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**The Structural Evolution of the Albula Pass Region,
Graubünden, Eastern Switzerland: The Origin of the Various
Vergences in the Structure of The Alps**

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1 THE STRUCTURAL EVOLUTION OF THE ALBULA PASS REGION,
2 GRAUBÜNDEN, EASTERN SWITZERLAND: THE ORIGIN OF THE
3 VARIOUS VERGENCES IN THE STRUCTURE OF THE ALPS

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13 *"Die Natur erreichen Sie doch nicht"*

14
15 Viktor Uhlig

16
17 **ABSTRACT**

18 The Albula Pass region lies between the Lower Austroalpine Err Nappe and the
19 Middle Austroalpine Silvretta Nappe. They will be treated here as the frame of
20 the non- to gently metamorphic sedimentary units between the two during the
21 Alpine times. Sedimentation started on a metamorphic Hercynian basement
22 during the latest Carboniferous (?) and continued into the Permian. Then a
23 sequence from the Alpine Buntsandstein to the medial Jurassic to early
24 Cretaceous Aptychenkalk (=Maiolica) and Radiolarites were deposited in an
25 environment of rifting and subsidence. The succeeding Palombini clastics were
26 laid down after the Aptychenkalk and mark the onset of shortening in the Alpine
27 realm. The initial structures that formed were at least two north-dipping normal
28 faults which formed before the deposition of the Jurassic sedimentary rocks.
29 When shortening set in, the first structure that came into being was the south-
30 vergent Elalbula Nappe, bending the normal faults into close antiforms. It

became further dismembered into two pieces creating parts of the future Ela and the Albula Nappes in the Albula region. This motion was later reversed, when the entire ensemble became bulldozed by the immense body of the Silvretta Nappe along numerous, closely spaced other thrust faults, some of which only very locally followed horizontal bits of the old normal faults, but in principle they determined their own course. No evidence for westerly motion could be identified, although microstructures in the structural fabric were not studied. The reason for this may be the pre-orogenic fabric in the bounding tectonic units.

Key words: Albula Pass, Graubünden, Switzerland, the Alps, Alpine Neo-Tethys, vergence change in orogens

INTRODUCTION: VERGENCE IN OROGENS

This paper deals with a very local, yet important, problem concerning the early vergence development in the Lower Austroalpine units in the Canton of Graubünden, eastern Switzerland, using my observations in the Albula Pass region as a basis. I mapped a small area there between 1979 and 1981. Although a new map appeared in 1987 by Bearth et al. and the explanatory text in 2015 by Furrer et al. I have not found a reason to change my original map largely because the 1987 map is based on very old observations and because I have not made use of it while mapping myself. Where there are serious disagreements, I comment

on them below. The 1987 map has many fewer structural observations, which is the strength of my map. As will be obvious below I disagree with the basic structural interpretations of Bearth et al. (1987) and Furrer et al. (2015) at the Albula Pass proper on the basis of my structural observations.

Most orogenic belts in the world have one dominant vergence, i.e., direction of fold overturning and transport direction of the upper plates of thrust faults (Stille 1930, p. 380 and fig. 1), recognised since the days of the Rogers brothers in the Appalachians (Rogers and Rogers 1843) and Eduard Suess' epoch-making little book *Die Entstehung der Alpen* (Suess 1875; see also •engör 2009). After the rise of plate tectonics it became obvious that an orogen is the product primarily of subduction and that the dip of subduction is the main determinant of orogenic vergence. Collisional orogens are constructed atop subduction-controlled orogens and they inherit their vergence. However, the origin of vergence is not so simple as these sentences express. When studied in detail, one encounters changes of vergence in time even in the smallest orogens created by the elimination of small oceans (barely a 1000 km across: see •engör 1991) or even by restraining bends on major keirogens, such as the Transverse Ranges in California (Davis et al. 1986, 1987; Namson and Davis 1988; •engör 1990) or the mountains in Svalbard (Lowell 1972). In major orogenic systems this may be a result of flip of subduction direction (e.g., in Irian Jaya and Papua/New Guinea:

Dow and Sukanto 1984; Cloos et al. 2005), but in smaller mountain belts the switching, or even the origin, of vergence has less obvious reasons.

One such small orogen is the Alps. Without doubt the best-studied orogenic belt in the world, it has long served as the showpiece of asymmetric orogens (Suess 1875; Argand 1911, 1912, 1916; so memorably summed up by Sir Edward Bailey: 'the principle that has been most fruitful has been the *one-way rule for mountain traffic*' : Bailey 1930, p. 25, italics his) until Kober (1911, 1914a, 1921, 1923) declared them to be symmetric with the south vergence of the Southern Alps forming a southern wing as opposed to the northern wing formed by the main body of the orogen. The two were supposedly joined by a scar (*Narbe*: Kober 1921, p. 140; Kober took this word from Suess, first introduced by Suess to designate volcanic lineaments: 1883, p. 213; then 1901, p. 422ff, in part in the sense that Kober later used it: see also Salomon-Calvi 1930, pp. 19-20) in the Alps, the Insubric Line, whereas elsewhere (e.g. in the Pannonian Basin, in Anatolia, in the North American Cordillera) older massifs called "betwixt mountains" (*Zwischengebirge*, Suess 1909a, p. 435; in the English translation: Suess 1909b, p. 380, there translated by Hertha Sollas as "intermediate range"; Kober 1914b; best Englished by Collet 1935, p. 24, as "betwixt mountains") are supposed to intervene between the two wings. The majority opinion followed Kober in the twentieth century (e.g., Stille 1924; Stille changed Kober's *Narbe* to *Scheitelung*, i.e. "vertex"), but after the advent of plate tectonics it was seen that Suess had been

right all along and that the southerly vergence of the Southern Alps was nothing more than back-thrusting in a north-vergent orogen, as Argand had shown in detail already in 1911; also see Argand 1924, fig. 19^{ter}).

However, the existence of *early* south vergence within the Alps soon began to be suspected. Some saw in it evidence for an early north-dipping subduction (e. g., Hsü and Schlanger 1971) others held gravity-gliding of nappes responsible (Tollmann 1981 and the references therein). However, these suggestions remained isolated and were bereft of detailed structural geological support. The purpose of the present paper is to present stratigraphic, but mainly structural evidence for an early south vergence in the Alps, its later flip to a north vergence and speculate briefly about their causes. The speculation I present applies only to the small area under discussion and does not imply a negation of other interpretations elsewhere in Graubünden or in the Alps in general. To generalise my results to the entire Lower Austroalpine units or even to those in Graubünden would be inadvisable without detailed studies on rotations of nappe fragments about vertical axes during the entire Alpine evolution. The Albula Pass in the eastern Swiss canton of Graubünden (=Grischun, Grisons) was chosen, because there a large number of structural elements come together in a very small region allowing a detailed structural analysis (Fig. 1). The area is small, not mapped previously from a structural viewpoint, hence the greater detail given. I give the rock descriptions also in some detail, because the last

descriptions date from Heierli's mapping published in 1955 (Heierli 1955), although the explanatory text of the Albula sheet carries the date of 2015 (Furrer et al. 2015).

GEOLOGY OF THE ALBULA PASS REGION

Location. The area investigated reaches in an east-west direction roughly from the culmination point of the Albula Pass (*Pass d'Alvra*: 2372 m; in the Swiss topographical grid 783 longitude) to about the latitude of the Gualdauna Hill (Swiss topographical grid 789 longitude). In a north-south direction it extends from the southern slopes of the Piz Kesch (Fuorcla Pischa: 2871 m a. s. l. and Vschinauncha da Zuoz: about 2700 m; about 165 latitude) to the northern slopes of the Crasta Mora including heights of some 2700 m (about 161 latitude; Figs. 1 and 2). This area is comprehended within the topographic map 1237 of the Swiss National Map series 1/25,000, entitled "Albulapass". The same area is covered geologically by Bearth et al. (1987) with an explanatory text by Furrer et al. (2015), to both of which reference is made frequently below. The entire area is above the tree limit and is covered with grasses and, higher up, mosses before bare rock is reached. There are no permanent human settlements in the pass. In the Summer, the *Ospiz* in the culmination point of the Pass serves the tourists and a cattle shed for the grazing cattle that come to the pass during the Summer is located in Alp Nova (see Fig. 2). The Es-cha hut (*Chamanna d'Es-cha*) run by the

140 Swiss Alpine Club lies at the foot of the Piz Kesch at an elevation of 2594 m a. s. l.
141 providing welcome shelter for the mapping geologist. In this paper whenever I
142 refer to my own figures I write Fig. with a capital F; when I refer to figures in
143 other publications, I write fig. with a lower case f. Since I did not formalise my
144 stratigraphic units, their names are spelled with lower case letters. Rock-
145 stratigraphic terms are capitalised (Lower, Middle, Upper), but time-
146 stratigraphic units are not (early, medial, late).

147

148

149 *Short history of interpretation and statement of the problem.* The area was first
150 mapped by Escher and Studer (1839) who identified the sedimentary and
151 metasedimentary rocks in the Albula area as "*Dolomit, Rauhwacke and Flysch,*
152 *Bündnerschiefer and Macigno*" (see their coloured foldout geological map on their
153 unnumbered plate II) and recognised the overall steepness of the structure with
154 bivergence towards the Crasta Mora in the south and towards the Piz Kesch in
155 the north. In the next important study of the Albula Pass, G. Theobald (1863)
156 recognised the basic internal structure of what is now recognised as the Ela
157 Nappe as a south-overtaken anticline, but was less successful in the Albula Pass
158 floor itself, what is called the "Albula steep zone" in Furrer et al.'s explanatory
159 text (Theobald 1863, p. 213 and cross-section 52). Yet Theobald did recognise the
160 main rock types we now know to be in the Albula area, although he did not in all
161 cases identify their correct stratigraphic position. Yet his naming of the

162 stratigraphic units later allowed a quick testing and rapid improvement.
163 Theobald's 1876 map, revised and added to by J. M. Ziegler, was a still more
164 important improvement in that the wide distribution of Liassic sedimentary
165 rocks in the Albula and the Val d'Es-cha was recognised. Karl Zoeppritz's
166 dissertation, completed under Gustav Steinmann, brought a great improvement
167 to the stratigraphy and the structural understanding of the Albula area, although
168 no nappes were as yet recognised (Zoeppritz 1906). That improvement came
169 with Zyndel (1912), who already knew of the main nappes in the area, including
170 the Ela and the Albula and he certainly recognised the great structural
171 complexity of the internal architecture of these metagranitic/metavolcanic
172 metasedimentary and sedimentary nappes. Parts of Eugster (1924) and Cornelius
173 (1935, 1950) have treated the geology of the westernmost and the southern parts
174 of the area I mapped. Eugster added little and in fact was really a step backwards
175 from Zyndel (1912) whereas Cornelius (1935, 1950) was a great improvement on
176 anything that had come before, but specifically in the Albula area he replaced
177 Zyndel's (1912) Albula Nappe with the *Albula Lappen* (=Albula slivers or rags).
178 Heierli's mapping (Heierli, 1955) improved still more the stratigraphy, but was a
179 serious step backwards in terms of structural interpretation, having been
180 shoehorned into a Staub-style multi-slice interpretation encouraged in part by
181 Cornelius' concept of the *Albula Lappen*, but mainly by his doctoral advisor
182 Rudolf Staub (R. Trümpy, pers. comm., 1979), which is unfortunately still
183 reflected in the interpretation by Furrer et al. (2015). Furrer et al. (2015)

184 summarise all that is discovered to date on the geology of the Albula Pass area.
185 The present paper is the first detailed mapping after the authors I just cited
186 concerning the Albula Pass area and its sole purpose was to understand the
187 structural geology of the region with a view to clarifying the palaeogeographic
188 position of the Ela Nappe. Its results turned out to be more significant than
189 originally hoped for. For a quick orientation in the tectonics of the Alps for the
190 extra-Alpine reader I would recommend two recent books: Weissert and Stössel
191 (2010) and Pfiffner (2014) and an excellent recent paper: Schuster (2015).

192

193 The geology of the region described in this paper consists of four major tectonic
194 units as seen in Figs. 1 and 2. From north to south, these are the Silvretta Nappe
195 (represented by the Piz Kesh, classified as one of the ultra-prominent peaks in
196 the world, i.e., above 1500 m prominence), the Ela Nappe, the Albula Nappe and
197 the Err Nappe represented by the Crasta Mora. I do not here indicate the
198 ascription of these to the larger Alpine entities as was done in Furrer et al.,
199 (2015), because of the latter's in part outdated classification going back to Rudolf
200 Staub's ideas and because some of the newer interpretations are less detailed
201 than mine. There are narrow belts of tectonic slices between my units consisting
202 of the Sub-Silvrettide lenses between the Silvretta and the Ela, the Infra-Ela slices
203 (herein defined for the first time) between the Ela and the Albula, and the Supra-
204 Err Slices (herein defined for the first time) between the Albula and the Err. I did
205 not look into the detailed petrography and structure of the Silvretta quartzo-

feldspathic gneisses or that of the Err Granite. I only wish to note that the lowermost contact of the Silvretta at the foot of Piz Kesch consists of a metre or so of greenish-grey mylonites in the sense of Lapworth (1883, p. 121; 1885, pp. 558-559), i.e., milled, banded, porcellaneous rocks.

In the following sections, I first describe the stratigraphy and then the structural geology. I end with a tectonic interpretation of the reasons of vergence switch in this part of the Alps, which I believe to have general implications for most small orogens.

Stratigraphy. The method I adopt for describing the stratigraphy is that I recount the stratigraphy of each independent tectonic unit (nappes and slices) as defined herein to give the reader an idea of its environment of origin and evolution (see Figs. 1, 2 and 3). Such a presentation, however, could give the mistaken impression that the tectonic units had already been known before their stratigraphy was worked out. Despite the fact that I am the fifth generation to map in the Albula Pass and the second to do so on a scale of 1/10,000, I had to redefine some of the previously named units and erect new ones. My units also differ from those by Furrer et al. (2015) who were constrained to use older observations and classification to bring their explanatory text with an older map. Establishment of stratigraphy and erection of tectonic units in such complexly-deformed areas as the Albula Pass region go hand-in-hand and simultaneously

in an iterative fashion. While mapping in such terrains one constantly employs the method of trial-and-error and that is why it is absolutely critical to publish the geological map in the form of an outcrop map. People disagreeing with my choice of boundaries may alter them as they see fit, which is only possible if there is an outcrop map at hand. Large-scale geological maps are of little value, if they do not show the outcrops.

I give no justification as to why I chose the stratigraphical and structural unit boundaries as I did, because such falls out of the descriptions and the geological map (Fig. 2). The stratigraphical descriptions are here kept as briefly as is consistent with the purpose of familiarising the reader with the environment of origin of the rocks. I erect no formal stratigraphic units, because the mapped area is so small. Other workers may take my units and incorporate them into theirs as they see fit. Identifiable index fossils in the Albula Pass region are extremely rare. I mention only those fossils found in units first established in this paper. For the previous stratigraphic terminology used in the Albula Pass region, see Heierli (1955), Trümpy and Haccard (1969), Eberli (1985a, 1985b, 1987, 1988), Bearth et al. (1987) and Furrer et al. (2015).

The Err basement and the Supra-Err Slices

In the field, the Albula Granite of the Err Nappe lies above Middle Triassic (Anisian to Carnian) and Liassic sedimentary rocks of the Supra-Err Slices across a sharp, mostly cataclasised, locally mylonitic, tectonic contact of about 1 m thickness. The brecciated rock consists of numerous elongated components that are mostly of granitic material. It abuts against a well-foliated shaly horizon, a possible fault rock, with a variable thickness averaging some half-a-metre. The shale contains elongated dolomite clasts ranging in size from mm to rare 3-4 cm pieces. Under this is the ± 3 m thick dolomite, the base of which is delimited by a number of closely-spaced, south-dipping thrusts. The contact between the Middle Triassic dolomites and the Liassic cherty limestones ("Kieselkalk") cannot be observed in the field, but it must be tectonic simply because no other rock units intervene except Rauhwanke, a notorious indicator for tectonic contacts.

Within the area shown in Fig. 2, the *Diplopora*-bearing grey to dark grey Middle Triassic dolomites (unit 14) of the Supra-Err slices, identical in character to the Albula Nappe Middle-Triassic dolomites (see below), are seen only in one place above a tiny pond at 2416 m elevation approximately 1 km SE of the *Ospiz*, although Bearth et al. (1987) identify these dolomites as Hauptdolomit. The outcrop consists of massive, grey to dark grey, fine-grained dolomites with local laminae of 1-3 mm thickness. Bedding is hard to make out to get a handle on the structure, but looking at the outcrop at a distance, I obtained a distinct impression that it forms an antiform with an axial plane paralleling the

271 bounding shear surfaces (Fig. 4A). Locally, the dolomite is shale-rich. Just under
272 it, but unfortunately not in direct contact, is a small outcrop of yellowish-orange
273 to light brown Rauhwaacke (for more detail on the Rauhwaacke see below).

274
275 The next unit of the Supra-Err slices is a dominantly brecciated one, consisting,
276 from bottom to top of micaschists, light grey cherty limestones and medium-
277 bedded, grey-brown dolomites.

278
279 The micaschist is a greenish-black, metre-and-a-half sliver located at the core of
280 what seem to be a large-scale fold defined by the dolomites and the cherty
281 limestones, although its facing is uncertain (Fig. 4B and B'). The schist is in clear
282 tectonic contact with the overlying dolomite. Here the dolomite has acquired a
283 phacoidal cleavage. The dolomite itself consists of alternating grey to buff-
284 coloured beds of about 50 cm to 1 m thickness. The grey beds contain numerous
285 neptunian dykes dying downwards and being filled with mm to cm size clasts of
286 the grey dolomite. Extensive stylolites are bedding-parallel indicating
287 considerable shortening of the beds vertically. There are 2-3 m diameter, lensoid
288 lumps of foliated, brown, shale within the dolomites. The total thickness of the
289 dolomites is about 15 to 20 m.

290
291 A blue-grey weathering cherty limestone conglomerate, cemented by grey-buff
292 weathering dolomitic matrix follows upwards the bedded dolomites. Generally,

the whitish-grey to grey chert layers seem confined to the light grey limestone clasts, but at least in one case the chert layer truncates the clast/matrix boundary and is thus younger than the brecciation/cementation. In fact, upwards, the dolomite-cemented breccia completely disappears and a massive, extremely tectonised limestone, with abundant boudined chert layers, dominates the outcrop.

The limestone is dull, blue grey and greatly resembles in outward appearance to the Aptychus Limestone ("Aptychenkalk"= "Maiolica" of the southern Alps and the Apennines), although it never assumes the typical whitish-buff weathering colour of the Aptychenkalk and remains grey on the weathering surface. On the fresh surface it is much darker and appears sparkly. Under the microscope it appears to be a completely recrystallised and chertified limestone, probably originally a micrite.

