, Aric, Cutanic). Lixisols are soils with pedogenetic clay differentiation (especially due to clay migration) between a topsoil with lower clay content and a subsoil with higher clay content, low activity clays (mainly kaolinite), and high base saturation at some depth (IUSS Working Group WRB, 2022). Lixisols are found in tropical, subtropical, and warm-temperate regions across Pleistocene and older surfaces – they occur mostly in the (sub)Sahelian and East Africa, but also in South and Central America, the Indian subcontinent, and Southeast Asia and Australia (IUSS Working Group WRB, 2022). Accordingly, the identification of a Lixisol in Istria represents a rare finding.

DURN et al. (2023) concluded that the finding of a Lixisol in the study area indicates the old age of the surface of the karst depression and provides the first data on the burial history of the southwestern Istrian planation surface. This suggests that some terra rossa soils previously classified as Cambisols or Luvisols may actually be Lixisols or other tropical soils (e.g., Nitisols) with preserved relict properties. DURN et al. (2023) also suggested that favourable periods for the formation of the studied soil in the northernmost part of the Mediterranean were older Quaternary interglacials, the mid-Piacenzian Warm Period (Pliocene), and/or the Miocene Climatic Optimum.

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FIELD TRIP A3
WHITE ISTRIA

FIELD TRIP A3 – WHITE ISTRIA (PLATFORM CARBONATES, ARCHITECTURAL-BUILDING STONE)

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INTRODUCTION

Based on the geological structure and different soil types, KREBS (1907) divided Istria into three large-scale units: White Istria, Gray Istria and Red Istria, based on historical Italian names *Istria bianca*, *Istria grigia* and *Istria rossa*. That division has remained widely accepted until today.

The term White Istria refers to the areas of Ćićarija Mt. and Učka Mt. in the northeastern part of Istria, as an elevated, karst dominated area predominantly built of light-coloured Cretaceous and Palaeogene limestones. The Gray Istria is the central part of Istria, which represents a depression filled with turbidite deposits (Pazin Flysch Basin), while the Red Istria represents the southwestern and western part of the Istrian peninsula, whose red colour in the name is due to the relatively large amount of red soil (*terra rossa*) that covers the Jurassic, Cretaceous and Palaeogene carbonate rocks.

Besides the White Istria (Učka Mt. and Ćićarija Mt.), the field trip A3 includes two stops that, according to the aforementioned division, belong to the area of Red Istria – Stop 1 (Kirmenjak Quarry) and Stop 5 (Fantasia Quarry), two well-known quarries in the area of the Western Istrian Anticline. The Kirmenjak Quarry represents the largest and stratigraphically oldest site of dinosaur tracks in Croatia found within stylolitized limestones of the Upper Tithonian age. The Fantasia Quarry, protected as a Geological Natural Monument, represents a unique example of the alternation of genetically different types of Lower Cretaceous dolomites with very well-preserved textures and structures.

STOP 1 – KIRMENJAK QUARRY: DINOSAUR TRACKSITE

Aleksandar Mezga, Damir Palenik, Ladislav Fuček, Igor Vlahović

Introduction

Nearly all the remains of dinosaurs on the Adriatic Carbonate Platform (AdCP; remains of which are now cropping out along the NE coast of the Adriatic Sea and its hinterland) are from Cretaceous deposits of the Istrian peninsula. Istria is located on the NW part of the Croatian coast, covering an area of app. 3000 km^2. Among those remains are numerous dinosaur footprints and bones. Although the AdCP was formed as an independent paleogeographic unit during the Early Jurassic (VLAHOVIĆ et al., 2005), there are no dinosaur finds older than Late Tithonian. The sauropod track records on the AdCP include numerous sites from the Upper Albian and Upper Cenomanian deposits. The main feature of AdCP sauropods is their small size when compared to other Cretaceous ichnites (DALLA VECCHIA, 2002), with pes length rarely exceeding 40 cm. Only narrow and medium-gauge sauropods have been recorded on the AdCP so far, with no evidence of gregarious behavior or parallel trackways.

The Kirmenjak dinosaur locality is situated in the Kirmenjak quarry, about 2 km to the south of the Sv. Lovreč–Poreč road, near Kirmenjak village in western Istria. It represents the largest site with dinosaur tracks discovered on the AdCP so far (MEZGA et al., 2007). There are two separated outcrops with footprints located on the formerly active quarry front. The distance between them is about 1 km but they belong to the same trackbearing layer (MEZGA et al., 2017). The northern outcrop (Kirmenjak I) is larger and contains majority of the footprints (Fig. 1) while the southern (Kirmenjak II) has some slightly larger tracks (Fig. 2). The locality is now under the formal protection by the Croatian Government.

Geological setting

The locality belongs to the lower part of the Kirmenjak unit, i.e., to the lowermost part of the second Istrian megasequence according to the presence of black pebble breccias at the site (VELIĆ & TIŠLJAR, 1988). Several shallowing-upward cycles from 0.5 to 1.5 m in thickness can be distinguished. The shallowing-upward cycles usually begin with black pebble breccia or mudstones, followed by fenestral mudstone, and usually end with peloidal packstone/grainstones and grainstones. Dinosaur

Fig. 1. Exposed surface with the dinosaur footprints (Kirmenjak I)

Fig. 2. Kirmenjak II outcrop with the sauropod footprints

footprints were imprinted on top of the shallowing-upward cycle in intertidal fenestral mudstones capped with a thin peloidal packstone/grainstone layer and overlying subtidal mudstone. The formation and preservation of footprints was favoured by a short duration of exposure of muddy sediment and its rapid burial underneath more mud. The oscillatory transgression over the emerged relief marked the beginning of the deposition of the Kirmenjak unit. Depositional transition in the lowermost parts of palaeorelief indicates fresh-water conditions at first, followed by the brackish environments (VLAHOVIĆ, 1999). Deposits in the lower part of the Kirmenjak unit indicate the presence of marsh environments which represent the source for the black pebble breccias (VLAHOVIĆ, 1999). Marshes were gradually replaced with shallow, protected lagoon environments surrounded by wide tidal flats. The dinosaurs walked across these wide tidal flats leaving their footprints.

Ichnology

A total of 971 footprints were registered on the Kirmenjak I outcrop (Fig. 3). Among them, 161 constitute 23 trackways while the others occur individually or in groups. The potential of discovering new tracks is even greater since the quarry front could progress even further in the hillside and expose the new potential areas of trackbearing horizon. The footprints are preserved as imprints (*epichnia* or negative epirelief). There are two different types of footprints discovered on the site – larger circular–elliptic ones and smaller semicircular, horseshoe-shaped ones. Regarding the fact that both types occur within same trackways, it could be concluded that they belong to the same animals. On most of the footprints an expulsion rim is visible, representing compressed waterlogged sediment squeezed from the print by the weight of the dinosaur. Such rims are usually more prominent than the footprints itself and they protect the prints from being obliterated by erosion.

All the footprints at the site belong to the same type of dinosaurs. The arrangement of the footprints in the trackways and their differences in the size and shape