The outcrops I describe have been described as representing a continuous stratigraphic series in Bearth et al. (1987) and Furrer et al. (2015) with which I cannot agree. There are unmistakable isoclinal folds and numerous shear zones in the outcrops and the sedimentary rocks and the metasediments are in clear tectonic contact with the Err Granite. Their present position clearly indicates that they come from on top the Err Granite and that they are sheared sediments that once sat on top of the Err as both Stöcklin (1949) and Manatschal (1995) maintain.

315

316 Most likely, the cherty limestone represents the Liassic *Kieselkalk* of the Lower
317 Austroalpine Units. Its preserved thickness is about 45 m. The extremely
318 scattered outcrops, strong deformation and the extreme rarity of fossils make a
319 detailed reconstruction of the stratigraphy of the Supra-Err slices exceedingly
320 difficult. From the extant information, the following evolutionary scenario seems
321 the most plausible:

322

323 Wherever they are seen in the Albula region they are always in tectonic contact
324 with the Err Granite. Not only nowhere can an original stratigraphic contact can
325 be observed, but there are no crystalline clasts within the breccia itself. They thus
326 cannot be the transgressive cover of the Err Granite here. Neither can they be
327 directly compared with the Salamun Breccia farther west described by Stöcklin
328 (1949). Because the breccias I just described are made up mainly of dolomitic
329 material and are found directly atop the definite Middle Triassic dolomites and
330 Rauhewacke (albeit with a tectonic contact) and because they are overlain by
331 chertified limestones of possibly Liassic (older?) age, they can be directly
332 correlated with Stöcklin's (1949, pp. 58-60) "Liassic breccias in the middle flank of
333 the Err Nappe." However, the components of the breccia in the Albula area here
334 described are not nearly as variable as those described by Stöcklin. Of the eleven
335 type of components he lists, I have been able to find only three in the Albula
336 outcrops. It seems as if the brecciation became ever more intensive westward,

whereby ever older rocks were involved. Evidence to support this view is found in the neptunian dykes cutting the grey dolomite beds in the Albula area. They seem to have controlled the brecciation of the dolomite. In other words, we here observe a breccia *in statu nascendi*, whereas in Stöcklin's area one sees fully-developed polygenic breccias.

Both to the west and to the east, in a supra-Err position, are a number of outcrops that have a common stratigraphy consisting, from bottom to top Aptychenkalk (=Maiolica of the Southern Alps and the Apennines: for a modern evaluation of the environment of deposition, see Weissert, 1981), Radiolarian cherts and Palombini shales (Fig. 3). Both the Aptychenkalk and the Radiolarian cherts represent the last phase of the growth of the Alpine Neo-Tethys (which was in reality a part of the early Atlantic Ocean together with the Liguride/Tellian/Rif/Betic ocean to the south and west: see Engör 1998, fig. 91a), whereas the Palombini Shales represent the onset of the destruction of it, as explained below, together with an alternative, climate-based, suggestion. Whether the thin metaclastic rocks I ascribe to the Palombini also contain the younger (Aptian-Albian) Val Lavagna Shale representatives I cannot tell. What I wish to note however that in the Northern Apennines, the Val Lavagna shales in places directly overlie ophiolitic pillow lavas, i.e., the oceanic floor, without the Palombini intervening (Zaccarini 2006, fig. 1A).

359 The radiolarian cherts are always associated with the Aptychenkalk in the Albula
360 region and barely have a collective volume of a few m³! That is why I do not
361 show them on my map (Fig. 1) as a separate unit from the Aptychenkalk. In fact
362 they are not even mentioned by Furrer et al. (2015, pp. 20-21) except to claim that
363 Heierli had incorrectly identified "*Aptychenkalk und Radiolarit*" at the crest west of
364 Murtel Trigld (coordinates 777.900/161.250, thus west of my area and west of the
365 area covered by Bearth et al. 1987), which they include in the Saluver Formation.
366 At outcrop, the radiolarian cherts appear as wine-red, green or greenish-grey,
367 massive, homogeneous cherts generally cut by numerous quartz veins. They are
368 thinly-bedded (2 to 5 cm at most) and generally complexly folded. The rock
369 seems thoroughly recrystallised so that I was unable to detect any fossils in it. It
370 passes into the Aptychenkalk through a limestone chert intercalation.

371

372 Together with the Palombini Shale, the Aptychenkalk makes up one of the two
373 main rock types of the Supra-Err slices. On the outcrop it is highly foliated, grey,
374 light grey, bluish dark grey, dull, very fine grained, locally cherty and locally
375 shaly limestone (for its structure, see Fig. 5). In some places chert layers have
376 been highly stretched and boudined giving the impression of lined-up chert
377 nodules. These layers allow one to recognise the F₁ fold structures within the
378 Aptychenkalk sequence (Fig. 5). Also, locally, there are thin (1 to 10 cm),
379 phyllitic-looking creamy green interlayers within the limestone, which only very
380 slowly fizz when treated with 10% HCl. An extreme stylolite development is

observed throughout the rock, delineated by thin films of black clayey material generally parallel with the dominant foliation of the rock, which is a solution cleavage. Although in general the thickness of such stylolitic films is on the order of a fraction of a mm, locally, and especially at the fold hinges, they can be as thick as 3 mm. Apart from the disrupted chert layers, no bedding or any other primary structure can be recognised at outcrop.

Under the microscope, the Aptychenkalk appears as a neomorphised, chertified limestone. Locally, coarsely crystalline sparry calcite masses are preserved in patches amidst the chertified sections. In general, however, the rock is a neomorphised microsparite with sparsely distributed silica nuclei. Before the onset of neomorphism, the rock was most likely a micrite consisting of coccolithic muds, although no coccoliths have been here preserved. No other distinct fossils could be recognised either.

The boundary between the Aptychenkalk and the Palombini shale is transitional. Towards the top, the Aptychenkalk becomes increasingly richer in shale and its foliation becomes finer and more pronounced. Because the Palombini shales still contain a considerable amount of limestone, the choice of stratigraphic boundary becomes arbitrary. I define the top of the Aptychenkalk in the Albula region as the level where the foliation becomes fine (2-3 mm), foliation surfaces shiny and the rock colour changes to dark, shiny grey to brown.

403

404 Because of the extreme tectonism (see Fig. 5) it is meaningless to speak of a
405 stratigraphic thickness of the Aptychenkalk in the Albula Pass region. Its present
406 structural thickness changes from a few metres to at most ten metres.

407

408 In the various sedimentary sub-units of the Err Nappe *s. l.*, the Aptychenkalk is
409 followed by limestones, shales with some calcarenitic layers. Nold (1972)
410 provisionally designated this rock assemblage as the *Emmat-Formation*, although
411 no type locality was given. Helmut J. Weissert (pers. comm., 2015) says that at
412 Piz Emmat this assemblage consists of limestones with flint and dark, silicic
413 shales. His attempts at dating it at Piz Nair and in Alp Flix, even with C isotopes,
414 was unsuccessful, but he endorses the idea that it probably spans a time from the
415 Valanginian to Albian. In the Albula area, similar rocks are found
416 stratigraphically above the Aptychenkalk. I informally call them the Palombini
417 shales upon the suggestion of Daniel Bernoulli (pers. comm. 2015), who was in
418 the field with me twice. In the Northern Calcareous Alps (Upper Austroalpine)
419 there are no direct equivalents of the Palombini and the Aptychenkalk there
420 becomes as young as Barremian-Lower Aptian (Helmut J. Weissert pers. comm.,
421 2015)

422

423 The Palombini shales in the Albula region actually consist of a number of
424 interlayered rock types in which shales and limestones make up some 80% of the

rock. The rest consists of fine-grained, iron- (?limonite) stained quartz sandstones. Thus with the onset of the Palombini sedimentation there occurred a gradual increase in the clastic influx into the Alpine Neo-Tethys eroded from lands containing quartz. There can be little doubt that such lands were the initial appearance of the Austroalpine crystalline units. Daniel Bernoulli (written communication, 24th April 2016) pointed out that the quartz influx may be a climatic signal, although the most recent global sea-level curve (see Bilal Haq in •engör 2015, Fig. 15) indicates that the early Cretaceous sea-level lowering was very gentle and still about 100 metres higher than today's. The sea-level started rising again beginning with the Valanginian, but shale deposition continued in the Val Lavagna Shales. Moreover the temporal agreement with other turbiditic clastics in the Northern Calcareous Alps, such as the Roßfeld beds, interpreted as products of trench deposition, makes me think that the Palombini in my area was probably a result of the onset of shortening in the Alpine edifice.

At outcrop, the Palombini is a shiny grey, grey brown and locally even shiny black phyllite. It is penetratively foliated except for the limestone-rich layers. In such places, completely recrystallised grey to black limestones appear as extensive lenses with a thickness variation between 1 mm to at most a cm. The presence of the limestone lenses give the foliation a phacoidal character. Also in lenses are the commonly rusty-brown coloured quartz sandstones. Neither

446 bedding nor any other primary structure seem preserved within the Palombini.
447 Its present structural thickness is nowhere more than 25 m.

448

449 The Palombini shales are the youngest rock unit in the Supra-Err slices, in fact in
450 the entire Albula Pass area, save for the Quaternary scree deposits. Although I
451 have been unable to find any fossils in them, just to the south of Albula, in the
452 Samaden area, Finger (1978) dated them as early Cretaceous.

453

454 Along a discontinuous Rauhewacke horizon, the Palombini Shales abut against
455 the base of the Ela Nappe in the north in the western part of my mapping area
456 and against the various rock types of the Albula Nappe elsewhere (Fig. 2).

457

458 *Albula Nappe*

459

460 The rock types within the Albula Nappe (Zyndel 1912) consist dominantly of
461 metaplutonic and metavolcanic rocks, metaclastics and carbonates representing a
462 time interval from the ?latest Carboniferous to the medial Triassic. This unit is
463 comprehended within the Mezzaun-Schuppe of Furrer et al. (2015, Fig. 12). For
464 mapping purposes only five units have been distinguished. These are, from
465 bottom to top, as follows: 1) Proliebas Assemblage (Permo-Carboniferous?), 2)
466 Funtaunas Assemblage (Buntsandstein), 3) lower Rauhewacke, 4) Middle Triassic
467 dolomites, 5) upper Rauhewacke. The first two of the units are called

"assemblages" simply because they are extremely heterogeneous from a lithological viewpoint, in which a number of genetically unrelated units have been lumped together tectonically. Even a mapping at the scale of 1/10,000 under the outcrop conditions in the Albula Pass area does not permit mapping these units individually. However, both assemblages are easily recognisable as distinct units forming a good mapping basis. In the following I describe them as if they were members in a formation, although it has not been possible to show them on the map because of the impossibility of separating them from one another easily without the help of a microscope.

1. Proliebas assemblage: The Proliebas assemblage is defined to be a heterogeneous group of mainly green, grey-green, blackish green and black coloured rock types consisting mainly of meta-plutonic rocks (a cataclastically brecciated granite {alaskite}, possibly equivalent to the Albula-Julier granite: Cornelius 1935), mafic metavolcanics, metamorphic silicic tuff-like rocks that are probably other cataclasites and psammities. Bearth et al. (1987) identified in this assemblage paragneiss, augengneiss, and sericite quartzite (Palaeozoic), sericite and quartzite schist ("Casanna-Schists": ?Carboniferous) and mainly conglomeratic Verrucano. They are shown to constitute a concordant series in Bearth et al. (1987), although in the field their structural relationships cannot be established because their outcrops are separated from one another by the younger soil cover. Their deformation is extreme having led to extensive

490 cataclasis (see also Furrer et al. 2015, p. 21). However, they all clearly lie below
491 the Funtaunas Assemblage. Elsewhere within the Err Nappe similar rocks have
492 been identified as the fault rocks of Jurassic low-angle extensional faults
493 (Manatschal 1995; Manatschal and Nievergelt 1997). I shall discuss this
494 suggestion below in the structural section.

495

496 1a. Cataclastically brecciated alaskite. In the field, this rock type has a massive
497 appearance cut by a large number of quartz veins. It has a dominant foliation
498 and the quartz veins are mostly parallel with it. However, when observed in
499 detail, the quartz veins seem to be folded isoclinally about the foliation. Most of
500 the folds are floating folds and many have sheared hinges. These veins are rusty
501 brown in colour. Another set of quartz veins, at high angles to the foliation, is
502 less deformed and white in colour.

503

504 The deformation has led to considerable grain size reduction by cataclasis and,
505 because of that, the rock, where not traversed by quartz veins, has the
506 characteristic massive look making field identification difficult. Locally,
507 however, a medium-grained texture is preserved. It is clearly a fault rock, but of
508 what kind cannot be told by studying the outcrops of the Proliebas assemblage
509 along in the Albula area. All one can say that the deformation along it had
510 become more brittle from the time of the injection of the first quartz veins to the
511 time of the injection of the second generation quartz veins cutting the S_1 foliation.

If a judgement only on this observation is to be passed, one would pronounce the cataclasis as resulting from extensional faulting, as overburden became less during the deformation leading to progressively more brittle deformation.

Under the microscope, the main constituents of this rock are seen to be perthitic K-feldspar with widely deformed deformation lamellae and quartz. Plagioclase is rare and there are hardly any mafic minerals. Mainly wart-like myrmekite has developed along the grain boundaries of the K-feldspars, particularly where they abut against plagioclase. The myrmekitic texture here indicates K-metasomatism under intense cataclasis (cf. Simpson and Wintsch 1989). In some of the quartz-rich domains there has been complete recrystallisation resulting from deformation forming typical sugary texture triple points. Locally, quartz crystals have dendritic edges resulting from irregular, anhedral growth.

Elongation of mineral constituents are confined to the shear zones within the rock. Under the microscope it is seen that the rock has been milled down to a fine matrix giving the massive outcrop appearance and giving it away as a fault rock, a protomylonite.

1b. Metamorphic silicic tuff-like rocks. This rock, probably originally a plutonic rock of felsic to intermediate composition, occurs in a small isolated area between the Gualdauna Hill and the Muntische and is found mainly in the form

534 of surface float, which, however, clearly indicates the existence of this rock type
535 as bedrock in the immediate vicinity. In hand specimen the rock is black to
536 greyish black to brown and has a porphyritic appearance with small pieces of
537 quartz, feldspar, and rock fragments embedded in a massive matrix. There is a
538 kind of foliation, which may well be a result of flow. Along it there is a preferred
539 orientation of grains.

540

541 Under the microscope it is seen that the rock consists of a very fine-grained
542 matrix with locally spaced crystals and rock fragments. The matrix is heavily
543 altered and locally totally replaced by a chloritic mass of fine-grained particles
544 showing a well-developed elongation forming cleavage-like weakness planes.
545 Embedded in the matrix are fine particles of mainly quartz and plagioclase
546 crystals, most of which have corroded and rounded edges and corners. They are
547 locally cracked and lost the unity of the original crystal shape. The feldspars have
548 been heavily altered to sericite. Few well-preserved ones show complex twinning
549 according to pericline and carlsbad-albite laws.

550

551 The rock fragments appear to belong to a well-crystallised, possibly deep-seated
552 magmatic body (our alaskite?).

553

554 1c. Metabasite. This rock crops out in Val Bella, about one hundred m directly N
555 of a small pond at the 2375 m elevation mark, along the stream that drains the

pond. It forms a 2 m-thick layer at the base of a highly tectonised section reaching into the lower Rauhewacke. To the south it is in contact with another Rauhewacke through a rusty-brown weathering highly altered rock.

At outcrop, the metabasite is a well-foliated, dark green weathering phyllite or phyllitic schist. About 50 to 60 % of its outcrop consists of quartz veins that weather whitish brown. Its dominant schistosity, which seems to be an S_1 , is finely phacoidal and by far the most prominent structure in the rock.

As a consequence of the deformation and metamorphism, extensive recrystallisation and crystal reorientation seem to have occurred. The protolith looks to have possessed a matrix with crystals and rock fragments embedded in it. These features are still recognisable in recrystallised form. In the matrix, the well-developed schistosity was generated by the preferred orientation of the chlorite. To a lesser extent, albite and quartz crystals also have preferred shapes and orientations. Epidote, sphene and possibly stilpnomelane are present as accessory minerals. The rock fragments have also been recrystallised into quartz- and albite-rich, schistose lenses. The metamorphic grade is within greenschist facies. The protolith may have been a mafic to intermediate lava, but more likely a mafic to intermediate volcanoclastic rock.

A complexly folded, green-white laminated, calcite- and mica-rich rock occurs in outcrops directly south of Alp Nova along the 2200-metre contour line. Under the microscope, the rock looks calcite- and clay-rich with a dominant mylonitic aspect. Both the calcite and the clay are possibly secondary in origin. This rock, too, may have had a mafic volcanic/volcaniclastic parent and probably belongs to the same metabasite group. They are all now highly sheared mylonites or protomylonites.

1d. Psammites and pelites. In the field these rocks occur as olive-green, greyish-green and greenish black phyllites and phyllonites. These rocks were originally sandstones, siltstones and shales.

The phyllites constitute a fairly monotonous group of rocks characterised by a rather narrow range of variation between slaty phyllites and quartz-rich phyllites; although all varieties contain quartz, some are richer in quartz than others. With the increase of mica flakes and foliae, the phyllitic character of the rock increases.

Under the microscope, the components of the fine-grained minerals are not coarse enough to permit a precise determination. But it is clear that the rock is composed mainly of quartz, mica and subordinate amounts of opaque minerals. It is cut by closely-spaced fractures slicing through the minerals forming the

599 dominant S_1 cleavage. One of the most characteristic features of the phyllites and
600 the phyllonites is the development of "eyes" and lenses of very different sizes.
601 The relatively undeformed mineral grains (most frequently quartz), the
602 porphyroclasts, have survived as the nuclei of each lens within a finely
603 granulated matrix. Under increasing strain, the original quartz grains became
604 more rounded or elongated into lensoid shapes. In extreme cases, the varieties of
605 phyllites appear as mixtures of fine-grained microcrystals and mica foliae.
606 Commonly both ends of the quartz porphyroclasts have fibrous appearance
607 while the core retains the texture described previously. The ends merge with the
608 surrounding micaceous material creating a microstructure resembling a
609 "pressure fringe". The metamorphism is within the crystallisation limits of the
610 muscovite, i.e., in the greenschist facies, in agreement with the rest of the
611 Proliebas Assemblage.

612
613 The upper boundary of the Proliebas assemblage against the Funtaunas
614 Assemblage is nearly everywhere tectonic and as it is parallel with the dominant
615 schistosity it is not possible to express any strong opinions whether the Proliebas
616 actually experienced pre-Triassic deformation(s) as Finger (1978) believed
617 possible for the Vairana-Schuppe farther south in the Samaden area, although
618 Staub showed already that its deformation created much mylonitisation and
619 implied that it was entirely Alpine (see Staub 1914, pp. 354-359; also Bearth 1964).
620 The question of the age of deformation of the Proliebas Assemblage has acquired

621 renewed significance in view of the recent discovery of post-Hercynian Permian
622 deformation and metamorphism within the Austroalpine edifice (Thöni 1999;
623 Thöni and Miller 2000, 2009; Schuster et al. 2001; Schuster and Stüwe 2008;
624 •engör 2013) and the interpretation of similar rocks as marking extensional low-
625 angle faults cutting into the crystalline basement of the Err Nappe (Froitzheim
626 and Eberli 1990; Manatschal, 1995; Froitzheim and Manatschal 1996; Manatschal
627 and Nievergelt 1997; Masini 2011, Masini et al. 2011). My present interpretation
628 is that the cataclastic deformation of the alaskite in the Funtaunas assemblage is
629 the same deformation related to Mesozoic extension known from elsewhere in
630 the Err Nappe (see Manatschal 1995, for a detailed discussion; also Froitzheim
631 and Eberli 1990), but the presently-available data alone from the Albula area
632 cannot preclude a Permian extensional event as envisaged by Thöni (1999) Thöni
633 and Miller (2000, 2009) Schuster et al. (2001) and Schuster and Stüwe (2008) that
634 may have been associated with large strike-slip zones (•engör 2013).