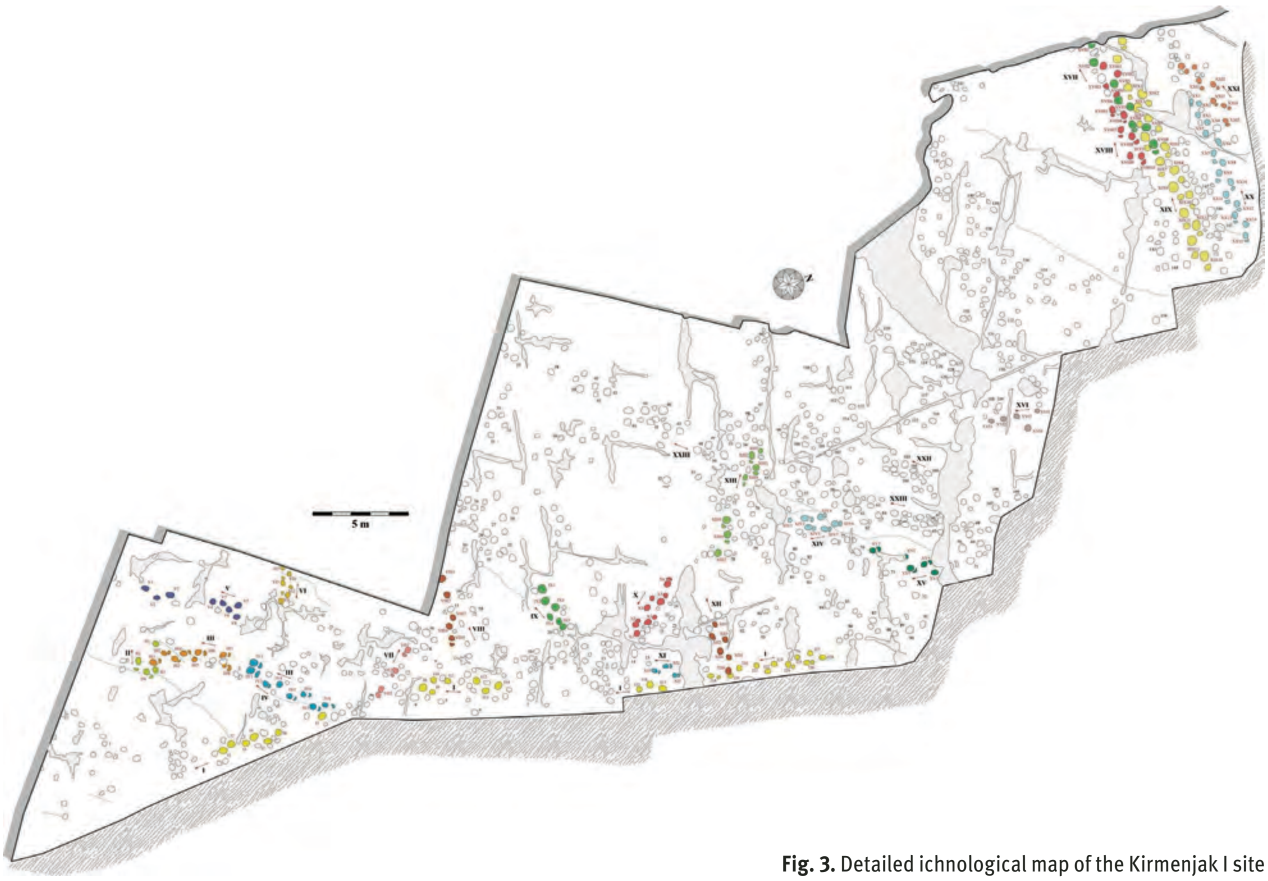

Fig. 3. Detailed ichnological map of the Kirmenjak I site

clearly indicate that the trackways were left by quadrupedal animals. The smaller semicircular footprints represent the manus prints and the larger circular ones the pes prints (Fig. 4). The footprints are attributed to sauropod dinosaurs based on their morphology. The footprints are relatively shallow when compared to their overall dimensions. Although numerous, the state of footprint preservation is far from ideal. In fact, it is difficult to find a footprint with clearly pronounced morphology where the digit impressions would also be recognizable. A large number of footprints are infilled with the sediment that further complicates the recognition of their morphology. When compared to the average size of the Late Jurassic sauropod footprints (e.g., FARLOW et al., 1989; THULBORN, 1990), the Kirmenjak footprints are relatively small, with an average size of 34.5 cm for the pes prints. Manus prints are semicircular to semilunate in shape, wider than long, with the longer axis almost perpendicular to the midline of the trackway. The length of the manus ranges from 5.5 to 26.5 cm with average of 14.2 cm. The manus width ranges from 15 to 29 cm averaging 21.5 cm. The manus prints are wider than long because of the collapse of the footprint margin, which occur anteroposteriorly, causing deformation along its length. Pes prints are oval or elliptical in shape, anteroposteriorly elongated with a somewhat narrower 'heel' impression. The length of the pes prints ranges from 23 to 52 cm with an average of 34.5 cm while their widths range from 18.5 to 43 cm averaging 29.1 cm. None of the pedal prints bears clearly visible digit impressions.

Fig. 4. A couple of sauropod footprints from the Kirmenjak I site. Larger, circular print, represent the pes print and the smaller semicircular, the manus print.

Twenty–three trackways are recognized on the Kirmenjak I outcrop. There are also numerous groups of footprints on the site in which the trackways of several individuals overlap but the precise allocation of single prints to a specific individual is difficult. Trackways in the northern part of the site are especially interesting because they are closely spaced, parallel and have nearly identical orientation. These trackways are better preserved than the other trackways and they also have a greater depth. Because of the same state of preservation and the same direction it could be assumed that the individuals moved through this area in a small-time span or even together. All the trackways in the Kirmenjak quarry are of the narrow-gauge type where the internal trackway width rarely exceeds 10 cm. This excludes titanosaurians but not diplodocoids as possible trackmakers.

Discussion

The footprints found in Kirmenjak quarry are generally shallow for their dimensions which implies that the carbonate mud in which they were formed was rather solid and hard, with greater imprint resistance. Clearly visible expulsion rims around the tracks verify their preservation as true tracks. The relatively shallow depth of Kirmenjak tracks does not necessarily mean that the substrate on which the dinosaurs have walked was dry and exposed to atmospheric conditions; instead, they could have been formed under the water level where the sediment was more saturated by water. The footprint depths increase from the southern to the northern part of the outcrop, what could reflect the different water saturation of the substrate or a longer exposure to erosion. Obviously, all of the tracks at the outcrop were not formed at the same time. Those which were formed first are more eroded than the tracks which were formed among the last.

According to the calculated parameters, the smallest sauropods on the site were about 7.5 m in length while the largest individuals attained some 14.5 m in length. The size of the sauropods from Kirmenjak quarry was smaller than the size of the other Late Jurassic sauropod 'titans' established from skeletal remains. This could be explained by the presence of new sauropod taxa or simply by the presence of smaller species of a larger genus, if we keep in mind that even all species of the same genus were not of the same size.

The estimated speed of Kirmenjak sauropods ranges from 0.5–2.5 km/h, which is not surprising when compared to the speeds estimated for sauropods in other localities. This is the speed in which sauropods usually moved around. It is observed that the distance between the manus and pes prints in the Kirmenjak trackways depends on the speed of the animal – the greater the speed, the longer the distance. In the case of the slow walk, overlap of prints occurs. The pes prints cover the manus prints. This feature indicates that the animals move their legs in the pairs, e.g., front left foot–hind left foot–front right foot–hind right foot, similar to modern elephants. It is the further proof that the gait of sauropods resembled that of the modern quadrupedal mammals more than reptiles.

The Kirmenjak footprints could be assigned to *Parabrontopodus* ichnogenus and the ichnocoenosis could be assigned to the *Brontopodus* ichnofacies which is characterized by sauropod footprints in the carbonate platform environment (LOCKLEY et al., 1994a). The presence of the large sauropod dinosaurs on the Adriatic Carbonate Platform during the Late Jurassic could be explained by its connection with the Gondwana via its southern margins. The sauropods could have migrated in the area during the emergence phase when the platform was exposed to continental conditions. There had to exist a widespread continental area in order to support a survival of such large terrestrial herbivores as the sauropod dinosaurs.

STOP 2 – UČKA: A BRIEF OVERVIEW OF THE GEOLOGY OF UČKA MT. AT THE VELA DRAGA VIEWPOINT

Igor Vlahović, Bojan Matoš, Darko Matešić, Damir Palenik, Ladislav Fuček, Aleksandar Mezga

Introduction

The area of the Učka Nature Park is about 160 km^2. The northern part of the Park includes the southeasternmost portion of the Ćićarija Mt., while the southern part includes the greater portion of the Učka Mt. More information on the geology of the Park can be found in the geological guidebook of the Učka Nature Park (VLAHOVIĆ & MATOŠ, 2021), on which the description of this stop is based. From the geological point of view the area of the Park can be divided into three structural-tectonic units (Figs. 5 and 6): the Imbricate and Nappe Structure of Ćićarija in the north covering an area of about 78 km^2, the Učka Anticline of roughly the same area in the south, and the Učka Klippe to which the summit area of Učka belongs, with a total area of only 4 km^2.

The part of the Park's area that belongs to Ćićarija is actually almost the same size as the part belonging to Učka. Another interesting fact is that new Poklon Visitor Center of the Učka Nature Park is located on the Ćićarija Mt., a few hundred meters NE of the surface border with the Učka Mt. (Ćićarija structure is here thrusted over the Učka Anticline). The same is the case with the 5,062 m long Učka road tunnel, which is drilled through the contact area of Učka and Ćićarija Mts. – the first third of the tunnel on the Kvarner side geologically belongs to Ćićarija, and if we look at the terrain above the tunnel tube, as much as two-thirds of the tunnel would actually belong to Ćićarija Mt. (Fig. 7).

One of the most important features of all three structural-tectonic units of the Učka Nature Park is their exceptional tectonic disturbance, due to which numerous

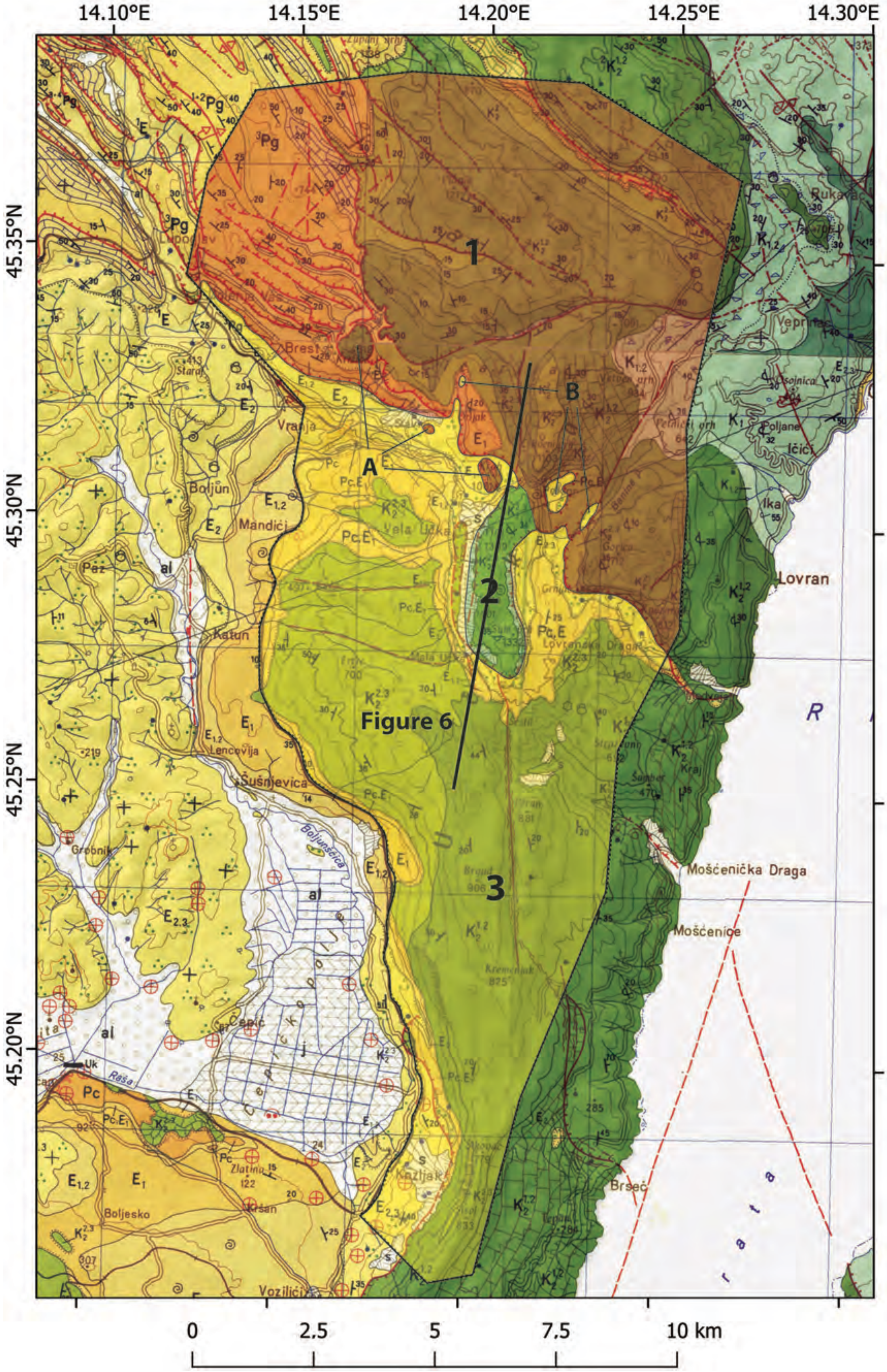

Fig. 5. Division of Učka Nature Park into three structural-tectonic units (VLAHOVIĆ & MATOŠ, 2021): 1) Imbricate and Nappe Structure of Ćićarija in the northern part with three tectonic klippes (A) and three tectonic windows (B); 2) Učka Klippe in the central part; 3) Učka Anticline in the central and southern part. In the middle part the profile line of the geological cross-section shown in Figure 6. The geological map combines the southern part of the Ilirska Bistrica sheet (ŠIKIĆ et al., 1972) and the northern part of the Labin sheet (ŠIKIĆ et al., 1969) of the Basic Geological Map of SFRY in 1:100,000 scale.

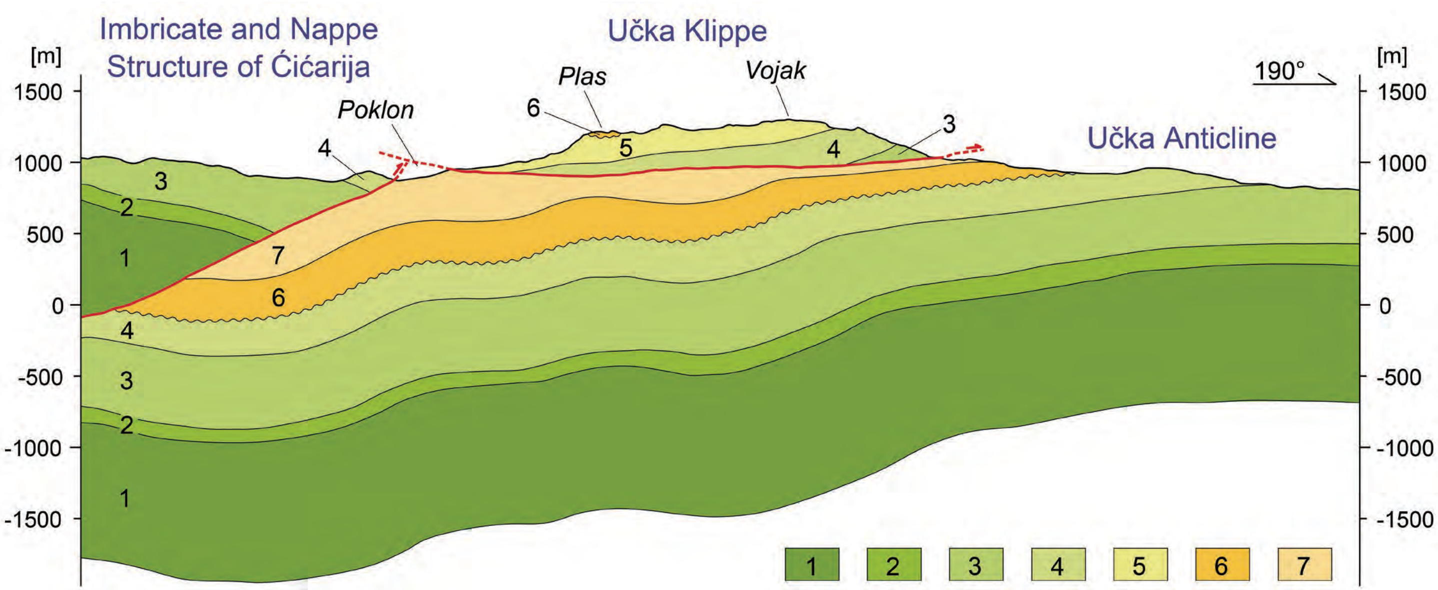

Fig. 6. Schematic geological cross-section showing contact area between three structural-tectonic units of the Učka Nature Park – Imbricate and Nappe structure of Ćićarija, Učka Klippe and Učka Anticline (for position of the profile line see Figure 5). Legend: 1 – Dragozetići–Crna fms. (Valanginian–Albian); 2 – Sis fm. (Upper Albian–Lower Cenomanian); 3 – Milna–Sv. Duh fms. (Middle Cenomanian–Lower Turonian); 4 – Gornji Humac fm. (Upper Turonian–Coniacian); 5 – Učka fm. (Coniacian–Santonian); 6 – Foraminiferal limestones fm. (Lower–Middle Eocene); 7 – Transitional and turbidite deposits fm. (Middle–Upper Eocene).

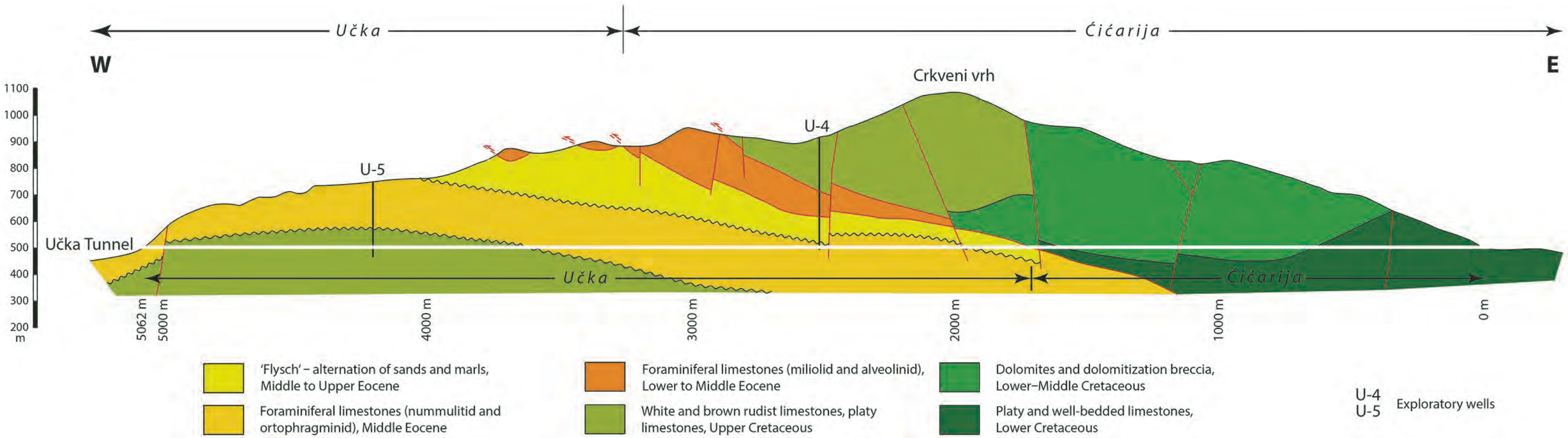

Fig. 7. Schematic geological cross-section made for the Učka road tunnel project (after BABIĆ et al., 1974, in VLAHOVIĆ & MATOŠ, 2021). As the area of the tunnel includes thrust contact between two large structural-tectonic units (the Učka Anticline on the left, western part and the Imbricate and Nappe Structure of Ćićarija on the right, eastern part of the profile), contacts on the surface and in the tunnel do not match. On the surface, approximately two thirds of the area above the tunnel belongs to the Ćićarija mountain range (see black markings above the profile; in adition both klippens west of the contact belong to Ćićarija), while approximately two thirds of the tunnel route belong to Učka (see markings in the lower part of the profile). At the very contact point of Učka and Ćićarija, a large cavern was drilled in the tunnel, which today is used for water supply.

fracture systems are developed, greatly facilitating the chemical weathering of carbonates and the mechanical weathering of turbidite deposits ('flysch'). Therefore, a large part of the Park area is covered with thinner or thicker soil and vegetation, while the outcrops are mostly tectonically very disturbed.