635

636 In a number of places, especially in the outcrop group immediately to the west of
637 Alp Nova, there exists what looks like a gradual passage from the psammities
638 and the pelites of the Proliebas Assemblage to the psammities and the
639 metaconglomerates of the Funtaunas Assemblage. Whether this gradual passage
640 is solely a result of the similarity of the rock types juxtaposed across a now-
641 obliterated angular unconformity or whether it really represents a continuous
642 sequence cannot be decided, because the later tectonism and the low-grade

metamorphism have completely obliterated any primary structure that may have existed between the two. In such cases, the boundary was drawn arbitrarily taking colour and grain size as guides: finer-grained green rocks were mapped as parts of the Proliebas Assemblage, whereas coarser-grained reddish brown rocks were considered as belonging to the Funtaunas Assemblage. An identical problem exists in the sedimentary zone of Samaden and its surroundings to the south of the Albula region, where Finger wrote: "Possibly the 'Buntsandstein' occurring in the various units ought to be considered Permian" (Finger 1978, p. 41). A careful isotopic dating campaign of the Proliebas assemblage thus appears to be an urgent desideratum from a number of viewpoints.

2. Funtaunas Assemblage. Much like the underlying Proliebas Assemblage, the Funtaunas Assemblage is also a lithologically rather heterogeneous group of rocks, although on the outcrop it invariably gives the impression of a rather monotonous succession of schists and phyllites. What distinguishes it from the Proliebas is its red, brown and buff colouring and coarser grain size (locally assuming a gneissic texture) compared with the Proliebas Assemblage. Towards the upper parts, the carbonate content of the rock increases visibly and locally it passes directly into dark mouse-grey dolomites. Much like the Proliebas rocks, extreme tectonism and isolation of outcrops do not allow an internal stratigraphy to be erected, for neither bedding nor sedimentary younging indicators are

preserved. The following rock-type groups have been distinguished in the Funtaunas Assemblage:

2a. Metaconglomerate-psammite group. Best exposed in the stream cut between the Punt Granda and Alp Nova, the metaconglomerate psammite group is seen in direct contact with and commonly "grading into" (see above) the underlying Proliebas rocks wherever seen. The younging direction is ascribed on the basis of the entire stratigraphy of the Albula Nappe compared with other Lower Austroalpine units (Finger 1978). This is what has been mapped as Buntsandstein by Bearth et al. (1987). With the reservation noted above, I have no objection to this age ascription.

At outcrop, the rock is a reddish brown, brown, red-green, medium- to coarse-grained quartz- and mica-rich one. It is clearly conglomeratic in origin. Simply because of its texture and with no reference to the metamorphic grade I called it a paragneiss during mapping on account of its crude, undulating, yet distinct foliation which is a gneissosity. It shows frequent interlayering of milky white quartz lenses or layers with a distinct pinch-and-swell structure and reddish brownish phyllite. Interlayering of quartz lenses with fine-grained aggregates of phyllitic character becomes more prominent where the grain size of the original conglomerate increases. At outcrop, there is a conspicuous continuity between the coarse conglomeratic gneiss and its finer sandy equivalent. In one isolated

686 outcrop, between the Gualdauna Hill and the Muntische, approximately 500 m
687 ESE of the Gualdauna Hill where the Funtaunas is probably in direct contact
688 with the Proliebas (in this case with the metatuff-like rock of the latter), it is more
689 of a psammite in character, with finer-grained quartz and an increased
690 proportion of sericitic matrix reaching to more than 50% of the rock.

691
692 Under the microscope, the typical mineral constituents of the metaconglomerate-
693 psammite group are mainly quartz, and muscovite with subordinate amounts of
694 opaque minerals. The prominent foliation of the rock is defined by fine, powdery
695 or lenticular masses of sericite and, to a lesser extent, by quartz. In the rock,
696 quartz makes up more than 50 % of the constituent minerals, whereas mica
697 varies between 20 and 40 %. The mica content increases with decreasing grain
698 size.

699
700 All of the rocks in the metaconglomerate-psammite group have undergone
701 important degrees of cataclasis. Except along the granulated borders of the large
702 quartz lumps and in the pelitic matrix where micas grew, recrystallisation does
703 not seem to have reached an advanced stage. As a consequence, this rock has
704 largely retained its granular texture. Lenses or "Augen" are the most conspicuous
705 textural features of this group. In the field, they are not nearly as prominent as
706 they are under the microscope. At the outcrop located W of Alp Nova, near the
707 point where the 2200 m contour crosses the road, they are also well-developed on

708 a macro scale. Mica flakes and small recrystallised or granulated quartz particles
709 wrap around the lenses. With respect to the brittle deformation that has affected
710 these rocks, most of the minerals are pre-kinematic. Undulose extinction is very
711 common.

712

713 Towards the upper parts of the metaconglomerate-psammite group, the matrix
714 becomes locally calcareous. This supports its age as Buntsandstein.

715

716 The protoliths of the metaconglomerate-psammite group were quartz-rich
717 conglomerates with clast size reaching fist dimensions, sandstones and shales.
718 They may well have been deposited in a high-energy fluvial environment,
719 possibly on and around alluvial fans. So far, attempts to interpret the
720 sedimentary environments of the equivalents of these rocks in the Lower
721 Austroalpine units have not been successful (generally grouped as Alpine
722 Buntsandstein; cf. Stöcklin 1949; Trümpy and Haccard 1969; Finger 1978), but a
723 comparison with the Chazforà Formation in the Engadiner Dolomiten, both with
724 respect to the rock type of the formation itself and the overlying and the
725 underlying units in both places (Dössegger 1974; Dössegger and Müller 1976)
726 makes it clear that the environment was fluvial in which energy of transported
727 decreased with time as shown by the calcareous upper parts of the
728 metaconglomerate-psammite group corresponding to the Chazforà-Fuorn

transition rocks (see Dössegger 1974; Dössegger and Müller 1976; Furrer et al. 2015, p. 18, fig. 3).

2b. Quartzite. The quartzite occurs in close association with the metaconglomerate-psammite group and in outcrops near it very much resembles it. It is best seen at the outcrop near Alp Nova. It is a coarse, brown, extremely quartz-rich rock with a crude, highly phacoidal foliation that is essentially a gneissosity. Under the microscope, the rock consists almost wholly of quartz occurring as fine-grained aggregates and coarse-grained porphyroclasts. Fine mica streaks and blastomylonitic calcite are also present separating quartz laminae consisting of elongated quartz grains.

Recrystallisation has resulted in a sugary, polygonal texture with triple junctions between the grains. The observed crude foliation is formed by the elongation of quartz crystals. There are quartz veins cutting through the rock that are clearly pre-foliation. Thus, the vein-filling minerals have also adopted the same elongation as the host rock.

2c. Metaconglomerate Group. This group of rocks greatly resembles the highly deformed metaconglomerates and the psammities of 2a both at outcrop and under the microscope. My reason for treating them as different from the 2a group of rocks is that they appear to be located immediately below the lower

751 Rauhewacke in Val Bella, as opposed to the position of the 2a group immediately
752 above the Proliebas Assemblage. However, the section in which it occurs is
753 located on the highly attenuated limb of a major fold where the rocks are much
754 strained. If a redistribution of rocks has occurred here, the 2c and 2a may be the
755 same rocks simply redistributed structurally. The state of outcrop does not,
756 however, permit a decision.

757

758 The metaconglomerate of the 2c group is a 1 to 2 m thick, greenish brown schist.
759 Under the microscope it is seen to consist dominantly of quartz and mica. It has
760 undergone a mild degree of dynamo-metamorphism which nowhere goes
761 beyond the very limit of the greenschist facies, as indicated by sericite and very
762 small, stout albite crystal growths. Quartz crystals have been elongated along the
763 foliation planes; around them very fine grains of quartz growth form thin halos.
764 Towards the top, the matrix of the metaconglomerate becomes carbonate rich,
765 similar to the case of the 2a group, but in the 2c group there are no actual
766 carbonate interlayers as they exist in the 2a group. In any case, these upper
767 carbonate-rich layers seem to provide a transition to the overlying calcschist-
768 marble group.

769

770 2d. Calcschist-semimarble group. Perhaps the best-defined stratigraphic unit in
771 the Funtaunas Assemblage is the calcschist-semimarble group. It can be
772 immediately recognised even in a hand specimen. It is best exposed along the

road north of Alp Nova. At outcrop the rock is a well-foliated, cream-coloured, light yellowish brown, or buff coloured phyllite. On the fresh surface it is grey to silvery grey. When treated with 10% HCl it fizzes slowly. Downwards it passes into carbonate-rich, grey-brown psammites, which, in their turn, pass into metaconglomerates of the unit 2c, all within about 10 m.

The calcschist-semimarble group has a distinct foliation that is itself crenulated giving rise to a very distinct crenulation cleavage with an average spacing of about 0.5 cm. A large number of calcite veins cut through the outcrop.

Under the microscope, the rock consists dominantly of calcite and, to a lesser extent, quartz and muscovite. The pronounced foliation results from an alignment of mica flakes. Where metamorphism reached greenschist grade, the calcite matrix became totally recrystallised to a fine-grained mosaic of calcite with subordinate quartz and mica. Towards the top of the section the effects of metamorphism diminish and the rock becomes a faintly metamorphic, fine-grained impure limestone with local recrystallisation. Impurities are represented by elongated fine muscovite, minute quartz crystals and possibly some albite.

In a number of outcrops, the calcschist-semimarble group passes upwards into a rather massive-looking, dark grey, fine to medium-grained dolomite. However, when regarded closely, one observes that the dolomite has a mini-spongy aspect,

795 possibly because of the dissolution of tiny amounts of anhydrite by ground
796 water recently (in the Quaternary, in any case). In other places, the group comes
797 very close to the outcrops of the lower Rauhewacke with no intervening micro-
798 spongy dolomite. In one place the calcschists actually interfinger, most probably
799 stratigraphically, with faintly metamorphosed impure limestones.

800

801 The entire Funtaunas Assemblage records a progressively more quiet
802 environment of sedimentation concurrent with the advance of the sea. The lower
803 members, such as the metaconglomerate-psammite unit and the quartzite,
804 probably record a terrestrial, more specifically a high- to medium-energy
805 fluviatile environment. Then the sea gradually came to cover the previous
806 terrestrial environment. In the ensuing beaches dolomites and anhydrite was
807 probably laid down much like in the present-day Persian Gulf sabkhas (i.e., salt
808 flats). During the early to medial Triassic, the world-wide sea-level rose to some
809 100 m higher than the present-day (see Bilal Haq in •engör 2015, fig. 15). What
810 we see in the stratigraphic record of the Funtaunas Assemblage can be accounted
811 for by this transgression without resorting to any extra tectonic subsidence.
812 However, there was much transtension in the future Alpine area both during the
813 Permian, but especially during the Triassic to create extension that would
814 inevitably have resulted in subsidence, although in the Albula area there is no
815 well-dated, unequivocal evidence for such extension before the latest Triassic to

Liassic. The Funtaunas Assemblage corresponds to the S-charl Formation of Furrer et al. (2015, pp. 17-18).

3. The lower Rauhewacke. Together with the Middle Triassic dolomites, the lower Rauhewacke forms the key horizon delineating the large-scale internal structure of the Albula Nappe (Figs. 2 and 6). It is best seen in Val Bella and between Punt Granda and Alp Nova. Its thickness changes from nil to 150 m, but these figures have little meaning because of the extreme deformation of the rock with strong thinning on the limbs of the major folds and thickening at their noses. It also slides easily localising thrusts.

The Rauhewacke (= *Zellendolomit*, i.e. cellular dolomite or *cargneule* or *cornieule* in France and the French speaking part of Switzerland) is a famous rock (the term seems to date back to the eighteenth century: see Krauter 1971) because of its massive and spongy appearance and in places foul smell. Its colour changes from dark grey through light grey, whitish grey, dirty yellow, buff to almost white. It fizzes strongly when treated by 10% HCl and generally crumbles when hit by a hammer. Locally, it is a calcite-cemented breccia, containing mainly clasts from the underlying and/or overlying units. It is commonly a tectonic breccia suggested to have formed by hydraulic fracturing because of overpressurisation either under a developing sedimentary pile diagenetically (Krauter 1971) or under the Alpine nappes (Masson 1972). In the Albula area, the rock types of the

838 Funtaunas Assemblage seem to have supplied most of the clasts to the lower
839 Rauhwacke breccias. Also present are small (mm to at most 2 cm diameter), platy
840 pieces of red shale. These shaly components remind one very much of the tuff
841 horizons present in the overlying Middle Triassic Dolomites and may well
842 represent the remnants of similar tuff horizons that may have once existed in the
843 lower Rauhwacke.

844

845 Under the microscope, the lower Rauhwacke gives the impression of a
846 completely recrystallised limestone with numerous open cavities (hence the
847 spongy look) vacated by dissolved evaporites. Some of the cavities are
848 rectangular in shape, suggesting the former presence of halite in addition to
849 anhydrite (see Heierli 1955).

850

851 Within the creek bed of the Ova d'Alvra, between Punt Granda and Alp Alesch,
852 there are a number of outcrops showing a clear stratigraphic transition from the
853 lower Rauhwacke into the Middle Triassic dolomites. Although much tectonic
854 disruption has occurred here as shown by the aberrant younging (shown by
855 coarser bases of laminae confined to channels) reversals within the dolomite, it
856 seems clear that the lower Rauhwacke passes upwards into grey dolomites
857 generally across a thin (0.5 to 2 m), rose-coloured limestone. It was named by
858 Eggenberger (1926, p. 546) as the Bergnelken-coloured layer (*Bergnelkenfarbene*
859 *Bank*) consisting of limestone and he believed it to be of Carnian age. Heierli

(1955, p. 67) has provided evidence on the basis of rock type and regional comparisons to argue for an Anisian age for the Bergnelken Limestone and located it directly above the Buntsandstein. My mapping entirely vindicated Heierli's interpretation of the stratigraphic position of the Bergnelken Limestone, mainly on the basis of the following section, which was also accepted by Bearth et al. (1987). Along the creek bed of Ova d'Alvra the lower Rauhewacke-Middle Triassic dolomites transition occurs as follows:

4. Massive grey dolomites

3. Pink, red, or white, fine-grained limestone, very coarsely foliated
(Bergnelken-coloured limestone)

2. "Bedded" lower Rauhewacke, grey, cream-coloured, yellow

1. Grey to buff or yellow weathering massive lower Rauhewacke

This zone of transition changes its thickness from more than 20 m to only 3 m.

Generally no sedimentary structures can be seen in any of the Rauhewacke.

Locally, however, erosion has preferentially removed 50 cm- to 1 m-thick layers on the face of some of the outcrops. The resulting striped pattern is commonly

parallel with the lithological boundaries of the Rauhewacke and may represent some kind of ghost bedding. As I show below, bedding is much more clearly developed (and preserved) in the upper Rauhewacke.

Because Heierli (1955) presented an extensive discussion on the origin of the Rauhewacke in the Albula region, I omit a genetic discussion here. I should like to point out only that the presence of perfectly angular clasts within the Rauhewacke indicates clearly that it is mainly a tectonic breccia as is apparently also the case in the Préalpes (see Baud and Masson 1975). That Rauhewacke almost always occurs along major shear zones supports this view. I fully agree with Heierli (1955) on the other hand that the appearance of Rauhewacke as we see it at outcrop today is dominantly an expression of karstic phenomena and at depth where neither surface nor subterranean water can affect it, it is encountered as a dolomite-anhydrite breccia. But, in contrast to Heierli (1955), I think the breccia in the Albula area is a product of tectonic events not sedimentary phenomena, because pieces of both the underlying *and* the overlying formations are found in it.

4. Middle Triassic dolomites. The dolomites of the Albula Nappe form the most conspicuous prominences in the landscape of the Albula Pass between the ridges of the Crasta Mora and the Blaisun-Gualdauna. The rock can be dark grey, bluish grey, buff or yellow and is massive. Considerable tectonic disruption and the

monotony of the rock type make a detailed stratigraphic subdivision impossible, although two common varieties can be distinguished solely on the basis of their colour at outcrop: a dark bluish variety containing *Diplopora* and a light grey, buff, or yellowish variety seemingly free of fossils. Conventionally, the darker variety has been ascribed to the Anisian, whereas the lighter variety to the Ladinian (e.g., Heierli 1955). Because of the lack of fossils and the apparent lack of any stratigraphic order I shall not here follow this date ascription. The minimum thickness of the Middle Triassic dolomites is on the order of 120 to 150 m.

4a. The darker dolomites. The colour of the darker dolomites range anywhere from dark, almost blackish grey through a mouse grey to bluish grey. The dark grey weathering surfaces are generally shiny, whereas those of the lighter ones are dull. The rock is commonly massive, medium to fine-grained and locally laminated. The laminae are parallel with one another, the thicknesses varying from about a mm to at most 7-8 mm. Locally, there are tiny channels filled with dolomite microbreccias. The younging evidence in these dolomites comes entirely from these mini-channels. The laminae containing these small channels are of a lighter tone of grey than the surrounding rocks.

Very rarely, the rock acquires a foliated appearance, which is probably associated with increased shale content. Commonly a brown-buff, rarely red, shale occurs

926 along bedding planes. The thickness of such shale layers rarely exceeds 0.5 cm
927 and they take up a weak foliation. By analogy with the occurrences in the
928 sedimentary zone of Samaden, I interpret these layers as possible tuff horizons
929 (Finger 1978), coeval with the Pietra Verde volcanism of the Southern Alps (e.g.
930 Oberholzner 1991; Krainer and Lutz 1995).

931
932 Locally, the darker dolomites are brecciated, with clast sizes ranging from 1-2
933 mm to 5 cm. The clasts are angular and are embedded in a dolomitic material of
934 a lighter shade of grey. In only one place I found a breccia, the clast size of which
935 is up to more than half a m in diameter with little or no matrix in between.
936 Although it is possible that some of the breccias of the Middle Triassic dolomites
937 may be storm breccias, a number of them are clearly collapse breccias resulting in
938 part from karstic phenomena of Triassic time.

939
940 These dolomites have a bituminous smell upon fresh fracture. Also, locally, they
941 have been invaded by innumerable calcite veins giving the rock a "brecciated"
942 look ("*Weisspat-Brekzie*").