At this stop, we will briefly consider two structural-tectonic units, the Učka Anticline, which belongs to the stable Istria, i.e., it represents a much less disturbed part of the Adriatic Microplate, and the Učka Klippe, which belongs to the Dinaric Mountain range. The third unit, the Imbricate and Nappe Structure of Ćićarija, also representing a part of the Dinarides, will be discussed at the next Stop.

It is important to note that the given order of description does not correspond to the time sequence of the structural formation: the extensive Učka Anticline is the oldest structure, a subsequent change in the main compression direction resulted in a collision alongside its northern edge and the formation of the Imbricate and Nappe Structure of Ćićarija, and ultimately the tectonic transport of the Učka Klippe to the south. Present-day relations are masked due to the erosion that occurred during the long-lasting subaerial exposure phase (the study area has been exposed to intense karstification over the last 30 million years), so it can be assumed that a significant portion of the original deposits, especially in the Ćićarija area, is missing.

The Učka Anticline

The structural-tectonic unit of the Učka Anticline occupies the central and southern part of the Učka Nature Park (Fig. 5), with a total area slightly larger than 78 km^2. It is a gentle, open fold with an approximate N–S axial plane orientation, which clearly differs from the orientation of the structures in the neighbouring Dinarides (characterised by typical Dinaric strike – NW–SE).

According to data from the Basic Geological Map of the Labin sheet (ŠIKIĆ et al., 1969), the oldest Cretaceous deposits of the Učka Anticline are of Cenomanian age, and the youngest are of the Upper Turonian–Coniacian age. The recent data provided by the Croatian Geological Survey indicate that very close to the Park area, in the area of Mošćenica, there are also older, Albian rocks. The western limb of that extensive fold is composed of a continuous succession from older to younger Upper Cretaceous deposits, numerous bauxite occurrences at the Cretaceous–Palaeogene boundary, and a sequence of Kozina deposits and foraminiferal limestones dipping outside the Park boundaries under the turbidite deposits of the Pazin Flysch Basin. The Palaeogene deposits also crop out on the surface at the very top of the Anticline, at an altitude of c. 1000 m above the sea level, where they represent the footwall of the overlying Učka Klippe. In addition, Palaeogene deposits cover the northern edge of the Učka Anticline, i.e., the area bordering the Imbricate and Nappe Structure of Ćićarija, and descend to the east towards Lovranska Draga and the area close to the sea near Medveja. The eastern limb of the Učka Anticline descends all the way to the coast of the Kvarner Gulf, but there are no elements on the adjacent islands that would indicate the continuation of similar structures. By comparing the successions, it is evident that the youngest Cretaceous rocks on northern parts of the islands of Cres and Krk are significantly older than those on Učka, but even more important is a totaly different orientation of their structures, characterised by a typical Dinaric strike, NW–SE. That is why the so-called Kvarner fault is often mentioned in geological literature (Korbar, 2009, and references therein), which is assumed under the sea between the eastern coast of Istria and the islands of Cres and Lošinj. The Kvarner fault represents the boundary between the relatively undisturbed part of the Adriatic Microplate in Istria and the Kvarner islands as parts of the External Dinarides.

In the northernmost part of the Učka Anticline very important elements indicating the character of the tectonic contact with the Ćićarija structures can be recognized, which also indicate the relative age and order of formation of the structures in the Učka Nature Park. The area of the NW slopes of Učka, between Vranja, Vela Učka and the Peruč pass (the highest point of the Lupoglav–Veprinac road, where the road to the highest peak of Učka, Vojak, branches off) is marked by the refolded structures of the Učka Anticline, with folds of the Dinaric strike visible on the surface. A good example of such folds is the anticlinal structure of Vela Draga, because the intense weathering of heavily fractured rocks in its core resulted in gradual widening and deep canyon incision. This indicates that the large-scale Učka Anticline was formed first, followed by several folds with Dinaric strike (NW–SE), as a result of the tectonic collision with the Imbricate and Nappe Structure of Ćićarija in its frontal part. Based on the succession of deposits and the aforementioned structural relationships, it is possible to reconstruct an approximate time frame of the formation of these structures. The Učka Anticline was formed after the deposition of turbidite deposits (because it contains such deposits even on its highest part), which means probably by the end of the Eocene. This fact implies that the folding of the northern part of the Učka Anticline due to the tectonic collision with the Ćićarija range probably took place at the beginning of the Oligocene, a little more than 30 million years ago. It should also be noted that the Anticline was significantly uplifted, because rocks originally deposited in marine environments at a depth of several hundred meters are today cropping out at the elevation of 1000 m above the sea.

The contact with the neighbouring Imbricate and Nappe Structure of Ćićarija along the NE part of the Učka Anticline is also interesting. Two occurrences of younger Palaeogene deposits surrounded by Cretaceous rocks in the form of tectonic windows are visible on the surface, one relatively close to Poklon and the other somewhat further to the southeast. The occurrence of tectonic windows together with several small klippe structures composed of Ćićarija rocks found on top of the marginal part of the Učka Anticline clearly indicate that the contact of Ćićarija Mt. and the Učka Anticline was formed by thrust tectonics, which was definitely proven during the construction of the Učka road tunnel (Fig 7).

The Učka Klippe

The structural-tectonic unit of the Učka Klippe is a very small area (less than 4 km^2) occupying the summit area of Učka (Figs. 5 and 6). It is the highest part of the Park, with altitudes mostly above 1000 m a.s.l., so it also includes the highest peaks: Vojak (1396 m), the nameless peak on the ridge north of Vojak (1352 m), Suhi vrh (1333 m), Plas (1285 m) and Jazvina (1104 m).

The Učka Klippe consists of mostly Cretaceous deposits thrusted over the Palaeogene foraminiferal limestones and turbidite deposits ('flysch') belonging to the underlying structural unit of the Učka Anticline. Although the thrust contact is mostly covered by thick vegetation and screes, the Učka Klippe represents one of the best examples of thrust structures in the Dinarides. The contact with the Imbricate and Nappe Structure of Ćićarija in the north is not well visible because similar stratigraphic units are in contact, but it can be clearly recognized in the area of Peruč, where the road to the Vojak peak branches off from the main road: at a distance of only several tens of meters, there are outcrops of turbidites belonging to the Učka Anticline to the west, Upper Cretaceous deposits of the Učka Klippe to the south and the lowest part of foraminiferal limestones belonging to the Imbricate and Nappe Structure of Ćićarija to the east.

In addition to the well-defined thrust structure the stratigraphic succession of the Učka Klippe also attracts the attention of geologists, especially its younger Cretaceous deposits different in age and origin from all Cretaceous rocks found in the underlying unit of the Učka Anticline and the neighbouring Ćićarija massif. The older part of the Cretaceous deposits in the Učka Klippe unit, clearly visible in its southern part, corresponds to the youngest Cretaceous deposits in the other two structural units, but in the Učka Klippe these rocks are overlain by even younger Cretaceous rocks not found in other parts of the Park (those areas were at that time already uplifted and subaerially exposed). This younger part of the succession consists of two units: the older one characterised by the deposition of calcisphaera limestones in deeper marine environments similar to the deposits of the older Sveti Duh unit, which forms the cliffs along the road to the Vojak peak, and the overlying younger shallow marine rudist limestones building the summit area of Učka Mt. and the area to the north. Within the deposits close the Vojak peak a rich microfossil assemblage has been found (including *Murgella lata*, *Pseudorhapydionina mediterranea*, *Scandonea samnitica*, *Dicyclina schlumbergeri*, *Nummofalotia apula*, etc.), indicating the Upper Santonian age (VLAHOVIĆ et al., 2003), i.e., an age of approximately 84–85 Ma.

Since in the area of the Učka Klippe Paleogene deposits were not determined on geological maps, the question arose whether younger Cretaceous deposits were also deposited but subsequently eroded. However, around the Plas peak an outcrop with transgressive Palaeogene foraminiferal limestones on top of Upper Santonian deposits was found (MATEŠIĆ, 2017), and during the research for the geological guidebook of the Učka Nature Park another similar outcrop was documented a little further south.

The presence of the deposits that cannot be found anywhere else in the vicinity is evidence that the Učka Klippe was transported tectonically from some more remote areas. In the whole area of the stable Istria, such young Cretaceous deposits are only found in the very south, on the Marlera peninsula SE of Medulin (MORO et al., 2002). However, that area is certainly not the place from which Učka Klippe originates: not only because there are no structural elements that would indicate such a significant tectonic transport from the south, but also because those rocks differ significantly from the contemporaneous rocks near the Vojak peak. Therefore, only the area NE of Ćićarija can be assumed as the potential source area for the Učka Klippe, and preliminary research did not rule out such a possibility. Therefore, it may be concluded that the Učka Klippe unit is just a small erosional remnant of a former nappe which thrusted over the entire Imbricate and Nappe Structure of Ćićarija onto the apical part of the Učka Anticline, therefore being the youngest of three structural-tectonic units in the Učka Nature Park.

Viewpoint over Vela Draga – view of the Učka Anticline and the Učka Klippe

The viewpoint located on the northern slopes of the Vela Draga provides a magnificent view of almost the entire canyon, from the eastern end over which the summit area of of the Učka Klippe is clearly visible (Fig. 8), to the south to the abandoned railway line from Lupoglav to Raša (Fig. 9). all the way to the west towards Boljunsko polje (Fig. 10). Vela Draga is around hundred metres deep canyon stretching from the entrance to the Učka road tunnel to the west close to the Vranja village, and then turning to the SW to Boljunsko polje.

The origin of Vela Draga is connected to tectonic processes which resulted in the present-day tectonic structure of the northern edge of the Učka Anticline. Namely, after the uplift of the Anticline, when the turbidite deposits of the eastern part of the Pazin Flysch Basin were significantly uplifted from their original position, a change in the stress field took place, resulting in compression almost perpendicular to the previous one: instead of the main stress oriented E–W, which formed the Učka Anticline structure, the new stress was oriented NE–SW, typical for the Dinarides. Therefore, the area of the Učka Anticline northern edge is folded into several large folds, and the part of Vela Draga between the Učka tunnel and Vranja represents the core of relatively large anticline.

The erosion gradually opened and widened fracture systems formed parallel to the Vela Draga anticline axis, enabling gradual incision of the canyon. The influence of surface water was also important, especially because in the hinge zone of the Učka Anticline impermeable turbidite deposits crop out, resulting in large quantities of surficial water concentrated above the entrance into the canyon. The dissolution of carbonate rocks during both surfical and underground long-lasting karstification process could have been significant, and it is possible that at least part of the present-day appearance of the canyon was also a consequence of the cavern collapses, especially at the very eastern end of Vela Draga.

As a result of intense tectonic fracturation of rocks and significant vertical erosion, numerous erosional columns composed of foraminiferal limestones can be found along the canyon (Figs. 11 and 12). These interesting geomorphological forms also represent a great challenge for alpinists and sport climbers, so that for almost 100 years, from the first ascent of the famous Italian climber Emilio

Fig. 9. View from the Vela Draga viewpoint to the south: at the bottom of Vela Draga there is an embankment of the abandoned railway line from Lupoglav to Raša, at both ends of which there are tunnels. On the other side, high cliffs of foraminiferal limestones can be seen, which represent the western limb of the extensive Učka Anticline, which dips to the west under the turbidite deposits of the Pazin Flysch Basin.

Fig. 8. View from the Vela Draga viewpoint to the east: in the foreground the canyon of Vela Draga, which forms the eroded central part of the extensive anticline structure in the NW part of the Učka Anticline structural-tectonic unit, and in the distance a much steeper relief of the Učka Klippe structural-tectonic unit

Fig. 11. View to the east from the beginning of the embankment of the abandoned Lupoglav to Raša railway line: in the higher part of the Vela Draga canyon, in the northern limb of the large-scale anticline formed by the refolding of deposits of the Učka Anticline, close to a contact with the Imbricate and Nappe Structure of Ćićarija, there is an entrance to the Učka road tunnel (in the central part of the photo)

Fig. 10. View from the Vela Draga viewpoint to the west: the canyon of Vela Draga continues towards Boljun Polje where the Palaeogene foraminiferal limestones dip under the turbidite deposits of the Pazin Flysch Basin

Fig. 12. View from the embankment of the abandoned Lupoglav to Raša railway line towards the Vela Draga viewpoint. The high cliffs composed of Palaeogene foraminiferal limestones have been used for sport climbing ever since 1931. Those geomorphologically interesting tower-like forms were formed due to intense tectonic fracturation and selective erosion along fracture systems.

Comici back in 1931, they have remained popular meeting place until today.

At the very bottom of the canyon the oldest deposits crop out, dark gray Kozina limestones with numerous miliolids and gastropods of the genera *Cosinia and Stomatopsis*, and along the canyon slopes younger foraminiferal limestones mostly composed of entire and broken alveolinids and nummulitids crop out.

Pupićina Peć in Vela Draga – what were people doing there several thousand years ago?

Vela draga is also a very important archaeological site where it has been proved that humans lived from approximately 12,000 years ago (about the end of the last ice age) until the Roman times. Pupićina Peć, which was described for the first time by MALEZ (1960), was thoroughly researched, and after careful and long-lasting excavations, archaeologists found very valuable evidences about who and how lived in that semi-cave (MIRACLE, 2001; MIRACLE & FORENBACHER, 2005; BOSCHIAN, 2006; LIGHTFOOT et al., 2011).

In the period between ten and eight thousand years BCE the people who lived in Pupićina peć (Fig. 13) fed on the meat of wild game they haunted in the vicinity (e.g. bones of roe deer, deer, boar, but also of wild cattle, chamois, badgers, rabbits, beavers, foxes, weasels and wild cats were found there) and made use of their fur to make clothing, their teeth to make jewellery and their horns to make weapons and tools. Besides that, the remains of shells that they brought from the sea some 20 km away that were, as well as traces indicating that they gathered plants were also found.

There is no evidence that Pupićina Peć was used in the period between 8000 and 5500 BCE either as a dwelling place or a shelter, but the period between 5500 and 3500 BCE is very well documented. Major changes affecting the lives of people took place at that time as systematic agriculture and animal husbandry were introduced, so that shepherds who raised sheep and goats, less often cows and pigs, lived in Pupićina Peć. Abundant remains of dishes made of baked clay and more advanced tools made of chert and volcanic glass were also found there (although chert can be found in vicinity, volcanic glass must have come from far away areas).

An even younger layer at Pupićina Peć dates back approximately to 1500 BCE, with the remains of typical Bronze Age dishes, but also of needles and spindles as proof of more advanced techniques of clothes making. The youngest layer dates to the first centuries CE, when shepherds probably brought Roman ceramics with them, of which only a few remains left.

Fig. 13. Archaeological excavation in Pupićina Peć, which explored the layers with extensive relics of the life activities of the humans who inhabited the semi-cave for shorter or longer periods during the last 12,000 years

STOP 3 – ĆIĆARIJA: GEOLOGICAL AND STRUCTURAL ARCHITECTURE OF THE CENTRAL PART OF THE ĆIĆARIJA MT.

Damir Palenik, Igor Vlahović, Dubravko Matičec, Ladislav Fuček, Bojan Matoš, Aleksandar Mezga

Introduction

Ćićarija Mt., with its characteristic imbricate and thrust structure, belongs to the tectonically very complex marginal area of the NW Dinarides. The fold and thrust belt of Ćićarija Mt. stretches along the NE edge of the Istrian peninsula, and along its SW margin, Ćićarija is morphologically uplifted above the surrounding terrain, including its highest peak Veliki Planik (1,272 m a.s.l.). With its NW–SE strike typical for the Dinarides, Ćićarija Mt. is located between the Brkini syncline (comprised of Palaeogene deposits – limestones and flysch deposits) to the NE, the Pazin flysch basin to the SW, and the Rječina valley to the SE (ŠIKIĆ & PLENIČAR, 1975). Ćićarija Mt. can be subdivided into two parts: the SW part predominantly built of Palaeogene deposits and NE area built mostly of Cretaceous deposits. The entire area is intensely tectonized by numerous faults with a dominant typical Dinaric NW–SE strike (locally deviated to NNW–SSE), but also by NNE–SSW striking transversal faults.

In a morphological sense, the thrust and imbricated structure of SW Ćićarija forms a series of morphological terraces on top of each other (Fig. 14).

Each individual morphological terrace typically consists of a sequence of older Palaeogene foraminiferal limestones thrusted above younger foraminiferal limestones or above transitional marly deposits, only locally including thin flysch deposits.

Description of the central part of the Ćićarija Mt.

The study of central part of the Ćićarija Mt. was conducted within a work on the Basic Geological Map of the Republic of Croatia 1:50,000 Scale project. The main objectives of investigation were focused on the geological composition, reconstruction of the tectonic movements, and identification of the kinematics of the formation of

Fig. 14. Panoramic picture of the Ćićarija Mt. and surrounding area. View from Učka Mt. towards imbricated structures of Ćićarija in the northwest. The tectonic transport of the Ćićarija structures is towards the southwest (to the left) (modified from PALENIK, 2020).

geological structures in this part of the NW Dinarides. The new Basic Geological Map of the central part of the Ćićarija Mt. and the marginal area of the Pazin flysch basin in the 1:50,000 scale covers a total area larger than 220 km^2 (Fig. 15). On the new geological map, thirteen informal lithostratigraphic units have been identified, named after typical penecontemporaneous units already defined in other parts of the former Adriatic Carbonate Platform (GUŠIĆ & JELASKA, 1990; FUČEK et al., 1995, 2012; VLAHOVIĆ et al., 2005). A well-established vertical and lateral relationships of the studied deposits, along with a comparison to penecontemporaneous units of the wider area of Istria and other parts of the former AdCP, were necessary prerequisites for the interpretation of the Ćićarija area tectogenesis.

The reconstruction of the tectogenesis of the study area included construction of 11 geological cross-sections normal to the typical Dinaric strike of the main structures. Geological cross-sections show the interpretation of subsurface relationships of structures based mostly on surface data (Fig. 16).

Interpretation

The Palaeogene imbricated structure of the SW part of the Ćićarija Mt. was formed as a result of compression and thrusting of northeastern hinterland composed of Cretaceous and Palaeogene deposits causing significant contraction of the area. Based on constructed geological cross-sections, it can be concluded that the Palaeogene imbricated structure of the SW part of the Ćićarija represents a *thin-skinned deformation* of exclusively Palaeogene deposits that are multiple times repeated, which are physically separated from the underlying Cretaceous basement by a shallow, regional-scale, detachment fault that dips towards the NE at a very low angle (Fig. 17). Regional detachment probably involved the lowermost part of the Foraminiferal limestones, i.e., miliolid limestones, which contain a substantial bituminous content and as such could be suitable for the development of the sliding surfaces and shear planes. Thrust contact of the entire Ćićarija Palaeogene imbricated structure and overlaying Cretaceous deposits dips at the very low angle towards the NE, indicating SW-directed tectonic transport (Fig. 18).