943
944 The dark-coloured dolomites yielded in a number of places *Diplopora* and
945 unidentifiable mollusc fragments.

4b. The lighter-coloured dolomites. The lighter-coloured dolomites have a colour range from very light bluish grey to yellow. Similar to the dark-coloured ones, they are massive at outcrop and commonly laminated. Locally, they contain buff or pink chert nodules in clusters or in layers that stand out because of their greater resistance to weathering. Possible tuff horizons, occurring as shaly or cherty material in them are seen as discontinuous layers with an average thickness of some 0.5 cm. Calcite veining, so characteristic of the darker variety dolomites, have developed only rarely in the lighter-coloured ones.

The upper boundary of the Middle Triassic dolomites is rather abrupt wherever it is exposed in the Albula area. Across that boundary, which is clearly tectonic in nature, they abut against the upper Rauhewacke.

5. Upper Rauhewacke. As the "upper Rauhewacke" I designate what previous authors have called the "Rauhewacke train of the Albula Hospiz-Zuoz". Furrer et al. (2105) have designated within it a Middle and Upper Rauhewacke independent of my classification. My mapping does not justify following their scheme. As Fig. 2 shows, this rock has a wide distribution in the Albula Pass region and is located along the most important tectonic boundaries. The Albula Nappe stratigraphy is truncated along the upper Rauhewacke that displays the greatest thickness variations. The Ela nappe stratigraphy begins with the Carnian. It is thus impossible to tell how much of the upper Rauhewacke belongs

969 to the Albula nappe and how much of it to the Ela Nappe. However, the internal
970 structure of the Albula Nappe may be used as the basis of a working hypothesis
971 to assign all of the upper Rauhewacke provisionally to the Albula Nappe and to
972 the Infra-Ela slices. I adopt this scheme here for the sake of a coherent discussion.

973

974 Both the upper and the lower boundaries of the upper Rauhewacke are tectonic.
975 The highly irregular map pattern, albeit crudely conforming to the outermost
976 layer of an anticline, suggests that it was strained readily during the deformation
977 making any thickness measurement meaningless with respect to its original
978 stratigraphic thickness.

979

980 At outcrop, the upper Rauhewacke resembles the lower one in every aspect. It is
981 dark grey, mouse grey, cream-coloured or buff, commonly forming massive
982 outcrops devoid of any suggestion of internal structure. Locally, however, there
983 exists a crude "layering" brought out by differential erosion. Only in the outcrops
984 at Chaunt de la Crasta, a distinct colour and texture change defines what was
985 once clearly bedding. This ghost bedding defines mesoscopic folds with
986 amplitudes and wavelengths around 75 cm.

987

988 Fig. 7 shows a typical hand sample of the Rauhewacke. Its characteristic
989 appearance is its spongy aspect and the presence of foreign, angular clasts in it.
990 The upper Rauhewacke is in fact much richer in such clasts than the lower one

991 and this is in accordance with the fact that it lies along a boundary of much
992 greater displacement than the one along which the lower Rauhewacke had
993 become brecciated. In at least one place one such clast of the Middle Triassic
994 dolomites of the Albula Nappe in the upper Rauhewacke has a length of some 200
995 m and a width of some 60 m (location: 1 km east of the *Ospiz*). This "clast" clearly
996 originated as a tectonic sliver plucked from the top of the middle Triassic
997 dolomites during thrusting. The majority of the clasts are of much more modest
998 size, generally less than 10 cm. They are dominantly of red and green shales and
999 grey, fine-grained sandstones. In one place (S of the prominent bend in the road,
1000 1.3 km east of the *Ospiz*), there is a coherent outcrop of shales in stratigraphic
1001 contact with an impure dolomite breccia. By analogy with the occurrences at the
1002 base of the *Ela* Nappe, they probably represent the Carnian *Raibl* beds,
1003 tectonically incorporated into the Upper Rauhewacke (for a description of the
1004 *Raibl* Group, see Furrer et al. 2015, p. 19).

1005
1006 Under the microscope, the upper Rauhewacke appears as a completely
1007 recrystallised limestone with local silicification. It is essentially a highly
1008 disrupted, sparsely fossiliferous micrite. The fossils are molluscs. Some of the
1009 pores have rectangular edges indicating dissolution of the associated halite.
1010 Disruption of the rock is in part because of in-situ brecciation resulting from
1011 dissolution-related collapse, but the dominant mechanism of disruption seems

1012 tectonism shown by the abundant foreign clasts in it plucked from the
1013 underlying units.

1014

1015 Locally, the original dolomite has not been completely converted into a
1016 Rauhwacke and in such places the rock has a bituminous smell upon fresh
1017 fracture.

1018

1019 So far the stratigraphy of the Albula Nappe. In the next section I present a brief
1020 account of the stratigraphy of the other major sedimentary unit in the Albula
1021 Pass area, the Ela Nappe.

1022

1023 *Ela Nappe*

1024

1025 In the Albula Pass region the Ela Nappe (Zyndel 1912; Furrer et al. 2015, fig. 12)
1026 occupies the largest area and has a simpler internal structure than the other units.
1027 The rocks of the Ela Nappe range from the Carnian (Raibl Beds) to possibly the
1028 Upper Jurassic (the top sequences of the Allgäuschiefer, a designation not
1029 adhered to in this paper except for reader orientation purposes). This sequence
1030 can be divided into three main subdivisions, namely the Upper Triassic
1031 dolomites (Hauptdolomit with the heterogeneous Raibl Beds), early Liassic
1032 sedimentary breccias (the Alv-Breccia) and the Lias to Dogger calciturbidites.
1033 The following radiolarite deposition in the Alpine Neo-Tethys may have begun

already during the Bajocian, but certainly not later than Bathonian (Bill et al. 2001). I describe these below beginning with the oldest Hauptdolomit, in sufficient detail to follow the structural arguments.

1. Upper Triassic dolomites. In the Albula area, the Upper Triassic dolomites crop out in the majestic Piz Üertsch, in the small prominence just to the east of the small pond on the road lying east of the *Ospiz*, and along the south face of the Gualdauna Hill. In the Piz Üertsch, the dolomites sit clearly on younger beds across a clear tectonic contact. In places a massive, grey dolomite sits directly on the Ela cherty limestones and the Fuora da l'Üertsch turbidites. In other places, yellowish grey dolomite beds of about 30-70 cm thickness alternating with calcareous shales occur along the base of the massive grey dolomites. Wine-red, fine-grained sandstones and shales occur along this lower unit as 1 to 3 m wide lenses. Along the southern and the eastern face of the imposing dolomite massif of the Piz Üertsch, I have failed to find any stratigraphic contact of the dolomites or the Raibl Beds with other rock types, an observation in accord with Heierli's (1955) earlier mapping and with Furrer et al.'s (2015) synthesis (their geological profile 3 in the foldout plate).

A much more diverse development of rock types including buff-grey, highly veined dolomites, buff-yellow calcareous sandstones and grey-black shales most likely representing the (tectonically-intercalated?) Raibl Beds, and very fine-

1056 grained sandstones are exposed about 50 m E of the *Ospiz*, right on the southern
1057 side of the road (Bearth et al. 1987). The outcrop is isolated and lies structurally
1058 above the Upper Rauhewacke, although no contact between it and the Rauhewacke
1059 is exposed. The rock types are all thin (each at most 2 to 3 m thick, but clearly
1060 tectonically reduced) and seem folded about WNW-trending axes.

1061

1062 Immediately N of this outcrop, across the road, is an even smaller outcrop of the
1063 Alv Breccia (see below).

1064

1065 The two other occurrences of the Upper Triassic dolomites belonging to the Ela
1066 nappe are found at the base of the sections exposed on the small hill just east of
1067 the pond on the road and on the Gualdauna Hill.

1068

1069 At outcrop the rock is a light, somewhat bluish grey dolomite. Locally it may
1070 acquire a darker grey or a yellowish grey colour. Where yellowing occurs, it
1071 seems associated with an increase in the stylolite horizons traversing the rock,
1072 which, in such places, meet each other at syntaxes. Between stylolite virgations
1073 the rock is almost "brecciated" by the anastomosing stylolite horizons. On the
1074 fresh surface, the dolomite is somewhat darker, but still bluish grey. Where
1075 stylolite concentrations increase, the fresh surface assumes an almost a blackish
1076 blue-grey tone. Other than faint laminations that are discontinuous along the
1077 strike, no sign of any bedding or any facing indications are visible. Neither is

there any sign of brecciation. The rock itself is fine-grained, whereby crystal size diminishes as one goes upward in the sequence. Under the microscope, the rock is a finely crystalline hypidiomorphic dolomite. Locally, the dolomite is in the form of possibly biogenic nodules surrounded by thin clay laminae. These clay laminae resemble those caught up between calcareous stromatolite layers. There is chertification in places.

At outcrop, regular, lamination-parallel, stylolites are seen, along which brown, dirty blackish and reddish brown undissolvable material has accumulated. The rock is cut by numerous white calcite veins of variable thickness between 1 mm to at most 1 cm. Upon fracture the rock releases a bituminous smell that very quickly vanishes.

In the road section, the thickness of the dolomite is at most 5 m whereas at Gualdauna Hill it changes from 8 to 10 m.

In neither outcrop the bottom of the dolomite is exposed, but lies hidden under Quaternary (Hill section) or under scree (Gualdauna). However, especially in the road section, the upper Rauhewacke outcrops are only several m away from the dolomite. Clearly one of the numerous discontinuities that cut through the highly ductile rock that predated the Quaternary Rauhewacke formation passes between it and the Upper Triassic dolomites delineating the base of the Ela

1100 Nappe. In the Gualdauna section there are two tiny, entirely isolated occurrences
1101 of a dark, fine to medium-grained dolomite, highly disrupted by criss-crossing
1102 calcite veins, which are found at the eastern end of the base of the Gualdauna
1103 vertical south face. They can be none other than torn pieces of the Middle
1104 Triassic dolomites with *Weisspatbrekzie*. However, it is unlikely that they come
1105 from the Albula Nappe in its present geometry. I shall show below that they are
1106 related to the slices shaved off the Albula Nappe as the Ela Nappe was being
1107 separated from it during southward thrusting.

1108

1109 As I mentioned above, there are three, extremely thin (total thickness ± 50 cm on
1110 the road section) rock types underlying the Alv Breccia and separating it from
1111 the Hauptdolomit below. Because they are so thin, they could not be shown in
1112 Fig. 2, but I describe them below:

1113

1114 Immediately above the Hauptdolomit is a navy-grey limestone with a thickness
1115 of 10 cm to about 2 m. It is a finely laminated micrite with some fossil fragments.
1116 It is intricately folded together with a green to black shale. The thickness of the
1117 shale changes from a few cm to a m. Under the microscope, this rock turns out to
1118 be an extremely deformed, dedolomitised and silicified shale. Calcite crystals are
1119 distributed in a clayey matrix in the form of irregular aggregates. The rock is so
1120 tectonised that clearly a high strain zone goes through it, although stratigraphic

passage both downwards and upwards into neighbouring rock types is obvious.
It represents a slid original stratigraphic contact.

Above it an at most 2 m thick, red, locally shaly, laminated micrite interbedded
with medium-bedded, lighter coloured, fossiliferous micrite is observed. This
rock does not crop out in the road section, where in its place, a several cm-thick,
possibly dolomitic yellow-greyish chertified carbonate occurs.

It is above these that an environmentally critical rock unit sits in stratigraphic
contact: it is the Alv Breccia, or the Alv Formation as formalised by Schüpbach in
the Piz Alv area in the Bernina Nappe (Schüpbach 1973, p. 14; see also Decarlis et
al. 2015).

2. Alv Formation. One of the main objectives of my mapping was to determine
whether the Alv Breccia belongs to the Ela Nappe or tectonically separated from
it in a *Gualdauna Schuppe* as Heierli (1955) and Eberli (1985a, b, 1988: *Gualdauna
Slice*) believed (see also Bearth et al. 1987). This determination is critical in
deciding where the Ela Nappe belongs palaeogeographically. That is why, the
relationship between it and the overlying Kieselkalk, which clearly belongs to the
Ela Nappe, have been more closely studied than other sections in the Albula Pass
region. For comparisons with the Piz Alv occurrence, the reader is referred to

1142 Schüpbach's thesis (Schüpbach 1973) and to Decarlis et al. (2015). Neither
1143 publication gives nearly as much detail as in the following, though.

1144

1145 Both in the road section and in the Gualdauna Hill the Alv Breccia is followed, in
1146 clear stratigraphic continuity by the Kieselkalk. However, whereas in the road
1147 section the Kieselkalk is in direct contact with the Alv Breccia, in the Gualdauna
1148 Hill a number of other rocks intervene between the two. That is why, in the
1149 following it is the more complete Gualdauna section that is described:

1150

1151 2a. The Alv Breccia. This is really a lump name for a variety of breccia types that
1152 are all of carbonate composition. The matrix is commonly very fine-grained
1153 limestone which is a micrite; less commonly it is crystalline calcite filling
1154 between, and sometimes through, the breccia clasts. The colour of the rock is
1155 generally grey, creamy yellow or pinkish red, but often varicoloured, in which all
1156 the colours just mentioned occur. From bottom to top, but locally also
1157 interfingering laterally, the following breccia types can be distinguished in the
1158 field:

1159

1160 2a1. Limestone breccia-1. The clasts of this sub-type are made up of dark grey,
1161 fine-grained limestone. On fresh surface, the limestone appears somewhat darker
1162 and somewhat sparkly. The weathered surfaces are dull and rough. The matrix
1163 consists of buff-weathering, clay-rich limestone. It surrounds the clasts in such a

way as to define a very crude "phacoidal foliation". Clast thicknesses in this "foliation" are in the range of 6 to 15 cm.

2a2. Limestone breccia-2. This sub-type of the Alv Breccia consists of subangular, thoroughly jointed and veined, navy grey-weathering limestones. On the fresh surface, these limestones look homogeneous and sparkly dark grey. The matrix in this case is entirely made up of very dark grey-brown crystalline calcite (vein-filling type). Locally, buff weathering, shale-rich, matrix looks as if squirted between the clasts.

2a3. Multi-coloured limestone-dolomite breccia. Just above the limestone breccia of type 2a2, a multi-coloured breccia occurs, in which it is locally impossible to tell what is matrix and what is clast. The colour of the rock may be blue-grey, dark grey, yellow, pink, or wine-red, although these colours do not correspond to particular rock types. For example, micrites and very shale-rich limestones can be both wine-red in colour. Similarly both limestones and dolomites may be of the same tone of grey. Upwards in the section the clasts and the matrix become locally indistinguishable. In the lower parts, the following components are observed:

2a3-1. Very dark grey weathering limestone clasts that seem to have been tectonically flattened. The average width of some of the clasts is about 10 cm,

1186 whereas their length is never more than 20 cm. Within one clast, the colour
1187 changes to navy green with red patches; this latter part is almost a coquina, being
1188 made up entirely of shell fragments. Locally, the limestone becomes entirely
1189 pink, where the coquina character disappears.

1190

1191 2a3-2. Grey weathering massive dolomite with cream-coloured patches of about
1192 0.5-1 cm diameter with commonly irregular forms and diffuse boundaries; most
1193 likely pieces of the Hauptdolomit. The fresh surface of the dolomite is hardly
1194 distinguishable from the weathered surface. Clast size is similar to that in 2a3-1.

1195

1196 2a3-3. Beige, very smooth weathering limestone; locally dark beige with diffuse
1197 red spots. On fresh surfaces, the colour of the rock is nearly the same as that on
1198 the weathered surface, being somewhat darker and sparkly. This limestone fills
1199 in spaces between the components of 2a3-1 and 2a3-2. It is almost always
1200 surrounded by a rim of fibrous calcite of grey and locally dark brown colour. It
1201 appears that first, the limestone making up the 2a3-1 and the dolomite forming
1202 2a3-2 fractured and the resulting fissures, "neptunian dykes", were filled by
1203 fibrous calcite vein filling. As fissuring progressed filling by crystalline calcite
1204 could not keep up with the opening rate of the fissures and the resulting cavities
1205 became filled with the limestone of 2a3-3. In many cases, the rock broke up into
1206 small fragments (clasts) that became embedded into 2a3-3.

1207

2a3-4. A light, navy grey limestone with numerous darker spots of 1 to 2 mm diameter and thin (never thicker than 1 mm), highly irregular, red stripes. On the fresh surface, the colours appear unchanged. This rock forms clasts of varying sizes in a pink and locally buff weathering, fine-grained micrite. Locally, that micrite forms highly irregular pieces, completely surrounded by fibrous calcite of grey colour. Locally, red or purplish red shale makes up the matrix. In the matrix, one also finds large (40-50 cm) blocks of a dull light grey weathering fine-grained dolomite. The dolomites have resisted the weathering better than their surrounding matrix and consequently they stand out on the surface of the outcrop. Extensive *Karren* (=lapiès) characterise the surface of these dolomitic blocks which do not continue into the surrounding matrix. Locally, the breccia looks as if made up entirely of these dolomite clasts embedded in a buff-weathering, crystalline limestone, which is locally clay-rich. In such places, the clast to matrix ratio seems to be around 6 to 4. Also locally, the clast size can become very large and the shapes equant and angular; in such cases, blocks may have edges as long as 1 m.

In summary, the overall impression one obtains of the mode of formation of the Alv Breccia, to be tested below while investigating the succeeding rock types overlying it is that it consists of disrupted late Triassic carbonates and some products of their subaerial karstification. In places, the disruption never went beyond the production of neptunian dykes and associated cavities. In the

1230 neptunian dykes, the opposing margins can often be fitted back. In other places
1231 polygenetic breccias with a micrite matrix was produced, included subrounded
1232 grains. In the field, the two types are gradational into one another. The Alv
1233 Breccia greatly resembles the contemporary Macchia Vecchia Breccia of the
1234 Southern Alps (Bernoulli and Jenkyns, 1909, fig. 34).

1235

1236 The following scenario best fits the history of origin of the Alv Breccia in the
1237 Albula region: First, extensive normal faulting exposed various levels of the
1238 Upper Triassic dolomites to subaerial weathering, whereas other parts remained
1239 below sea-level; along fault margins collapse breccias formed at precipice feet;
1240 along the tilted backs extension created submarine neptunian dykes and ponors
1241 filled with pelagic limestones. Smaller ones just became veins of various widths
1242 and geometries and were filled with crystalline calcite. The parts that remained
1243 above sea-level became extensively karstified as a result of which irregular
1244 cavities, ponors of various sizes, formed. These were filled with carbonate rock
1245 debris and terra rossa; the latter now appears as iron- and aluminium-rich red
1246 clays. There are also a few hard ground horizons.

1247

1248 The process of karstification, under tropical conditions, was going on
1249 concurrently with a regime of extension, the ultimate cause of which may have
1250 been transtension. Ongoing extension encouraged widespread neptunian dyke
1251 and ponor building. Continuing fissuring locally led to complete disintegration

of the rock into angular fragments of very irregular sizes leading to the total destruction of some high-standing blocks. The disintegrated material, made up of extensively jointed blocks containing partly or completely filled karstic cavities and structural fissures, was carried along with masses of terra rossa under the sea by two main mechanisms: either they gently subsided on the backs of rotating normal fault footwall blocks, or they collapsed along the fault precipices, both subaerial and submarine. Entire karstic plateaux may have thus foundered as a result of ongoing stretching to be invaded by micritic limestones that quietly filled depressions and cavities and became mixed with the terra rossa clays. Finally, the entire area on which the Alv Breccia formed subsided to bathyal depths (□ -500 to -1000 m) and the turbiditic material of the Kiselkalk and the Allgäu turbidites were laid down on top of it.

The scenario outlined above explains all the characteristics of the Alv Breccia in the Albula area, but also outside it and is consistent with the model suggested by Schüpbach (1973) and Decarlis et al. (2015(in press)), except the extension probably had a strong transtensional component here. It also reconciles Schüpbach's model of a subaerial origin with Heierli's (1964) find of *Belemnites acutus* and *Millericrinus* sp, *Pentacrinus* sp. and unidentifiable foraminifera. According to the model developed here, various components of the Alv Breccia formed in various environments, which needed not to be far away from one another and which themselves were very unstable. In the beginning, some blocks

1274 of the extensive late Triassic carbonate platform forming the northern continental
1275 margin of the Palaeo-Tethys (Laubscher and Bernoulli 1977; •engör and
1276 Atayman 2009; Decarlis et al. (2015) were rising, whereas others were sinking.
1277 Because of normal-fault-related tilting a part of the same block may have been
1278 rising, while another part was sinking. Bertotti's (1991) magnificent work from
1279 the Monte Generoso Basin in the western part of the Southern Alps near Lugano
1280 has demonstrated that the normal faults disrupting the Triassic platform were
1281 mostly large lystric faults going as far deep as 10 km into the crust. This kind of
1282 normal fault behaviour is what current seismicity in the western Turkish
1283 extensional systems corroborates (Eyido•an and Jackson 1985). They were found
1284 and mapped also in the Lower Austroalpine units in Graubünden as we shall see
1285 below. Wholly submarine fissuring gave rise to entirely micrite-cemented
1286 breccias within neptunian dykes.

1287

1288 Now let us see whether this model stands up to the information gathered from
1289 the sedimentary rocks overlying the Alv Breccia.

1290

1291 3. Red, clayey, sparsely packed micrite. This is the first rock type that overlies the
1292 Alv Breccia in direct stratigraphic continuity in the Albula region. At outcrop it
1293 appears as a red to purplish red, thinly-bedded fossiliferous biomicrite. Under
1294 the microscope it is seen to be sparsely packed, echinodermal biomicrite. 20 % of
1295 the rock consists of completely broken up and worn down echinoderm

fragments. Some pelagic mollusc fragments make up 2-3% of the rock. Throughout the rock there are scattered siliceous spicules. Echinoderm fragments rarely show syntaxial overgrowths. All of these fossil fragments appear to be floating in a micritic matrix. Grain size ranges from 40 μ m to 1 mm or so. Solution contacts and stylolite surfaces are marked by insoluble residue. These appear in most thin sections. Particularly the stylolites are parallel with the bedding and indicate extensive compaction and/or tectonic flattening under the extensional regime (cf. Baud and Masson 1975). The rock is in general poorly-sorted.

Broken and worn echinoderm fragments in a micritic matrix may indicate that such grains were transported from a shoal area with agitated water, where these animals were flourishing. The rock as a whole betrays a low energy environment during its final deposition, most likely below wave-base, but not too deep as no pre-induration solution features have been observed. Eberli (1988) thought that the Alv Breccia has been transported by rock falls. It may be only in part so, and in part much finer grained sediment such as the one here described may have been carried by storms to deeper water, similar to the hurricanes that regularly do much damage to the benthonic communities in the present-day Windward Islands in the Caribbean.

1317 4. Bluish grey, chertified, sparsely packed biomicrite. This rock follows that of 3
1318 in stratigraphic continuity and is itself 4 m thick. It consists of thinly-to medium
1319 bedded, bluish grey, hard, chertified, sparsely packed biomicrite. Under the
1320 microscope the allochems are seen to consist of *in situ* spicules (about 20%) that
1321 are represented by medium-size crystalline calcite and broken and probably
1322 transported echinoderm fragments (about 3%), all of which lie embedded in a
1323 micritic matrix.

1324

1325 The presence of abundant calcareous spicules and micrite in this rock indicates a
1326 low energy environment (probably no deeper than 10 to 15 m) at which depth
1327 the sponges live abundantly (Wilson 1975). The scarcity of echinoderms indicates
1328 that this area was far away from the echinodermal shoal or perhaps the primary
1329 absence of echinoderms.

1330

1331 5. Black biomicrite. In stratigraphic continuity with the former unit, this rock
1332 consists of black, thinly to medium-bedded, packed echinodermal biomicrite.
1333 The total thickness of the unit is two metres. Under the microscope, it is seen to
1334 be a tightly packed echinodermal biomicrite with some probable sponge spicules
1335 or calcspheres. Ecihoderm fragments (40%) are badly worn and now and then
1336 show syntaxial overgrowth. Possible spicules make up 15% of the rock. One or
1337 two ostracode fragments are also present. Stylolites and pressure solution
1338 contacts are seen between the echinoderm fragments. The rock is moderately

sorted and may be called an encrinite indicating a moderate energy environment. Alternatively, all of its constituent grains could have been generated in an agitated environment and then transported into their micrite-depositing quieter environment. Given the very unstable tectonic environment of normal faulting here at the time, it is difficult to choose between these two possibilities. The black colour, however, may indicate transportation into a poorly-oxygenated fault-bounded small basin.

6. Fossiliferous grey micrite. This is a grey, blackish dull grey, thinly- to medium-bedded and in part laminated, fossiliferous micrite. The macrofossils it contains are small belemnites (2-3 cm-long) and fucoids. Under the microscope it is seen to be an echinodermal biomicrite with 25% mostly broken and severely worn echinoderm fragments. Up to 10% mollusc fragments are also present, represented mostly by casts. Locally, the echinoderms are silicified. The rock is cut by numerous stylolites. The size of the belemnites show that the age is probably Lias.

This rock is the product of a low energy environment, most likely below the wave base. As indicated by their state of preservation, all of the fossil fragments appear to have been transported into the quiet environment of micrite deposition.

1361 7. Fossiliferous and lithic, laminated micrite. This rock represents the onset of
1362 what one may term the Allgäuschiefer regime of dominantly deeper water
1363 turbiditic sedimentation in basins formed by the widespread extension and
1364 subsidence during the Alv Breccia time. The thickness of this entire unit changes
1365 between 5 to 10 m and it finally leads into the Kieselkalk deposition.

1366 At the base, the unit 7 consists of fossiliferous micrite with 5% echinoderm and
1367 mollusc fragments, all badly worn and obviously transported into a low energy
1368 environment. Upwards in the rock, one observes an influx of lithoclastics into the
1369 area made up mainly of quartz and limestones. It is at this level that
1370 microlaminations appear. Finally, the rock becomes an extremely sparsely
1371 fossiliferous, laminated micrite that has been mostly neomorphised into
1372 microspar.

1373

1374 Before I begin the description of the Allgäuschiefer calciturbidites, it would
1375 perhaps be useful to summarise the evolution of the sedimentary environments
1376 in the Ela Nappe since the Upper Triassic to see how my model for the origin of
1377 the Alv breccia has so far fared. This will also set the scene for the succeeding
1378 turbidite deposition.

1379

1380 During the late Triassic a supratidal environment probably developed in the
1381 Albula pass area, where stromatolitic (?) dolomites (Norian Hauptdolomit)
1382 formed. The existence of horizons of storm breccias within it at the Piz Üertsch

substantiate the presence of a very shallow environment where the lower shoreface may be well on the order of tens of metres (as wavelength of the hurricane waves decrease when the waves reach the shelves; wave base is usually 1/2 of the wavelength).

Later, during the Rhaetian a slight sea-level rise resulted in the appearance of normal marine conditions, under which siltstones, shales and the laminated micrites were deposited. The progressive, tranquil transgression of the sea continued until the Lias, but the overall environment remained shallow marine. In the following discussion, one should bear in mind that from the Norian to the Hettangian, the world-wide sea-level dropped by nearly 80 m (Bilal Haq in •engör 2015). So every time a sea-level rise is recorded in the sedimentary record until the Hettangian, it was the bottom subsiding, not the world-wide sea-level rising.

Atop the Triassic supratidal environment, the Alv Breccia formed by the disintegration of the former fissuring-aided karstification, collapse and tectonic break-up. Because certain areas became emergent they were karstified. But this emergence was not the consequence of the world-wide regression, but the rise of the footwall blocks of normal faults accomplishing the extension, as I said above. There may have been here a significant strike-slip component also, which I discuss in the penultimate section of this paper.

1405

1406 Following the formation of the Alv Breccia during the latest Triassic-Hettangian
1407 interval, the region foundered somewhat and carbonate slope conditions became
1408 established, represented by the echinodermal, sparsely packed biomicrites and
1409 fossiliferous micrites containing sponge spicules. As shown by many authors,
1410 calcareous sponges live only under normal marine conditions at depths less than
1411 100 m. The kind of echinodermal facies encountered at the Gualdauna Hill has
1412 been very common on carbonate slopes flanking Palaeozoic and Mesozoic
1413 shelves (see, for example, plate XXI in Wilson 1975). This slope moved up and
1414 down as indicated by the succeeding rock types until it definitely subsided to
1415 receive the Allgäuschiefer turbidites beginning by the Sinemurian to
1416 Pliensbachian interval (Eberli 1985a, 1985b, 1987, 1988). From what has so far
1417 been said, it is clear that we are here dealing mostly with the footwall block of a
1418 normal fault that progressively subsided from a subaerial location to bathyal
1419 depths.

1420

1421 3. Lias to Dogger (?Malm) calciturbidites. The Pliensbachian to Dogger (to
1422 possibly Malm) turbidites form the next and the final stratigraphic unit of the
1423 Albula area. Eberli (1985a, 1987) in his sedimentological study called what used
1424 to be known under the designation Allgäuschiefer (first introduced by Gumbel
1425 1856, p. 38) or Algäuschichten (*sic!* first introduced by von Richthofen 1859, p.
1426 109), the Allgäu Formation (for the history of investigation of the Allgäuschiefer,

also known as Fleckenmergel, see Jacobshagen 1965; for a nomenclatorial review predating both my and Eberli's work, see Fischer 1971, p. 5). For mapping purposes at a scale of 1/10,000 or even 1/25,000, this classification is too broad and needs to be further subdivided, which was not done by Heierli (1955), Eberli (1985a, 1985b, 1987, 1988) or Furrer et al. (2015). For my purposes, the term Allgäu needs to be raised to group status and be divided into the following formations. In naming them I stuck to traditional names wherever they seemed present and appropriate. I also do not formalise them in view of the small area mapped, as mentioned earlier. When a larger area is mapped, these names may be formalised if found appropriate:

3a. Kieselkalk (cherty limestone) formation. Exposed mainly in small creek beds along the southern face of the Blaisun Ridge, the Kieselkalk is the first unit of the calciturbidites deposited on the previously described basal units of the Ela Nappe. At outcrop, it is a thinly-bedded, blue-grey to blackish grey, fine-grained limestone of about 100 m thickness with shaly (grey to dirty buff-weathering) and cherty (yellowish cream to grey) streaks and nodules (Fig. 8). On the fresh surface, the limestone bands are sparkly dark grey; the shaly sectors are nearly of the same colour with the exception that when they are broken along the foliation surfaces, the colour of the latter looks a shiny blackish-grey suggesting considerable flattening of phyllosilicates across the bedding planes. The fresh surfaces of the cherty sections are grey, but of a lighter and bluer than that of the

1449 limestones. Under the microscope, the limestone is seen as a largely
1450 recrystallised sparry calcite-microspar with abundant quartz crystals.

1451

1452 In the Kieselkalk the most obvious structure that is seen at outcrop is an ever-
1453 present, thin lamination along which the rock is easily split. Except for rare
1454 breccia lenses, composed mainly of bluish grey limestone and dull light grey
1455 dolomites and the lamination, neither any sign of bedding nor any primary
1456 structure is observed? Kieselkalk never shows any indications of younging to the
1457 naked eye again with the exception of the breccia lenses (that were clearly former
1458 channels). A faint grading near the contact of the Alv Breccia in the road section
1459 shows that there the Kieselkalk youngs away from the Alv Breccia, as expected.

1460

1461 As mentioned above, the shaly sectors are grey to dirty buff. However, in detail,
1462 the buff coloured sectors (generally a few mm to about 0.5 cm) are separated by
1463 thin, locally sub-parallel and locally anastomosing grey to blackish grey laminae,
1464 the thickness of which is never beyond 2 mm. They are papery and on parting
1465 surfaces shiny. In few places, as in the easternmost outcrops of the Blaisun Ridge,
1466 the Kieselkalk occurs in beds of 50-60 cm thickness in the form of hornstein. In
1467 such places no shales are seen in it.

1468

1469 The existence of parallel laminae in the limestones (probably micrites before
1470 recrystallisation) and the almost ubiquitous association with shales gives the

impression that the Kieselkalk corresponds with the T4 type (i.e., T d-e: Bouma 1962, p. 50, corresponding to what Mutti and Ricci-Lucchi 1978, p. 8-10 called the D, i.e., pelitic arenaceous facies; the turbidite facies of the Kieselkalk corresponds to the D3 facies of Mutti 1979, fig. 7), but in calcareous turbidites a somewhat different facies subdivision is necessary that takes into account the recrystallisation and dissolution features of carbonates. Fig. 8 shows the parallel laminations and the beds in the Kieselkalk, where the terrigenous component is closer to sand size than to clay size. Because of complicated folding and abundant internal imbrication, no stratigraphic thickness of the Kieselkalk can be given with any certainty. Moreover, there is at least 30% bedding-perpendicular shortening in many sections (Fig. 9). The present thickness may be on the order of 100 m. Eberli (1988, Fig. 9) estimated its original thickness around 80 m, which, given the bedding-perpendicular shortening, is a serious underestimation. The original thickness may have been 120 to at most 160 m.

3b. Fuora da l'Üertsch formation. This formation consists of black to grey-black, locally mouse grey to buff-brown calci-turbidites. The distinction from the Kieselkalk is provided by an increased amount of shaly portions and the black-buff-brown colour. The limestones are cm- to dm-thick, regularly-bedded, locally neomorphised fossiliferous micrites. The present unit thickness is on the order of 110 m; it was probably thicker originally, but I cannot now tell how much. The fossils are mostly echinoderm and mollusc fragments. On the NE ridge of the Piz

1493 Üertsch, where the best section of the Fuora da l'Üertsch formation is exposed, I
1494 have found a 15 cm-long broken Belemnite testifying to a post-Hettangian age
1495 (Dr. Gregor Eberli, pers. comm. 1981). The echinoderm fragments are partly and
1496 selectively silicified. There are sparse opaque materials. Numerous stylolites are
1497 parallel with the bedding, indicating considerable bedding-perpendicular
1498 shortening. Towards the upper parts of the Fuora da l'Üertsch formation, the
1499 limestones become completely recrystallised and silicified.

1500

1501 At outcrop, the most conspicuous structure of this formation is a parallel
1502 lamination that is generally parallel with a pronounced solution cleavage in the
1503 limestones and a slaty cleavage in the shales. North of the Blaisun Ridge one sees
1504 convolute bedding. A cleavage/bedding intersection lineation is seen, but as this
1505 gives little structural insight in this well-exposed area, I did not bother
1506 measuring it systematically.

1507

1508 In the lower parts of the Fuora da l'Üertsch formation, there are a number of
1509 breccia lenses very reminiscent of those found in the Kieselkalk. They look like
1510 channel fills and are composed of angular to sub-angular, very finely crystalline
1511 dolomite fragments, ranging in size from a few mm to 2.5 cm, embedded in a
1512 micrite matrix. The colour of the clasts is light grey, whereas that of the matrix is
1513 black grey. It is possible that a normal fault block exposing Hauptdolomit may
1514 have been supplying these clasts to the turbidites.

1515

1516 Along the easternmost part of the Blaisun ridge and on the outcrop extending
1517 north of the peak at 2923 m a. s. l., the individual turbidites of the Fuora da
1518 l'Üertsch formation have bottoms composed of very finely conglomeratic (at
1519 most 4 mm clast size) horizons of about 1 to 3 cm thickness. In some cases,
1520 bottom structures are observed.

1521

1522 These considerations lead me to conclude that the Fuora da l'Üertsch formation
1523 contains nearly all the members of a complete turbidite (a to e in the sense of
1524 Bouma 1962), but never in any one place! They are distributed so as to give
1525 support to the submarine fan model of Mutti and Ricci-Lucchi (1978), which is
1526 what one would expect. In the Albula region, one is probably looking at a small,
1527 facies-cutting, slice of such a fan. The Fuora da l'Üertsch turbidites are mostly
1528 base cut-out sequences extending as far as Bouma's c-horizon. Cross-laminations
1529 are only rarely observed. In one outcrop, north of the Blaisun ridge, the sequence
1530 contains bedded chert, i.e., already a part of the surrounding deep basin around
1531 the fan(s).

1532

1533 3c. Rust-coloured turbidite formation. This very thin sequence (at most 5 m) is
1534 perhaps the most conspicuous rock of the entire calciturbidite sequence of the Ela
1535 Nappe and can be recognised even from a long distance. It consists of black to
1536 rust red, thinly-bedded, laminated calciturbidites composed of interbedded

1537 limestones and calcareous shales of nearly equal amounts. They are typical distal
1538 turbidites.

1539

1540 Under the microscope, the limestones appear to be intra-sparites with 3-4 % lithic
1541 elements composed mainly of quartz and plagioclase feldspars. Abundant in the
1542 rock are opaque, rounded to sub-rounded grains of manganiferous micrite
1543 (Eberli 1985a, 1988; these grains could be manganese coated peloids). These
1544 opaque grains in fact give the rock its distinctive colour and indicate their
1545 deposition most likely during the Toarcian ocean anoxia identified by Jenkyns
1546 (1985, 1988). Other grains consist of recrystallised limestones and abundant
1547 echinoderm fragments. The matrix in which these grains are dispersed is a
1548 poorly-washed sparry calcitic groundmass which may be neomorphised micrite.
1549 Locally diagenetic silicification is seen. At the base of each turbidite sequence
1550 there is a thin (0.5-1.5 cm), graded microconglomerate. In one outcrop in the
1551 Fuora da l'Üertsch I have seen a flute cast plunging and trending as follows:
1552 15°/75° N on a right-side-up bed striking 110° and dipping 65° N. This shows a ±
1553 NW-SE orientation for the direction of the original turbidite transport.