In the area of the Palaeogene imbricated structure reverse faulting show typical *ramp and flat geometry*. Marls from the Transitional deposits, and very rarely (or locally marls from the lowermost part of the flysch sequence) are proposed as detachment horizons within the imbricated structure system (Figs. 19 and 20).

For determination of fault kinematics in relation to the past and present stress fields, collected data of field

Fig. 15. Location map with shaded study area (modified from PALENIK, 2020)

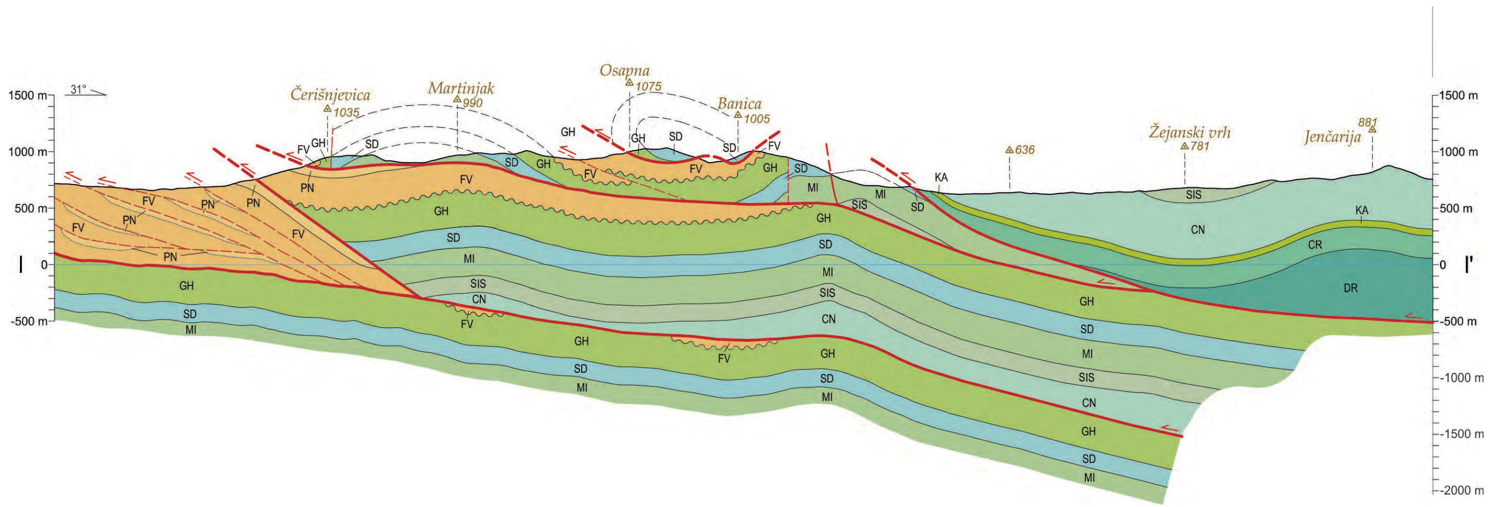

Fig. 16. An example of a typical transversal (SW–NE oriented) geological cross-section in the Ćićarija Mt. area. (PALENIK, 2020) Legend: DR – Dragozetići fm. (Berriasian–Hauterivian); CR – Cres fm. (Barremian); KA – Kanfanar fm. (Aptian); CN – Crna fm. (Albian); SIS – Sis fm. (Upper Albian–Lower Cenomanian); MI – Milna fm. (Middle–Upper Cenomanian); SD – Sv. Duh fm. (Upper Cenomanian–Lower Turonian); GH – Gornji Humac fm. (Upper Turonian–Coniacian); FV – Foraminiferal limestones fm. (Lower–Middle Eocene); PN – Transitional deposits fm. (Middle Eocene).

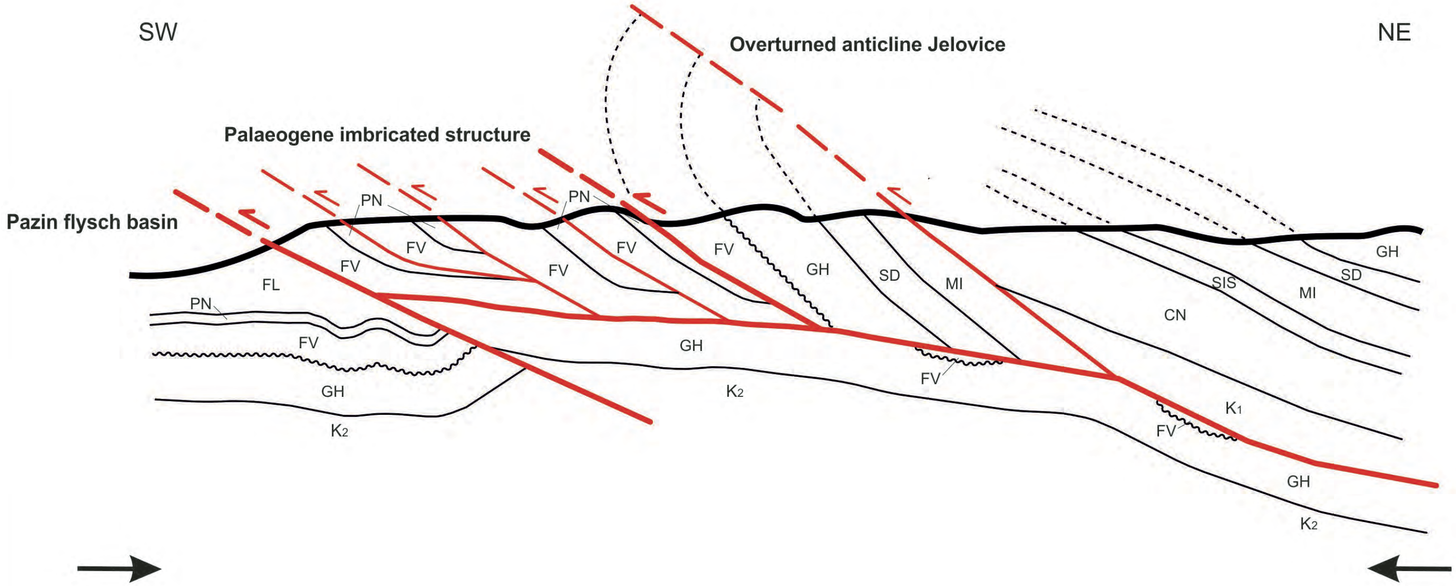

Fig. 17. Schematic presentation of the thrust sheet of the Cretaceous deposits tectonically transported towards the SW and the formation of the Palaeogene imbricate structure of the SW part of the Ćićarija Mt. (PALENIK, 2020; not to scale).. Legend: K_1 – Lower Cretaceous deposits; K_2 – Upper Cretaceous deposits; CN – Crna fm. (Albian); SIS – Sis fm. (Upper Albian–Lower Cenomanian); MI – Milna fm. (Middle–Upper Cenomanian); SD – Sv. Duh fm. (Upper Cenomanian–Lower Turonian); GH – Gornji Humac fm. (Upper Turonian–Coniacian); FV – Foraminiferal limestones fm. (Lower–Middle Eocene); PN – Transitional deposits fm. (Middle Eocene); Flysch deposits (Middle–Upper Eocene). The arrows at the bottom of the picture indicate the NE–SW oriented compressional stress.

Fig. 18. Thrust fault of the Upper Cretaceous deposits of the lithostratigraphic unit Sv. Duh (SD) on top of the Palaeogene foraminiferal limestones. Location: 45°25'6.7''N, 14°7'3.4''E.

Fig. 19. The frontal part of the imbricate structure composed of the Palaeogene foraminiferal limestones, which represents the top of the thrusted overturned fold. The thrust contact of foraminiferal limestones over the flysch deposits of the Pazin Flysch Basin. View from the SE to the NW. Location: 45°24'24.2''N, 14°0'22.2''E.

measurements of dip directions and dip angles of fault planes and orientation of carbonate slickensides, as well as their sense of movement were used. Based on kinematic criteria, the structural data were separated into compatible datasets and processed by appropriate softwares. By using the P–T axis method theoretical maximum (σ_1), intermediate (σ_2) and minimum stress axes (σ_3) have been calculated. Additionally, using the Right Dihedra Method, synthetic focal mechanisms for the analysed fault segments, i.e., pal-

Fig. 20. Example of a very gently inclined thrust contact of Palaeogene foraminiferal limestones on top of the marl deposits. Location: 45°23'21.3''N, 14°5'36.3''E.

aeo-synthetic focal mechanisms as representations of the palaeostress fields were determined. On the basis of the structural data measured in the field and the results of kinematic analysis two major tectonic cycles were defined in the study area. The older cycle, referred to as the Palaeogene Tectonic Cycle, is characterized by compressional palaeostress field with the P-axis dominantly trending NE–SW. This tectonic cycle formed structures of the NW–SE strike, i.e., the Dinaric strike. The activity of the younger, Neotectonic Tectonic Cycle in the study area began probably during the Late Miocene and/or Pliocene and continues to the present time. This tectonic cycle is characterized by compressional/transpressional stress field with N(NE)–S(SW) oriented P-axis. Neotectonic activity resulted in newly formed structures, striking N(NE)–S(SW), with sporadic deviations to the NW–SE strike, and structural reactivation of the older Dinaridic strike faults resulting in a dextral/sinistral movements. These results are correlative with the description of both tectonic cycles that were recognized in the wider area of the NW Dinarides (PALENIK, 2020).