1554

1555 The rust-coloured calciturbidites contain only parallel laminations.

1556

1557 3d. Blaisun turbidite formation. The Blaisun turbidites follow the rust-red
1558 calciturbidites of the previous sequence with a similar turbidite sequence. The

passage is marked by the rather abrupt disappearance of the rust-red colour. Highly foliated calciturbidites rapidly give way to medium-bedded ($\approx$ 50 cm) calciturbidites alternating with 5 to 215 cm shale beds. Within about 200 m of the contact with the rust-red sequence, two thick shale horizons, each 2-3 m thick, interrupt the normal limestone-shale intercalation. These thick shale units have 2 or 3 to 20 cm-thick limestone horizons in them. The shales have a well-developed phacoidal cleavage.

In general, the Blaisun turbidites consist of black, medium to thickly-bedded, fossiliferous, sparsely packed, biomicrite to micrite, interbedded with calcareous shales. Generally, the limestone beds have conglomeratic bottoms made up of very angular intra-lithoclasts and in places abundant echinoderm fragments. As a whole, limestones predominate over shales at outcrop. The individual limestone beds are about 25 cm-thick, whereas the shales are on the order of 10-20 cm-thick. Grading at the bottoms of the limestones is common. The bottoms of individual turbidites are characterised by sole marks consisting of large-scale groove casts, cut-and-fill structures. The limestones are partly chertified.

Under the microscope, the bottom parts of the limestone beds consist of rounded to sub-angular clasts. These are partly chertified echinoderm fragments (abundant), fossiliferous micrite fragments, dolomites and recrystallised limestones. Matrix is micrite and the clast size changes between 1 and 3 mm. In

1581 the rock there are abundant stylolites and pressure solution boundaries. Farther
1582 up, towards the middle of an individual turbidite, the rock is a micrite, locally
1583 fossiliferous with commonly authigenic silica dispersed in the micrite.
1584 Silicification and chertification is common.

1585

1586 The thickness of the Blaisun turbidites is today some 100 metres. Originally, they
1587 may have been at least 1/3 thicker.

1588

1589 3e. Pischa turbidite formation. Through the thickening of the individual
1590 limestone beds, with a concomitant increase in the thickness and the grain size of
1591 the conglomeratic bottom parts, a lightening of the colour to pale blue grey and
1592 the nearly complete disappearance of the shaly horizons, the Blaisun turbidites
1593 lead over to the Pischa turbidites over a distance of 50 m. Ignoring the
1594 differences just listed, the Pischa turbidites are no different from the Blaisun
1595 turbidites. In terms of the Bouma classification, they represent commonly
1596 truncated sequences of the type a-d. Only locally do black shaly horizons
1597 represent the e part.

1598

1599 Towards the top, the Pischa turbidites become rather suddenly thinner and
1600 shale-richer, resembling very much the Blaisun turbidites. That these upper
1601 horizons are younger than the thick, shale-poor Pischa turbidites is very clear;
1602 i.e., they cannot represent some sort of structural repetition of the Blaisun

formation. Within about 20 m from the last shale-poor Pische turbidite, the thickness of the individual turbidites reduces to about 20 cm and the section becomes extremely shale-rich. The colour of the limestones changes to dark grey to almost black with a brownish weathering. The shales remain black.

The thickness of the Pische turbidite formation is hard to estimate because of the intensive folding, but it is now on the order of some 60 m. Originally they must have been considerably thicker. Above them now is the lower sub-Silvretide slice across a very sharp tectonic contact.

Summary of the Ela sedimentation (Fig. 10). The descriptions given above present a rather simple sedimentological evolution for the part of the Ela Nappe in the Albula Pass region. It begins with the subaerial to very shallow marine Alv Breccia and the succeeding units up to the base of the Kieselkalk record the gradual subsidence of the ground. Then a rather rapid subsidence coeval with the world-wide sea-level rise leads to the Kieselkalk distal turbidite deposition. With the Fuora da l'Üertsch we have enlargement of grain size possibly indicating the rise of a fault block into high-energy environment. Then with the Blaisun this regime continues but then abruptly ceases. This cessation is possibly in part related to rising sea-levels, but its abruptness is possibly tectonically-controlled. Elsewhere, we know that the Ela Nappe sedimentation continued into the medial Cretaceous. In the Albula area I mapped, the Ela Nappe is tectonically

1625 truncated at the level of Middle, possibly at the lower levels of the Upper
1626 Jurassic.

1627

1628 Eberli (1985a, 1987, 1988) presented a much more meticulous study of the
1629 Jurassic turbidite sedimentation in the Albula Pass area, but our conclusions are
1630 essentially the same, as summarised in Eberli (1985a, 1987, 1988) fig. 16, except
1631 the more detailed field subdivision I present and the structural observations
1632 concerning thicknesses, which Eberli did not consider (see also Fig. 10 herein for
1633 a summary of my sedimentological results).

1634

1635 *Sub-Silvrettide Slices*

1636

1637 There are two Sub-Silvrettide Slices in the Albula area below Piz Kesch. Both
1638 consist of separate lensoid tectonic bodies resembling shreds carried along the
1639 bottom of the massive Silvretta Nappe.

1640

1641 1. Lower slice: the lower slice consists only of two rock types: dark grey to black,
1642 very richly fossiliferous, micritic limestones and black to grey calcareous shales.
1643 The macrofossils consist mainly of corals, brachiopods and numerous shell
1644 fragments. Under the microscope, the rock is seen locally to be a fossiliferous,
1645 laminated micrite and locally a recrystallised limestone with some allochthonous
1646 lithoclasts (mainly quartz and feldspar). These lithoclasts might indicate erosion

1647 down to pre-Alpine basement at the upthrown edges of rotated normal faults.
1648 Despite recrystallisation, some echinoderm crystals are still recognisable.

1649

1650 This rock is about 40 m thick now, but owing to extreme tectonisation and
1651 stylolite building, this figure is meaningless with respect to its original
1652 stratigraphic thickness.

1653

1654 The limestone coquinas gradually pass stratigraphically upwards (not obvious in
1655 my area) into calcareous shales. This entire assemblage is known in the
1656 Austroalpine stratigraphy as the Kössen Beds of Rhaetic age. Concerning the
1657 history of the development of the concept of the Kössen Beds and their great
1658 significance in correlating the Alpine sequences with those of the foreland, see
1659 •engör (2014).

1660

1661 2. Upper Slice. The upper slice consists entirely of very finely crystalline, locally
1662 chertified dolomite. Chertification seems to have produced dedolomite-like
1663 crystals. The entire rock is cut by coarsely crystalline, deformed calcite and
1664 quartz veins. Locally, veining has led to complete brecciation. At outcrop, the
1665 rock is a massive, blue-grey Hauptdolomit.

1666

1667 My subdivision largely agrees with that by Bearth et al. (1987) but is more
1668 detailed than theirs.

1669

1670 *Structural geology.* The Albula region, as comprehended to lie between the two
1671 crystalline nappes of the Err in the south and the Silvretta in the north for the
1672 purposes of this paper, essentially consists of two major nappes, the Albula and
1673 the Ela, each formed from the slicing of a single anticline and intervening slices
1674 torn from them during their emplacement. The surprising result of this
1675 investigation is that the two nappes originally formed parts of a single, *south-*
1676 *vergent*, nappe of the first genre, i.e., a recumbent anticline in which the
1677 amplitude is many times larger than the wavelength (see Termier 1911, p. 5).
1678 This anticline was essentially cut into two main pieces during the reversal of the
1679 movement and two essentially sledrunner nappes (genre of the second kind)
1680 originated. This firmly places the Ela into a Lower Austroalpine position and
1681 initially on top of the Err Nappe as I already pointed out in my dissertation in
1682 1982 (•engör, 1982; but without realising that the Albula and the Ela nappes
1683 were once one), contrary to the previous interpretations and to the scheme still
1684 depicted in Furrer et al. (2015, foldout plate, profiles 2 and 3), but, inexplicably,
1685 in agreement with the scheme in Furrer et al. (2015, fig. 12). The *Tectonic Map of*
1686 *Switzerland* at a scale of 1/500,000 (Bundesamt für Wasser und Geologie, 2005)
1687 correctly puts the Ela Nappe into the Lower Austroalpine. The movements were
1688 simple in principle, but the thrust faults on which they moved frequently were
1689 abandoned and new ones formed creating numerous slices between them and
1690 their bounding crystalline nappes of Err and Silvretta. That is why it is extremely

difficult to correlate "generations of structures" from one nappe to another or between the nappes and the slices or even between two slices. The narrative below first describes the structures and then presents my interpretation of their history.

The Supra-Err Slices

Four different generations of folds and two prominent foliations are the dominant structures of the Supra Err slices as they are exposed at the considerable outcrop just south of the small lake at the western end of Fig. 2 (see Fig. 5A). As I had to describe the structures of the other Supra-Err slices already in the stratigraphy section, I here confine myself to this main outcrop. The dominant structure at outcrop is a penetrative slaty cleavage in the Palombini shales and a very closely-spaced solution cleavage in the Aptychenkalk. Both of these have a steep northerly dip and are axial plane to two large folds (see Fig. 2). Finding mesoscopic folds is extremely difficult as they seem to be almost completely transposed. No primary bedding could be identified with certainty. Some "ghost" beds are like the ones sketched in Figs. 5A and C. The hinges of those few I was able to find are plotted on Schmidt's stereographic net in Fig. 5A and two examples are sketched in Figs. 5B and C. The axes of the F_1 folds generally trend W-NW and plunge westward following the plunges of the two main F_1 folds. The poles perpendicular to the plane of the S_1 foliation (Fig. 5A)

1713 define a pronounced N-S girdle, indicating the E-W trend of the F_2 folds. The
1714 actually observed F_2 fold axes corroborate this inference indicating a slight
1715 preference for northwesterly axes exactly as in the case of the F_1 folds. The
1716 associated second foliation is a crenulation cleavage (Fig. 11) with a spacing of
1717 about a cm, which, however, has rarely developed and I was able to measure
1718 only five orientations. Three of the measurements indicate gently SSW-dipping
1719 surfaces, whereas the other two dip gently northeastward. However, the gently
1720 SSW- to S- dipping surfaces must be predominant, because most of the F_2 folds I
1721 was able to see had Z shapes when viewed towards the W and had gently S-
1722 dipping axial planes.

1723

1724 The D_3 deformation generated conjugate kink bands. The orientation of their axes
1725 is E-W. These kinks take up roughly N-S shortening and are related to the last
1726 deformations seen in the area. They deform the S_2 crenulation cleavage.

1727

1728 An additional phase of deformation to those observed in the Supra-Err slices is
1729 the long wavelength undulations of the widespread S_1 surfaces. Fig. 12 shows
1730 one example of these undulations. In Fig. 5, these undulations are shown as F_4
1731 structures. In only a single instance did I see an intersection of two similar
1732 undulations, whereby one folded the other. It turns out that the younger
1733 undulation is the more widespread of the two (poles to the axes of which are
1734 shown in red in Fig. 5A). These undulations indicate a shortening of NW-SE

orientation and may well be a result of late right-lateral strike-slip motion along the Albula Pass region, which is the opposite of the sense suggested by Furrer et al. (2015, fig. 76).

Finally, an extremely interesting aspect of the structural geology of the Supra-Err slices is the widespread occurrence of two main sets of tension gashes opened across the S_1 foliation surfaces. These gashes are filled with calcite crystals. Most are antitaxial by anybody's definition, but at least one in the example shown in Fig. 13 is ataxial according to some, but syntaxial according to Bons (2000 also see Bons and Montenari 2005), because the crystals seem to have grown from one side to the other (inset in Fig. 13). Commonly, the first generation tension gashes formed perpendicular to a very fine crenulation lineation on the S_1 surface. This crenulation parallels the F_2 fold axes. The strikes of these older tension gashes are around 170 and they are commonly vertical or very close to it. Figure 5A shows poles to those tension gashes that could be measured as surfaces owing to the weathering out of their calcite filling (in red). It is clear that their strike varies around a N-S plane that is almost vertical. This is one of the very rare cases in nature that the direction of the σ_1 can be read directly out of the structure, because the fluid-filled gashes could not have sustained any shear stresses as they were filling. In this case the σ_1 had to have lain somewhere parallel with the plane of the gash.

1757 The second generation tension gashes are larger than those of the first generation
1758 and they form a nearly right angle with the second generation crenulations.
1759 When studied in detail, the orientation of crystal growth within the second
1760 generation shows that it was bent after the crystals had grown so as to be able to
1761 sustain a shear stress. The sense of motion along them is shown in Fig. 13 inset,
1762 and is consistent with the shortening indicated by the second crenulation. It may
1763 be that these tension gashes are coeval with the F_4 undulations (see Fig. 5A) and
1764 related to an E-W dextral strike-slip motion along the Albula Pass area.

1765

1766 Summary of the Supra-Err slices deformation. The Supra-Err slices were
1767 deformed by at least 3 generations of folds that have the same orientation; only a
1768 fourth generation very open folding has an orientation with fold axes trending
1769 about 90° to the older generations (Fig. 14A-D). I consider the first generation a
1770 result of a first thrusting, the direction of which I could not ascertain from this
1771 outcrop. Whether the first deformation has only one generation of folds or
1772 perhaps two (as the phacoidal cleavage leads one to suspect) could not be told
1773 because of the very intense shear. In the Ela Nappe, as we shall see below the F_1 's
1774 and the F_2 's could be separately mapped, yet they are entirely coaxial and have
1775 essentially the same style. But their shear seems less than the shear exhibited by
1776 the supra-Err slices. Fig. 14A shows a possibility, which is supported by the
1777 overall evolution of the area. But the F_2 folds suggest top to north transport (Fig.
1778 14B). The kink folds also indicate N-S shortening and they clearly postdate the F_2

1779 folds (Fig. 14C). They are possibly the continuation of the same deformation in a
1780 medium that lost much water and has become more brittle. Only the F_4 structures
1781 indicate roughly NW-SE shortening (Fig. 14D), possibly related to right-lateral
1782 shear along the Albula Pass that extends in an E-W direction. Froitzheim et al.
1783 (1994) postulated left-lateral shear along the Albula Pass, adopted by Furrer et al.
1784 (2015, fig. 16) but showed no evidence for it. My data indicate the opposite. A
1785 right-lateral shear along the Albula Pass is also compatible with the presence of
1786 an X-shear here related to the sinistral Engadine Fault Zone.

The Albula Nappe

1789
1790 Early normal faulting. The Albula Nappe, first defined by Zyndel (1912), is here
1791 used in a more restricted sense simply because of the more detailed structure
1792 that could be established than could be done by Zyndel (1912). His basic
1793 anticlinal structure, also given by Furrer et al. (2015, foldout plate, profile 3) in a
1794 much simplified manner, however, is fully corroborated. In fact, Fig. 2 shows a
1795 roughly westward plunging anticline with the Proliebas Assemblage at its core
1796 and upper Rauhewacke at its top. However, this anticline is itself folded so as to
1797 wrap the Middle Triassic dolomites and the upper Rauhewacke around the
1798 Funtaunas Assemblage (Fig. 2 and 6), a structure first established in this paper.
1799 But this wrapping shows that units have been eliminated from the section, where
1800 the Middle Triassic dolomites and the upper Rauhewacke come into direct contact

1801 with the Funtaunas Assemblage. Since this contact has been deformed during all
1802 subsequent deformations of the area it represents a contact that originated
1803 during the earliest deformation and I thus feel justified in seeing in it a normal
1804 fault contact that was active sometime after the deposition of the upper
1805 Rauhwacke. This is thus a normal fault of late Triassic to possibly earliest Jurassic
1806 age (Fig. 6) and related to the opening of the Alpine Neo-Tethys. When the
1807 subsequent deformations are undone palinspastically, the normal faults
1808 accomplishing the elision of units in the Albula Nappe turn out to be north-
1809 dipping normal faults (assuming no rotation around vertical axes of the entire
1810 Albula area). Although north- or west-dipping normal faults have often been
1811 mentioned in the literature, in my area I have been able to find only north-
1812 dipping normal faults.

1813

1814 Although the outcrop pattern clearly gives away the anticlinal structure of the
1815 Albula nappe because of its contrasting rock types, it has proven extremely
1816 difficult to find mesoscopic folds at outcrop. Of the F_1 folds I have only been able
1817 to find three to measure and they show westerly plunging axes in accordance
1818 with the westerly plunge of the main anticline of the Albula Nappe (Figs. 15 and
1819 16). By contrast the S_0 (bedding) surfaces are more common and they indicate
1820 considerable scatter with E- to NW-SE strikes and southwesterly dips
1821 predominating. The strike and dip measurements in the Middle Triassic
1822 dolomites have given me much trouble in the field, because it initially seemed

impossible to make sense of their orientations. I finally realised that these dolomites had probably burst into independent blocks during the folding giving rise to the aberrant strike and dip measurements because of the separate motions of the blocks (Fig. 15). Fig. 15A shows a hand specimen from the Upper Devonian deep-water nodular limestones in Istanbul, in which the limestone layers between the more plastic shale layers had burst into numerous "nodules", although the nodules here may have been predetermined by burrowing and carbonate segregation plus solution cleavage (Daniel Bernoulli, written communication, 20th April 2016; Bernoulli had examined the Istanbul outcrops together with me earlier) Some of these nodules formed because of boudinaging, others by snapping during folding. I imagine a similar thing must have happened to the Middle Triassic dolomites of the Albula Nappe, shown in Fig. 15B. That this inference was possible is entirely because of the preservation of the original bedding and the sedimentary laminae in the Middle Triassic dolomites.

The F_1 folds were all detected in the Funtana and the Proliebas clastic rocks. They are all isoclinal and have the prominent S_1 foliation associated with them (Fig. 15, A and B). The poles to the orientation of the S_1 foliation planes show a pronounced grouping around the north and south poles showing a weak girdle grouping and indicate folds with E-W axes and axial planes dipping to the north (Fig. 15). The S_1 assumes different characters in different rock types in which it has formed. The most common type is slaty cleavage occurring in pelites and

1845 fine-grained psammities. A coarser variety of this cleavage, which may be termed
1846 "domainal cleavage", occurs in psammities and fine-grained metaconglomerates.

1847

1848 The folding of the S_1 by the F_2 folds shows that the F_2 folds are coaxial with the
1849 F_1 's. The F_2 folds are more abundantly detectable at outcrop as shown in Fig. 16.

1850 The F_1 's seem to have a nearly perfect E-W orientation in the Albula Nappe as
1851 can be seen in the pole distribution of their axial foliation, S_2 in Fig. 15. That axial

1852 foliation is gently south-dipping crenulation cleavage. The wavelength and the
1853 amplitude of the crenulations vary from 0.5 cm to about 1 cm according to the

1854 rock type in which they occur. As a rule the metaconglomerates and the
1855 psammities have larger wavelengths and amplitudes than those in the finer

1856 grained psammities and the pelites. *Based only on my local observations so far, the D1*
1857 *and the D2 deformations in the Albula Nappe had north-south tectonic transport only,*

1858 contrary to the widespread opinion that *all* the Cretaceous nappe movements in
1859 Graubünden and Vorarlberg had an east-west to east-northeast-west-southwest

1860 tectonic transport. Whereas *many* structures may indeed have had east-to-west
1861 movement, this is perhaps not valid for *all* the structures and hat the kinematics

1862 in Graubünden may have been more complex than previously believed.

1863

1864 The third generation deformation affecting the Albula Nappe rocks has
1865 produced gently south-dipping thrust faults of limited displacement (Fig. 16C)

1866 and conjugate kink bands (Fig. 16D). That this phase may be related to the

1867 generation of very long wavelength and high amplitude folds affecting the entire
1868 Alpine edifice, creating essentially *plis de fond* of Argand (1924), I shall discuss
1869 later.

1870

1871 As opposed to the older rock types of the Albula Nappe, it is exceedingly
1872 difficult to say anything about the internal structure of the enormous mass of the
1873 upper Rauhewacke separating the Albula Nappe from the Ela Nappe with the
1874 intervening Infra-Ela slices, all of which are really various rock types of mainly of
1875 the Albula Nappe and only the Aptychenkalk floating in the upper Rauhewacke.
1876 The reason for this structural opaqueness of the upper Rauhewacke is the
1877 extensive micro-karstification that has given the rock a homogeneous
1878 appearance. A single fold was seen in it by me and a few measurable "ghost"
1879 bedding remnants.

1880

1881 That a number of thrust-bounded slices must be present in it (the infra-Ela slices)
1882 is obvious from the fact that it contains lensoid slices of both the lower (from the
1883 Albula Nappe) and the upper (Ela Nappe) edifices, but it has not been possible to
1884 follow any structures beyond the boundaries of such slices that could have
1885 indicated the location and the orientation of the major thrust faults along which
1886 the upper Rauhewacke rocks have moved and enveloped the other rock types.
1887 The fact that much of the upper Rauhewacke is actually a tectonic breccia suggests
1888 that the entire mass we now see must be riddled with numerous internal surfaces

of structural discontinuity and motion. But neither the rock type, nor the state of outcrop, nor indeed the chosen mapping scale would have been suitable to find and map them. Bearth et al. (1987) did make an attempt, but their mapping is so generalised and bereft of structural detail that I was unable to use it for my structural analysis.

Since the Infra-Ela slices are only inferred from scraps of Aptychenkalk and Hauptdolomite and substantial folds of the Funtaunas and Middle Triassic dolomites, I shall not describe them in any detail here. Whatever is known can be gathered from the Figs. 2 and 6. But I shall return to a consideration of them once I have summarised the structure of the Ela Nappe.

The Ela Nappe

Of all the tectonic units in the Albula Pass area, i.e. between the crystalline nappes of the Err and the Silvretta, Ela occupies by far the largest area (Fig. 2 and 6). Unlike the Albula Nappe and the Supra-Err slices, the Ela underwent no metamorphism and consists mainly of well-bedded calciturbidites providing superb younging evidence. Moreover, except for the southernmost parts, near its main basal thrust, the internal structure of the Ela Nappe is very well exposed in the wonderful sections of the Blaisun Ridge, the sedimentological history of which was studied by Eberli in detail (Eberli 1985a, 1988). By far the largest part

of its structure can be seen directly on such slopes so that it is unnecessary to resort to a detailed mesoscopic structural analysis to be able to outline the structure. Fig. 17A displays a representative selection of the structural data.

The Ela itself is the lower half of a nappe of the first genre, the upper half of which had been sliced off during the thrusting of the higher units, namely the sub-Silvrettide lenses and the giant torso of the Silvretta itself. This slicing must have taken place right across the Hauptdolomit as both below and above it are remnants of the Hauptdolomit. It is extremely unusual that a major thrust would prefer to cut across dolomites when less competent rock types are available to function as detachment horizons, but it is not unknown from elsewhere (e.g., Burchfiel et al. 1982).

The Ela nappe is a southerly overturned, upward-facing cascade of antiformal synclines, descending northward as its contained stratigraphic units become younger (Fig. 6). This result is in complete agreement with the conclusions of Froitzheim et al. (1994), who also concluded that the Ela is a south-facing nappe of the first genre. In Fig. 6 I have only depicted the main F_1 folds as the F_2 do nothing except to fold the F_1 folds in sympathy with the continuing southward thrusting. In Fig. 17A it is seen that the F_1 and F_2 are entirely coaxial and that the F_2 folded the S_1 around folds very similar to the F_1 folds. I do not even think that in the Ela Nappe, the F_1 and the F_2 represent different phases in sharp contrast to

1933 the supra-Err slices, where the F_1 's and the F_2 's can be clearly identified as
1934 belonging to two contrasting displacement and strain regimes; in the Ela, they
1935 are parts of one progressive deformation that first generated the F_1 folds; as they
1936 grew, they themselves began behaving like layers and became re-folded. In the
1937 Albula Nappe also, the F_1 and F_2 belonged to two different phases that had two
1938 different causes and that is why they had to be described and discussed
1939 separately; in the Ela that is not the case. Moreover the F_2 has developed only
1940 locally in the Ela Nappe, and only near the lower thrust contact in the Kieselkalk.
1941 Moreover, some of the very tight F_1 's may in fact have originated originally as
1942 slump folds. That their style is completely different from the much more
1943 common tectonic folds and that they are extremely rare (I have seen only one!)
1944 support my suspicion that they are not tectonic folds.

1945

1946 Figs. 17B and 17C show folded stylolites in the Ela Nappe. These stylolites were
1947 bedding-parallel structures and their folding and slicing I think faithfully reflects
1948 what the Ela internal structure looks like (Pumpelly's rule!). Especially Fig. 17C
1949 shows that the Ela Folds originated in a broad shear zone with top to south
1950 movement. Larger forms of these folds, that are actually upwards-facing
1951 antiformal synclines, are seen in the Blaisun east ridge.

1952

1953 The dominant mesoscopic structures of the Ela nappe are close to tight folds with
1954 a well-developed axial plane foliation, which is a spaced solution cleavage in the

1955 micrites of the calci-turbidites and in the sparites of the Kieselkalk, commonly
1956 with a spacing of 0.3 to 1 cm. It is anastomosing and bounds limestone lenses of
1957 1/15 aspect ratio. Locally, the solution cleavage takes a stylolitic form. In the
1958 shaly units of the turbidites, the axial cleavage assumes a slaty character. Around
1959 the fold hinges it generates marvellous pencils, of which the top of the Piz
1960 Blaisun has a particularly abundant supply.

1961

1962 Many stretched belemnites indicate that considerable E-W extension occurred
1963 during the formation of the Ela Nappe. These stretched fossils are parallel with
1964 the pencils of the fold hinges. Also, some rodding is seen along the fold hinges of
1965 the limestone-rich layers.

1966

1967 The entire Ela Nappe plunges eastward and its Hauptdolomit core appears in the
1968 Uertsch Massif at the crest of the Albula Pass. That core is in tectonic contact with
1969 the younger turbidites everywhere. This is probably in part because of the slicing
1970 of the entire Ela into two during the southward thrusting and in part because of
1971 the great competence difference between the massive dolomites and the much
1972 less competent turbidites. That thrusting, as we shall see below, also explains the
1973 origin of the sub-Silvrettide lenses.

1974

1975 The Blaisun Ridge exposes renewed folding around steeply south dipping axial
1976 planes. These folds may have been generated by the latest southerly movement

1977 of the Silvretta along the Engadine strike-slip fault (Trümpy 1977; Schmid and
1978 Froitzheim 1993). They are large-scale (tens of m) and have no associated
1979 microstructure.

1980

1981 The extremely thin slices of the Hauptdolomit at the base of the Ela Nappe both
1982 within the Infra-Ela slices and near the Ospiz directly in contact with the
1983 Kieselkalk and the Alv Breccia gives us an extremely interesting clue as to the
1984 origin of the Ela and the Albula Nappes, which I shall discuss in the penultimate
1985 section of this paper. In the next short section I describe the structure of the sub-
1986 Silvrettide lenses.

1987

1988 *The Sub-Silvrettide Lenses*

1989

1990 The Lower Slice. An entirely new tectonic unit was discovered in my area within
1991 the sub-Silvrettide lenses during a visit by my friend Dr. Gregor Eberli, who
1992 identified the shales and limestones I had separated immediately as the Kössen
1993 Beds (for the Kössen beds, see •engör 2014). These Kössen beds form a lower
1994 slice, previously not recognised in the Albula area (see Heierli 1955; Furrer et al.
1995 2015). The Rhaetian Kössen Beds have an extremely complex structure, the study
1996 of which is rendered difficult by the lack of sharp contrast in appearance
1997 between the fissile, shaly limestone coquinas and the carbonate-rich shales. What

1998 is depicted on Fig. 2 illustrates only a schematic outline of what I think the
1999 structure of the lower slice to be.

2000

2001 The most conspicuous mesoscopic structure of the lower slice is an ever-present
2002 slaty cleavage which becomes phacoidal in the shaly limestone coquinas. Only
2003 rarely does one see a mesoscopic fold associated with this S_1 cleavage. I was able
2004 to find seven such F_1 fold closures and their orientation is shown in Fig. 18.
2005 Without exception, they are all isoclinal. The similarity of the orientations and
2006 the styles of these folds to those of the F_1 's of the Ela Nappe is striking. I have
2007 little doubt that they originated together. The poles to the S_1 cleavage surfaces
2008 define a nice girdle orientated nearly N-S indicating that they were later re-
2009 folded. The F_2 folds are Z folds when viewed towards the west, although only
2010 rarely an S_2 fabric is seen associated with them. The probably formed during the
2011 northwards thrusting of the Silvretta above them.

2012

2013 All the structures of the lower slice of the sub-Silvrettide lenses are truncated
2014 either by the sharp basal thrust of the Silvretta Nappe or by that of the upper
2015 sub-Silvrettide lenses.

2016

2017 The Upper Slice. The upper sub-Silvrettide slice consists of a single rock type,
2018 that of the commonly massive, locally brecciated Hauptdolomit, preserved in
2019 disjointed lenses between the main body of the Silvretta Gneiss (actually its

2020 mylonitic base) and the lower sub-Silvrettide slice. The Hauptdolomit is in sharp
2021 thrust contact with both of its bounding units and the thrusts have very steep
2022 (50° to 70°) northerly dips. No structures were observed within the
2023 Hauptdolomit itself with the exception of locally very dense calcite veining and
2024 jointing.

2025

2026 Fig. 19 shows the contact relations of the westernmost of the upper slices of the
2027 sub-Silvrettide lenses within my mapping area. One very interesting feature of
2028 this figure is the small, late thrust fault it shows. This structure may be related to
2029 the latest, southerly motion of the Silvretta along the strike-slip Engadine Fault
2030 (Trümpy 1977; Schmid and Froitzheim 1993).

2031

2032 *The Silvretta Nappe*

2033

2034 As I said earlier, I did not consider in this study the crystalline parts of the
2035 Silvretta and the Err nappes. One thing that I must repeat mentioning however,
2036 is the metre-thick, greenish-grey, true mylonites that parallel the lower boundary
2037 thrust of the Silvretta. The thickness of this mylonite might indicate that the
2038 motion the Silvretta experienced with respect to wherever it was detached from
2039 before the formation of its basal brittle thrust may not have been very
2040 considerable, contrary to some earlier considerations (see Wibberley 2005). This
2041 encourages me to think that the Northern Calcareous Alps have covered a much

larger distance than the Silvretta and probably did not come from under it, but from above and south of it as Tollmann argued many years ago (Tollmann 1963) and as, at least in this sense, recently been corroborated (see the superb article by Schuster 2015 and the literature cited therein).

THE TECTONIC EVOLUTION OF THE ALBULA PASS REGION

Figure 20 illustrates my interpretation of the tectonic evolution of the Albula Pass region on the basis of the data reported in this paper. Fig. 20A shows the initial situation, colours of the layers and their numbers corresponding to those in Figs 2 and 6, i.e., they are the stratigraphic units seen in the Albula and the Elanappes. Two normal faults, now preserved in a folded state in the Albula Nappe cut the Triassic section, including the Hauptdolomit and the Kössen Beds (not shown here because of their thinness). The thin blue line topping the Hauptdolomit is the Alv Breccia and its accompanying rock types. These normal faults dip to the north (assuming no later rotation around vertical axes affected my area) and create a basin in which the turbidites of the Elanappe were deposited to the end of the early middle Jurassic. The faults may have continued moving while the Alv Breccia was being deposited and possibly also after until the Sinemurian. There is no evidence to say when they stopped until the radiolarites and the pelagic Aptychenkalk mantled everything in the early

2064 Cretaceous. The general tendency is to consider the end of rifting when ocean-
2065 floor spreading started in the Alpine Neo-Tethys during the medial Jurassic (e.g.,
2066 Laubscher and Bernoulli, 1977; Trümpy 1982; Schaltegger et al. 2002; Li et al.
2067 2013), because this is believed to be the case in the modern Atlantic-type
2068 continental margins. That consideration is incorrect as shown by the ongoing
2069 normal faulting along the Eritrean margin of the Red Sea creating earthquakes
2070 exceeding Mw=6 simultaneously with those along the spreading centre farther
2071 east (e.g. Al Ahmadi et al. 2013)! But a more serious problem is whether there
2072 was ever a well-defined spreading axis in the Alpine Tethyan subbasins such as
2073 the Piemontais and Valais. There are indeed pillow lavas in the Penninic Platta
2074 Nappe (Dietrich 1969, 1970, 1976), but work by Daniel Bernoulli and his students
2075 since the beginning of the eighties has shown that the Alpine ophiolites and their
2076 relationship to the continental rocks remind one of the geology of the ultra-slow
2077 margins from the Indian spreading centres where there is hardly any magmatism
2078 and that serpentinites form much of the ocean floor, raised there by extensional
2079 detachment faults (compare Dick et al. 2003). Therefore it is not possible to tell
2080 when the normal faulting ceased in the Alps except by the onset of radiolarite
2081 deposition followed by the mantling of the older structures by the coccolith
2082 muds of the Aptychenkalk. At least I have no way of knowing from the data
2083 from my mapping area in the Albula Pass when the normal faulting ceased until
2084 the deposition of the Aptychenkalk.
2085

Another criterion to establish the end of normal faulting in the Alps has been the sealing of fault contacts by younger sedimentary units, such as the Saluver Group of Middle Jurassic age overlying the extensional low-angle faults and fault rocks at Fuorcla Cotschna (see Fig. 1. 27b in Manatschal 1995). But such observations constrain only the age of the fault *segments* that they seal. Also, clasts of the fault rocks occurring along the low-angle extensional faults were found in the Jurassic Saluver Group (see Manatschal 1995, p. 60, for the detailed observations). However, finding the fault rock clasts belonging to one extensional fault does not necessarily mean that the fault as a whole had ceased its activity (see, for example the rolling hinge model of low-angle extensional fault generation in Hamilton 1987, p. 40 and Buck 1988). All such observations only constrain the ages of fault segments and not whole faults, so care must be taken in assigning age to extensional regimes. In the Sea of Marmara, for example, Le Pichon et al. (2014) found major faults, covered in one place by Pliocene sediments while at another they remained seismically active!

The second problem concerning the late Triassic-early Jurassic normal faults pertains to the direction of stretching. My structural data tell me that the normal faults, all preserved within the Albula Nappe, were north-dipping, assuming no rotations around vertical axes have since affected the Albula area. If the folding of the fine quartz veins are to be associated with this extensional phase, and not with the Alpine thrusting, then the stretching must have varied about a N-S

2108 direction (cf. •engör and Bozkurt 2013). The displacement of the Albula granite
2109 body between the Crasta Mora and the Piz Jenasch and Piz Calderas (Manatschal
2110 1995, p. 88), combined with the highly scattered displacement data (Manatschal
2111 1995, fig. 1. 45; for a somewhat cleaned-up version of the same data set see
2112 Manatschal and Nievergelt 1997, fig. 13) give a variable extension clearly
2113 implying hanging wall rotations similar to the ones we see today in western
2114 Turkey (see •engör 1987; Kissel et al. 1987; compare especially with fig. 9 in
2115 Manatschal et al. 2006 and fig. 5 in Massini et al. 2012; especially the fig. 9 of
2116 Manatschal et al. 2006 show a clear NW-SE stretching). But the boundaries of the
2117 differently-moving blocks have unfortunately not been established. It is
2118 impossible to evaluate the reliability of the shear-sense indicators where the total
2119 strain data are unavailable (e.g. Flinn 1994; also Hanmer and Passchier 1995,
2120 where they concede Flinn's main point that unless the total strain is known, the
2121 kinematic indicators are of limited value; also see •engör and Bozkurt 2013, fig.
2122 1). Taking the data presented in Manatschal (1995, fig. 1.45 and more so in
2123 Manatschal and Nievergelt 1997, fig. 13) at face value, however, one can argue
2124 that the *stretching was in fact northwest-southeast* (see Schmid et al. 1997, fig. 22-4,
2125 reproduced in both Froitzheim and Manatschal 1996, fig. 3 and Ziegler et al.
2126 1996, fig. 2; Masini 2011, Masini et al. 2011 and 2012, see especially the
2127 uncertainty in the direction of tectonic transport in fig. 5) *simultaneous with the*
2128 *folding of the main low angle extensional faults in a WSW-ESE dextral transtensional*
2129 *zone* (cf. Dewey 2002 and •engör 2014, for the distribution of strain and

displacement in zones of transtension). It is likely that these faults originated during the latest Triassic and continued to be used during the Jurassic.

The reason that I have dwelt on the direction of extension during the Triassic-Jurassic is that the Alpine extensional regime between the Triassic and the Jurassic underwent a complete reversal (from dextral transtension to sinistral transtension) and yet the same major extensional structures were able to serve the both systems with minor modification (Nurbike Sa•dıç et al. in preparation). This has not been sufficiently emphasised in the literature, but this is not the place to deal with this large-scale problem.

With the onset of the Palombini deposition during the Valanginian to the Albian interval I presume that some form of shortening had begun in the Alps, as this is also the time interval of the onset of deposition of the Rossfeld Beds in the northern Calcareous Alps. Schlager and Schöllenger (1973/74) have shown that the onset of the Rossfeld Beds deposition following the Schrambach Beds (=Aptychenkalk) occurred during the Valanginian-Hauterivian interval (Weber 1942; Plöckinger 1968) and marked a change from the coccolith-rich pelagic Aptychenkalk of the Schrambach Beds to the sandy-marly Rossfeld beds. In the Northern Calcareous Alps, the "grey", i.e., deep-water Aptychenkalk deposition was accompanied locally by the red Agatha Limestones indicating shallower swell facies. The Rossfeld beds are distal turbidites, just like the Palombini in the

2152 Albula area. This I interpret to be a sort of pre-flysch phase of deposition, when
2153 shortening-related structures in the Alps had just begun moving. It is in this
2154 connexion significant that in Rettenbach near Bad Ischl in Austria, the initial
2155 appearance of quartz detritus was accompanied by the first debris flows and
2156 synsedimentary fold structures indicating the origin of significant slopes
2157 (Schlager and Schöllenger 1973/74). It is also interesting to note here,
2158 however, that palaeogeographically, the occurrence of the Rossfeld clastics lies
2159 much farther south than the Palombini, as they belong to the Tirolikum, i.e., the
2160 main, central nappe of the Northern Calcareous Alps, lying south of the
2161 Bajuvarikum. In fact, Neubauer et al. (2000, 2007) would now consider the
2162 Rossfeld beds to belong to an entirely different ocean, the Meliata, which they
2163 consider closed during the early Cretaceous and they see the Rossfeld Beds as the
2164 initial trench sediments of a south dipping subduction zone south of the Meliata
2165 Ocean. If the Rossfeld deposition really began during the Valanginian-
2166 Hauterivian interval, then they could not have been a consequence of the
2167 beginning convergence between Africa and Europe, because the first
2168 Africa/Europe convergence in the Mediterranean Alpides is seen between the
2169 M0 (125 ma) and 34o (118 ma) magnetic anomalies, i.e., between the end of the
2170 Barremian and before the end of the Aptian (see Engör and Stock 2014, fig. 18),
2171 i.e., at least 4 Ma later than the end of the Hauterivian (International Commission
2172 on Stratigraphy 2014/02). In fact, some now believe that shortening in the
2173 Meliata Ocean had started already during the late Jurassic (Missioni and Gawlick

2011), although it seems difficult to distinguish salt-assisted gravity gliding (e.g., Tollmann 1981; Mandl 1982) and shortening-related nappe movements in the Hallstatt area. North of the Bajuvarikum lay the Middle Austroalpine nappes and only to their north were the Lower Austroalpine Nappes such as the Err, Ela and the Albula. Helmut J. Weissert's estimate (written comm., 2015; see above) of the age of the Palombini is alone compatible with the magnetic anomaly interval permitting the onset of the Alpidic shortening in the Alps. I thus consider it nearly contemporaneous but somewhat older than the Rossfeld deposition.