STOP 4 – FANTASIA QUARRY: THE FANTAZIJA QUARRY GEOLOGICAL NATURAL MONUMENT

Ladislav Fuček, Igor Vlahović, Dubravko Matičec, Damir Palenik, Bojan Matoš, Darko Matešić, Aleksandar Mezga

Introduction

The Fantasia Quarry (*Cava di Monfiorenzo*) is located in the eastern part of the city of Rovinj, along the road towards Bale (Fig. 21). The Fantasia Quarry is an integral part of the recently opened Fantasia Visitor Center, accomplished by the Public Institution Natura Histrica with partners that include City of Rovinj and the Tourist bord of the city of Rovinj (EU funded project "geoIST3A – valorization of natural heritage in the County of Istria by improving the competitiveness of the tourist offer"). Due to the specific alternation of dark gray and light gray mac-

Fig. 21. The position of the Fantasia quarry within the Visitor Center "The Fantasia Quarry" – yellow arrow. Source: *Google Earth Pro*.

ro- and microcrystalline dolomites with numerous decorative and effective structural-textural features, there was an interest in exploiting Fantasia dolomites as an architectural-building stone. However, these dolomites proved to be an extremely hard material that delaminated along the sharp contacts of early-diagenetic and late-diagenetic dolomite, so this idea was soon abandoned. In the abandoned quarry, the western, northern and eastern fronts remained vertically sawed and polished, which enabled detailed geological, especially sedimentological research of different types of dolomites with numerous well-preserved structural and textural features (TIŠLJAR, 1976). In 1987, the Fantasia quarry (*Cava di Monfiorenzo*) was declared a protected geological monument of nature, for which late professor Josip Tišljar, fellow of the Croatian Academy of Sciences and Arts, was certainly the most deserving person.

Description of dolomites in the Fantasia quarry

A total of 13 layers of dolomite are visible in the Fantasia quarry, which are only one part of the sequence of deposits of the informal lithostratigraphic unit defined as the Rovinj fm. (VLAHOVIĆ, 1999; MATIČEC et al., 2020). On the surface, the outcrops of the Rovinj fm. are found in a continuous belt that stretches along the Western Istrian Anticline from Poreč, across the Lim channel, Žbandaj and all the way to the coast south of Rovinj. In a normal superposition sequence, the dolomites of the Rovinj fm. are located between the Zlatni rt fm. (ZR) below and the Materada fm. (MA) above (Fig. 22). The total thickness of the Rovinj fm. is very variable, in average around 35 m. The Berriasian age of the dolomite deposits of the Rovinj fm. was confirmed by the occurrence of green algae *Humiella sardiniensis* (OTT & FLAVI-

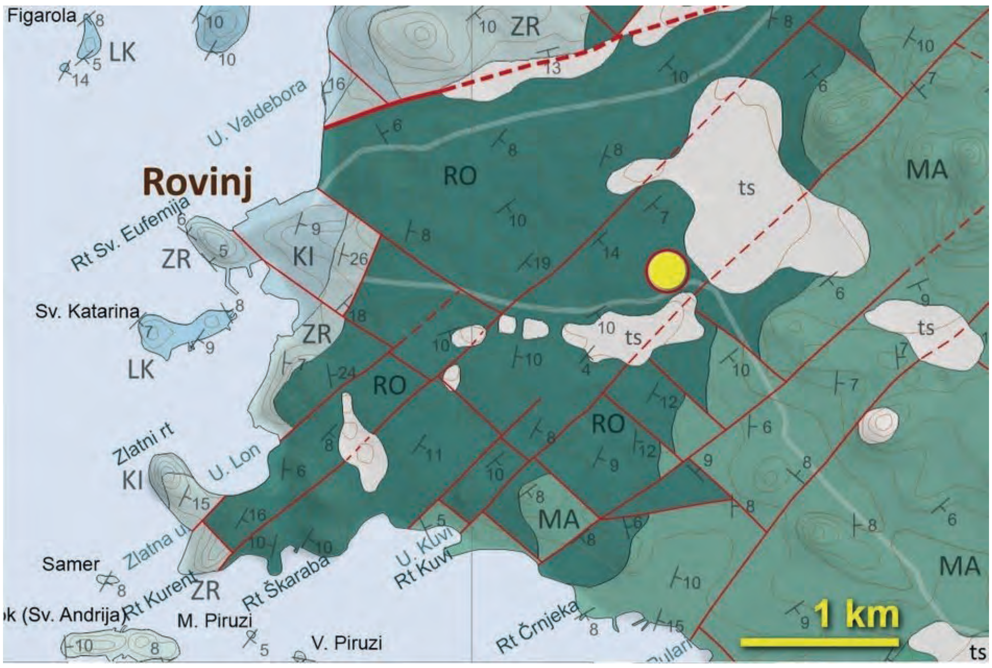

Fig. 22. A detail of the Geological Map 1:50,000 of the Rovinj area, the yellow circle marks the position of the Fantasia Quarry (MATIČEC et al., 2020)

Fig. 23. Eastern front of the Fantasia quarry before the construction of the Visitor Center with well visible alternation of light gray layers of early diagenetic and dark gray layers of late diagenetic dolomites

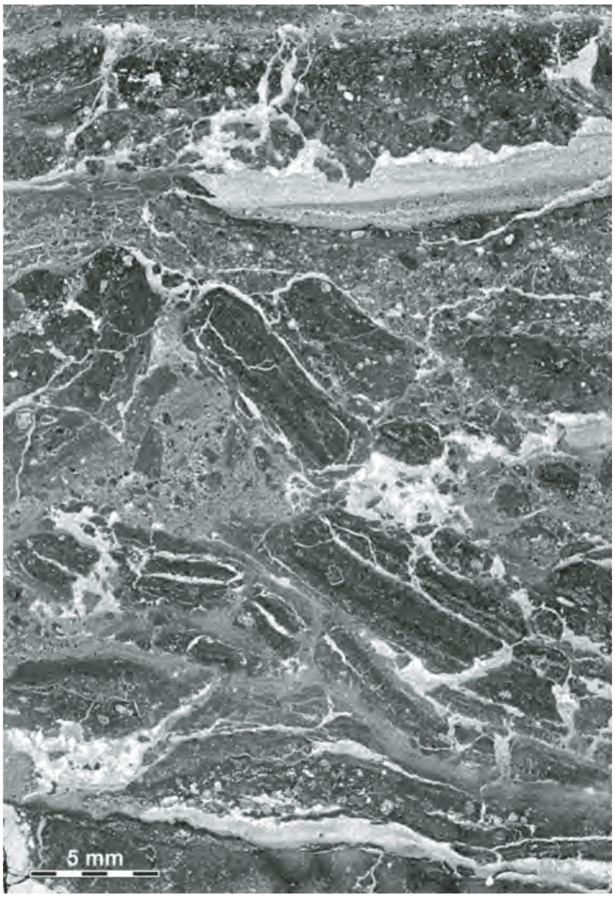

Fig. 24. Laminated dolostromatolite within a light-coloured microcrystalline dolomite

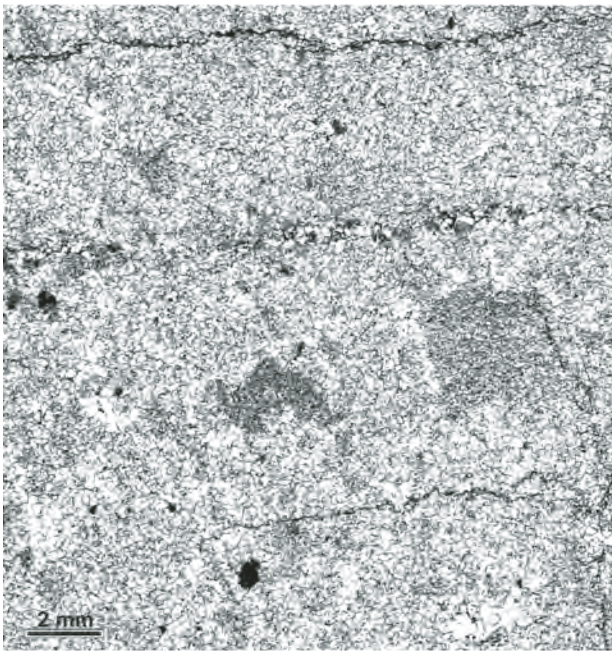

Fig. 25. The late diagenetic dolomite with a coarse crystalline structure composed of large hypidiomorphic dolomite crystals with almost completely destroyed primary structural and textural features

ANI) and *Clypeina radici* SOKAČ, small benthic foraminifera, gastropods similar to certain species of the genus *Nerinea*, and relatively frequent pellets from the *Favreina* group, but also based on the superpositional relationships of the Zlatni rt fm. below and the Materada fm. above.

The alternation of light gray and dark gray layers of dolomites is the macroscopically easily observable feature of the Rovinj fm. (also known as Fantasia dolomite after this locality; Fig. 23). The thickness of individual layers is variable, between 15 and 170 cm. The light gray to almost white layers represent synsedimentary, early diagenetic or primary dolomites of cryptocrystalline structure formed relatively quickly after the deposition and containing numerous preserved and visible structures as well as fossil content (Fig. 24).

The late diagenetic or "secondary" dolomites are represented with the dark gray layers formed subsequently, by post-sedimentary dolomitization. It is characterized by coarse crystalline structure and lack of visible textures and fossil content (Fig. 25). Bedding surfaces between different types of dolomites are mostly very uneven, but sharp.

Among the structural–textural features, especially in the light gray early diagenetic dolomites, the most common is a laminated texture as a reflection of the succession of flat, slightly wavy or curly cyanobacterial laminae–interlayers of varying thickness, and fenestral structures with irregular and laminoidally arranged fenestrae. In the light gray dolomite of the layer 3 there are uneven mottled intervals that could be the result of bioturbation, and in the top part of the layer there are three horizons of desiccation cracks with rare tepee-structures (Fig. 26). On the western wall of the quarry, various examples of the load-cast structures formed by intrusion of consolidated parts of the denser upper layer into the underlying semi-consolidated hydroplastic deposits are clearly visible. Good example is the intrusion of consolidated layer 5 into the semi-consolidated layer 4 (Fig. 26), as well as several such examples on the northern wall of the quarry. At the top of the layer 5 (on the western wall of the quarry) there are two examples of the eroded upper part of the layer 5 and channel-like filling with coarser-grained material from the layer 6. Besides, the layers 5 and 6 are marked by numerous shrinkage cracks, which are mostly filled with sediment from the upper layer.

Layer 8 contains numerous deformations of unconsolidated sediment (soft-sediment deformation), such as plastic deformations of sediment in the form of slump-folds, vertical extrusion water escape structures, i.e. escape of water in the flower-like deformed interlayers (Fig. 27), flame structures, diapir-like uplifts, injected sediment and numerous other examples of unconsolidated sediment deformations (including tearing, brecciation and deformation of semi-consolidated parts of the sediment). All of these structures may indicate seismites that were probably formed on a relatively gently tilted local slope inclined towards the south due to earthquakes.

Another type of structures that can be observed are the structures created within the consolidated, brittle rocks (brittle deformation structures). These are small-scale reverse faults found in layers 10 and 11. There are several of them, the largest one has a movement of about 5 cm (Fig. 28), the smaller ones are characterized by a movement of about 1 cm, but all of them indicate southward tectonic transport.

Interpretation of structures and textures in the Fantasia quarry dolomites

Based on the obtained results of research on structures, especially in the light gray early diagenetic dolomites or "primary dolomites", it can be concluded that the depos-

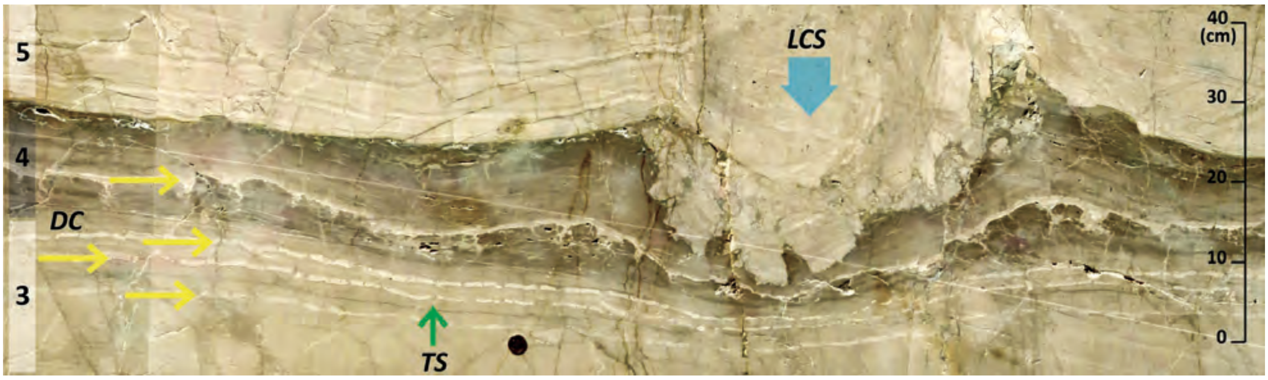

Fig. 26. In the top part of layer 3 there are three horizons of desiccation cracks – DC (yellow arrows) and rare tepee structures – TS (green arrow). LCS – pressing of consolidated sediment from layer 5 into semi-consolidated layer 4 (load-cast structure; blue arrow). Height of the outcrop c. 50 cm.

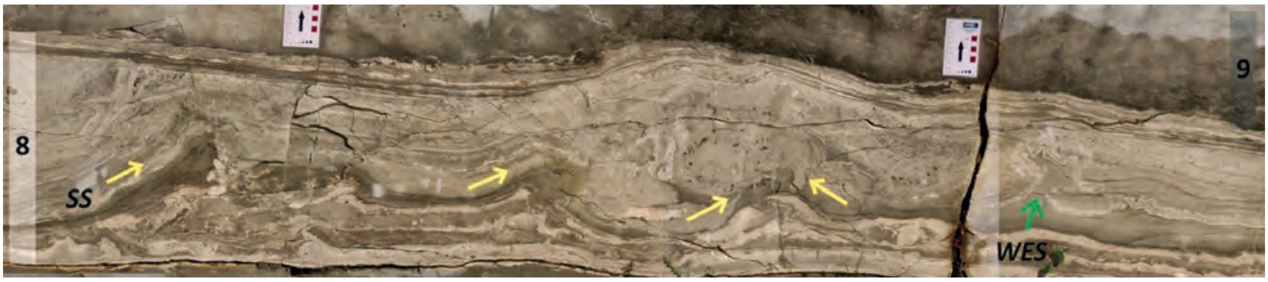

Fig. 27. Layer 8 with numerous soft-sediment deformation – SS plastic deformation slump structures – yellow arrows, WES water escape structures – green arrow. Height of the outcrop c. 70 cm.

Fig. 28. Brittle deformation structure – small-scale reverse fault cutting through the layers 10, 11 and 12 – red arrow. Height of the outcrop c. 80 cm.

its were formed in a very shallow marine environments where the effects of increased water energy under the influence of stronger waves or storms were occasionally recorded. In addition, numerous evidences for the synsedimentary deformation are visible in the rocks, partly in a soft, unconsolidated deposits, and partly in a partially or almost completely consolidated deposits.

Most of the structures created by the deformation of soft sediments are formed in water-saturated non-cohesive sediments in which liquefaction as a mechanism of deformation occurs in situ due to the disruption of the system of mutual support between the grains caused by the increase in the pore fluids pressure (ALLEN, 1982).

There are several types of structures indicating deformations in uncosolidated sediments (soft sediment deformation structures) – from load cast structures, diapir-like and dyke-like structures, plastic deformation slump-like structures, flame structures, etc.

Particularly noteworthy are the large load cast structures that are visible in layers 4, 5 and 6 on the western and northern quarry walls. They indicate the embossing of the consolidated sediment from above into the uncosolidated soft sediment below. This resulted in the displacement of soft sediment from the sides and the creation of diapir-like structures as well as sedimentary dykes that in places pass through thin cracks into the overlying sediment. Load cast structures are formed at the contact of two lithotypes of different densities and/or degrees of sediment connectivity, where the consolidated overlying deposits are pressed into the softer, unconsolidated underlying sediment. They are the result of the gravity force, so they have a pronounced vertical component, and as the cause is often proposed seismic activity that causes liquefaction of the sediment. Such assumed synsedimentary tectonic activity was already recognized in the evolution of this part of the former Adriatic Carbonate Platform (MATIČEC et al., 1996; TIŠLJAR et al., 1998; VLAHOVIĆ et al., 2005, 2011).

An alternative interpretation of the origin of structures in the Fantasia Quarry shown in Fig. 26 was given by LOCKLEY et al. (1994b), based exclusively on photos from TIŠLJAR et al. (1983). These authors assumed that these could be dinoturbations by sauropod dinosaurs. However, clear evidences for such interpretation are missing.

Another type of synsedimentary structures visible in the Fantasia quarry are deformation structures formed in almost or completely consolidated deposits – brittle deformation structures. Among them, the most important are the small-scale reverse faults cutting through layers 10 and 11 with dip of c. 50°. There are several such faults, with the largest and best visible fault having a displacement of about 5 cm, and the smaller ones having a displacement of approximately 1 cm. It is important to emphasize that the tectonic transport of these reverse faults is southward, which corresponds to the direction indicated by the aforementioned deformations soft sediment deformations.

Considering that almost all the structures in the layers 8, 10 and 11 indicate movement towards the south, it can be concluded that the entire block was probably gently inclined towards the south, which is in accordance with the position in the SE limb of the gentle Western Istrian Anticline.

Based on the observation of a limited number of outcrops, it can be assumed that these were very likely very gentle slopes, in the range of 1–3°. Namely, earthquakes with a magnitude greater than 5.5 on the Richter scale can cause slumping even on slopes of only 0.25° inclination, and an approximate earthquake magnitude of 5 on the Richter scale can be assumed as the limit below which sediment liquefaction occurs, and above which folding or faulting occurs under ductile–brittle conditions (CHAKRABORTY et al., 2019).

Significant influence of synsedimentary tectonics in Istria is also indicated by the occurrence of Palaeogene foraminiferal limestones on top of very different levels of Cretaceous deposits, oldest among them being just 10–12 My younger than deposits in the Fantasia quarry (MATIČEC et al., 1996), as well as a very clear tectonic activity recorded in the oldest Late Cretaceous deposits (TIŠLJAR et al., 1998; VLAHOVIĆ et al., 2005, 2011). This is evidence of the gradual uplifting of the anticlinal area of Istria much earlier than tectonic deformation in other parts of the Adriatic Carbonate Platform.

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