The Palombini deposition is the last sedimentary event in our area (except for Quaternary sedimentation). Sometime after that a single, south-vergent recumbent fold originated, a nappe of the first genre, that affected the entire sedimentary rock package in the Albula area. This nappe was thrust onto the Err crystalline rocks (Fig. 20B). The basis of this interpretation is the following observations:

1. According to my mapping, the stratigraphic content of the Albula Nappe ends where that of the Ela Nappe begins.

2. Both Nappes have the same structural styles, except the prominent F_2 of the Albula seems related to the overriding of it by the Ela. Whether it had " F_2 's"

2195 similar to those of the Ela cannot now be decided because of the extreme
2196 deformation and the unsuitable rock types.

2197

2198 3. Both nappes were originally south-vergent. Their reconstruction requires that
2199 the Ela Nappe sat to the south and in continuation of the Albula Nappe.

2200

2201 4. Ela Nappe sits directly on the Albula Nappe; all intermediary slices being torn
2202 either from one or the other.

2203

2204 The single nappe just mentioned I name the Elalbula Nappe (Fig. 20B) Figures 20
2205 C and D show how I think the Ela Nappe was torn from the Albula Nappe
2206 during the continuing southward thrusting.

2207

2208 One could ask whether similar upper-plate-vergent (often called "landward
2209 vergence" in the literature: e.g. Seely 1976) structures are seen in the presently
2210 active convergent systems. Seely has reported some examples from the present-
2211 day subduction zones and showed that such "anomalous vergences" occur where
2212 there is little friction at the base of the upper plate. In the Alpine Neo-Tethys (in
2213 our case the Piemontais Trough) the subducted sediments would have been the
2214 Bündnerschiefer, a pile of hemipelagic muds. But in this case what is probably
2215 more decisive would have been the north-dipping fabric generated by the earlier
2216 extensional phase. I have shown above that the normal faults I was able to map

in my area were north-dipping and the possible extensional structures immediately below them in the Proliebas Assemblage possibly had the same dip. But more importantly, the main detachments within the Err body in Graubünden (Froitzheim and Eberli 1990; Manatschal 1995) may have had northwest rather than purely west dips as has been often argued.

It is very likely that the Elalbula Nappe was thrust southward following this north-northwest dipping fabric, *but not following the individual faults*. This is extremely important! As Froitzheim and Eberli (1990) already pointed out, the thrust faults related to the Alpidic shortening very rarely followed the normal faults of the extensional phase(s). In my area, I cannot demonstrate a single instance where they did follow the old faults, but the two folded and imbricated normal faults clearly show where the thrusts did not follow them. This shows that vergence in orogens is dictated, among others by the pre-existing penetrative fabric in the rocks, rather than individual faults.

I cannot tell from the evidence in my area how long the southward thrusting lasted and when exactly the Ela and the Albula nappes were separated (Figs. 20C D and E), but it is clear that when the Silvretta was thrust northward during the Eocene, it overtook and reformed the south-vergent structures (Fig. 20F). The main F_2 fold seen in the Albula Nappe was a product of this phase, so was the thrusting of the Ela on top of the Albula. It also seems that the Ela Nappe had

2239 probably been dismembered already in the south-vergent motion at least into
2240 two main pieces along the Hauptdolomit (Fig. 20E) and the Silvretta thrusting
2241 during the north-vergent motion seems to have pushed the upper part of the Ela
2242 outside the Albula area and came to rest on the lower part of the Ela. This is the
2243 most likely explanation for the existence of the sub-Silvrettide lenses on top of
2244 the Ela Nappe. During the Silvretta thrusting, the Ela folds became re-sheared as
2245 shown in Fig. 20F, possibly creating their final mesoscopic structures. An
2246 analysis of the fold distortion of the kind suggested by Alvarez (1978) on the Ela
2247 folds might yield the direction of final re-shearing, but time shortage did not
2248 allow me to undertake it.

2249

2250 Fig. 20G shows in a highly schematic fashion the final geometry of the structural
2251 elements in the Albula Pass area. A final refolding of the Ela folds may have been
2252 related to the southwesterly motion of the Silvretta along the Engadine Line
2253 (Trümpy 1977; Schmid and Froitzheim 1993), but I did not analyse it. The entire
2254 steepening of the area was related to the late Alpine folding as shown in Fig. 23
2255 H. This "folding" as accompanied by much thrusting under the Lepontine area
2256 both from the north, depressing the Molassebasin and from the south, depressing
2257 the Po Basin. This was "basement folding" in the sense of Argand (1924),
2258 whereby much thrusting accompanies and causes the depressions and the uplifts
2259 on a "fold wave."

2260

2261 Figures 21 A-F show a highly schematised series of cross-sections showing the
2262 tectonic evolution of the Albula area. *Both geographic orientation and the time each*
2263 *cross-section represents have been deliberately omitted.*

2264

2265 Fig. 21A, inspired by Laubscher and Bernoulli's (1977) fig. 4, shows an Atlantic-
2266 type continental margin sometime during the late Permian-early Triassic in an
2267 orientation that could be anywhere from north-south (north being on the left) to
2268 northwest-southeast. The red line shows the limit of the future extension. In Fig.
2269 21B extension not only has started, but also progressed to the point of just
2270 generating a mafic to ultramafic basement for the Piemontais ocean (see esp.
2271 Trümpy 1985; Weissert and Bernoulli 1985; Bernoulli et al. 1990, 2001; Froitzheim
2272 and Manatschal 1996; Desmurs et al. 2001; Schaltegger et al. 2002; Manatschal et
2273 al. 2003; Ferrando et al. 2004; Li et al. 2013; Decarlis et al. 2015). This must have
2274 been sometime during the earliest Liassic. The orientation of the section can be
2275 just like the previous one. The Triassic dextral transtension has now already
2276 given place to Jurassic sinistral transtension as a consequence of the opening of
2277 the Atlantic Ocean. At this time the Briançonnais continental fragment had
2278 already begun to separate from Europe along a major extensional system: at its
2279 footwall the Allemanic High of Trümpy (1949) had formed to the north and the
2280 Valais trough to the south (see Decarlis et al. 2015 for the most recent
2281 interpretation of this phase). Notice in these two figures that all the major normal

2282 fault systems are south or south-southeast-dipping. They may have acquired it
2283 from the previous Palaeo-Tethyan Atlantic-type continental margin or even from
2284 the earlier orogenies here (Pan-African and Hercynian).

2285

2286 In Fig. 20C the extension has reached its apogee and possibly stopped. However,
2287 it is hard to say when extension stopped in the Alps before the northerly swing
2288 of Africa 125 ma ago, because until then the regime was one of transtension in
2289 which local extension proceeded with local compression simultaneously (Dewey
2290 2002), a fact hardly ever taken into account in the Alpine literature. Only after
2291 Africa began approaching Europe can one really talk about the onset of Alpidic
2292 shortening *per se* and that happened only after 125 ma ago (cf. Engöör and Stock,
2293 2014). The earlier shortening events may well have been still related to
2294 transtension.

2295

2296 The question now becomes when did the subduction begin? The onset of clastic
2297 sedimentation of the Palombini between the Valanginian and the Aptian is
2298 hardly a help, as topographic prominences can easily be created by other
2299 mechanisms in various tectonic environments. The most secure evidence is the
2300 presence of high-pressure/low temperature rocks. The oldest ones from the
2301 Graubünden area have been dated to 90 ma from the Platta Nappe and the
2302 Malenco Fragment (Bousquet et al. 2008; see also the maximum 89 ma age in the
2303 north Penninic flysch zone in Vorarlberg: Winkler and Bernoulli 1986). This gives

us about 35 ma between the onset of the major Alpidic shortening in the Alps deduced from the magnetic stripes and fracture zone trends in the Atlantic Ocean and the first high pressure/low temperature metamorphism. These are the outer limits of a time interval, in which the Alpine subduction could have started any time. We must bear in mind, however, that the first chromite grains appear in the Penninic flysch deposits already in the Cenomanian and even earlier in the Northern Calcareous Alps (Woletz 1963, 1967), although it is impossible to tell which ultramafic bodies were supplying these grains. If they were Alpine, at least an obduction of sorts would have to be assumed already in the early Cretaceous in the Alps. But the source could easily have been the now-obliterated Meliata Ocean also, although my recent visit to the alleged Meliata outcrops in Austria has not encouraged me to see an "ocean" in them. Also the intrusion of the basanitic Ehrwaldite dykes in the Northern Calcareous Alps seem to preclude any subduction beneath them during the Albian (Trommsdorff et al. 1990; Froitzheim et al. 1996). Fig. 20 D-E shows my interpretation. I assume that before subduction began, compression may have deformed the southern continental margin of the Alpine Neo-Tethys, as continental rocks would be more susceptible to deform than ultramafic rocks under the same stresses simply because they have more quartz and feldspar. This deformation may have taken the form of south-vergent nappes using the pre-existing fabric here. Why that fabric should have existed there and not the dominant south-dipping fabric farther south is simply a deduction from observations. In any case that south-

2326 vergent deformation created first the Elalbula Nappe and then divided it into the
2327 more (now) northerly Ela and the more southerly Albula. All of this may have
2328 taken place before subduction started or concurrently with its inception
2329 sometime before the Coniacian. Only during the Eocene, the Middle
2330 Austroalpine crystalline sheets were torn from their more southerly positions
2331 and thrust northward. It was at this time when the Silvretta overrode and
2332 reformed the previously south-vergent nappes. It was at this time also when
2333 the Ela was thrust on top of the Albula and the Silvretta on top of all the rest.
2334 Before I end this discussion let me enter here a disclaimer. All the story I told in
2335 this section is based on a tiny area in the Albula Pass plus the evidence that
2336 supports the story that can be read there. This story does not take into account
2337 possible rotations of moving nappe fragments around vertical axes during the
2338 Cretaceous or during the Eocene. It is thus a "naïve" interpretation, told only to
2339 summarise the Albula story. A more realistic story will only be possible when we
2340 have a lot more strain data from this part of the Alps.

2341

2342 CONCLUSIONS

2343

2344 Although the Alpine evolution was dominated by transtensional events until the
2345 end of the Barremian in the area of the canton of Graubünden in Switzerland,
2346 creating a bewildering variety of structures, I have found no evidence in the
2347 Albula area for any east-west transport until very late in the evolution of the area

when possible east-west strike-slip motion may have imposed an east-west transport here as a consequence of the sinistral motion on the southwest-northeast striking Engadine strike-slip fault (Trümpy 1977; Schmid and Froitzheim 1993). It seems that interpretations on east-west tectonic transport during the extensional phase come mainly from scattered fabric data and from east-west trending large-scale folds interpreted as normal fault channelures. If they are considered actual folds created by shortening already during the late Triassic-early Jurassic interval in the context of east-west orientated transtension, then the pure stretching direction becomes northwest-southeast as already surmised by Laubscher as early as 1975 (see his figs. 4 and 5), later corroborated by Kelts (1981) Lemoine and Trümpy (1987) and Trümpy (1988), as also shown most recently by Masini (2011), Masini et al. (2011 and 2012).

This makes it puzzling as to why the dominant north vergence of the entire orogen (taking into account the late origin of the bend of the western Alps) should be so uniform as already appreciated by Suess (1875), because there was little subduction (see Stampfli and Marthaler 1990, pp. 177-178, for a good discussion concerning the amount of subduction in the Alps) and even what little subduction existed was probably dominated by the pre-existing structures in the sense that it originated as a part of the dominant north-vergence. The purpose of this paper was to show that a demonstrated, local, north-dipping, fabric in a limited area in eastern Switzerland gave rise to south-vergent structures very

2370 early in the Alpine evolution, which were later overwhelmed and reformed by
2371 the large, north-vergent nappes. The presence of a possible south-dipping fabric
2372 in the Austroalpine basement related to an Atlantic-type continental margin in
2373 the northeastern margin of the Palaeo-Tethys or an even earlier Hercynian or
2374 Pan-African fabric may have guided these major structures as the local north-
2375 dipping fabric did in the opposite sense in the Albula area. This inference is
2376 based on the presence of the Palaeo-Tethyan Atlantic-type continental margin
2377 (Laubscher and Bernoulli 1977, fig. 4; •engör and Atayman 2009) and the close
2378 spacial association of the extensional north-dipping structures in the Albula area
2379 and the early south-vergent nappe formation in the same place. At this stage that
2380 is all one can say.

2381

2382 Manatschal et al. (2015) discussed the inheritance problem from the viewpoint of
2383 the Alpine rifting only, but considered structural, compositional and thermal
2384 inheritance. They concluded that the inherited structures did not control
2385 significantly the location of the rifting. But in their consideration of the
2386 Hercynian and the post-Hercynian structures they did not consider the immense
2387 strike-slip systems that slashed through the future Alpine edifice all the way
2388 from the external massifs to the Southern Alps in a direction parallel with the
2389 long axis of the later Alpine Neo-Tethys (the orientation of these immense shear
2390 zones is parallel with the trend illustrated in fig. 8 of von Raumer 1998; cf.
2391 •engör and Atayman 2009; •engör 2013). These lithospheric-scale structures do

2392 seem to have controlled the location of the rifting considering the multiple rifting
2393 loci in the Alps (Meliata-Hallstatt, Valais, Piemonte oceans) not only structurally,
2394 but also thermally and compositionally. As I mentioned above, the Alpine rifting
2395 in the Alps itself was not a clean east-west stretching as implied in Weissert and
2396 Bernoulli (1985), but a transtension in which shortening and extensional events,
2397 together with a strong constrictional strain, dominated the picture (as Argand
2398 rather prophetically mentioned in 1934). It will not do to carry northwards into
2399 the Alps the well-established, roughly east-west extension of the Apennines and
2400 even of the Southern Alps. Kelts' (1981) model of a Gulf of California-type mini
2401 ocean is a better model, but it needs to be considered in terms of the peculiarities
2402 of the strain evolution in transtensional belts (Dewey 2002) without the
2403 formation of discrete spreading centres in a very slowly-spreading sea-floor (R.
2404 Trümpy, oral communication, 1975¹ corroborated by Daniel Bernoulli and his
2405 associates and students).

2406
2407 Despite the immense amount of work done on the Alps, we still do not have a
2408 good picture of the evolution of strain in its constituent rocks, except in the
2409 Helvetic Nappes of Switzerland (e.g., Ramsay 1981; Pfiffner et al. 2011). Spot
2410 kinematic data are useful when available for narrow zones, essentially surfaces,

¹ This was a talk Rudolf Trümpy gave in the Hans Cloos-Kolloquium in 1975 in Bonn, in which he expressed doubt as to whether a well-developed mid-oceanic ridge had ever been present in the Alps. His reason was that the Alpine ophiolites nowhere displayed Penrose-Conference-type full ocean-floor sections. He further pointed out that the Alpine oceans may have been too small to develop full-fledged spreading centres. Regrettably, he could not be persuaded to publish his ideas concerning this issue.

between very large entities such as entire plates, as have been shown by the use of slip vectors in testing Wilson's hypothesis of plate tectonics (Wilson 1965) by McKenzie and Parker (1967) and Sykes (1967), but not in areas of complex deformation where different movement directions inevitably form in the same deforming body during its change of shape in a single episode of deformation. In such areas it is the analysis of strain paths that give away the tectonic evolution.

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FIGURE CAPTIONS

Fig. 1. Location of the Albula Pass area (boxed) in the overall tectonic framework of eastern Switzerland and western Austria (copied from the *Tektonische Karte der Schweiz 1:500 000*, published by the Bundesamt für Wasser und Geologie, 2005). The relevant unit names have been Englished by me.

Fig. 2. The geological map of the part of the Albula region reported in this paper.

Fig. 3. Simplified and schematised stratigraphic synopsis of the area shown in Fig. 2.

Fig. 4. Structure of the lower Mesozoic rocks of the supra-Err slices. See text for details.

Fig. 5. A. Orientation information on the upper Mesozoic rocks of the supra-Err slices shown on a lower hemisphere projection of a Schmidt net. All orientation data in the following figures have been plotted on the same net. B. Outcrop sketch of the supra-Err slices south of the Ospiz showing folded layers. The outcrop width shown is about 5 m. C. On the left, the light green are Emmat Shale layers (Palombini) and darker red are chert layers in the Aptychenkalk. On the right are ghost folds and the phacoidal cleavage of the Aptychenkalk in the outcrop south of the Ospitz. Thin green lines are my interpretation of the geometry of the folds. What is actually observed are the chert boudins shown as green patches.

Fig. 6. A schematised geological cross-section across the area shown in Fig. 2. The F_2 structures in the Ela nappe are not shown because they are coaxial with the F_1 's and their depiction would have greatly complicated the picture while adding little extra information.

Fig. 7. A typical Rauhwaacke hand specimen from the area shown in Fig. 2. Notice the breccia aspect of the rock, which is actually a fault rock.

Fig. 8. A typical hand specimen of the Ela Nappe Kieselkalk. Note the strong layering parallel with a foliation. This rock is a distal turbidite.

Fig. 9. Across-strike shortening of the Kieselkalk. The shortening in this sample is almost 50% as shown by the folded calcite veins.

Fig. 10. Simplified sedimentological depiction of the Allgäuschiefer (Eberli's, 1985a, Allgäu Formation). The long triangles indicate grain size. The grain size first coarsens and then re-fines. For more detail see Eberli (1985a, 1985b, 1987, 1988). The letters indicate Bouma's (1962) turbidite sections.

Fig. 11. Sketch of the style of the S_2 crenulation cleavage in the supra-Err slices in the outcrop south of the Ospiz.

Fig. 12. An example of the F_4 undulations in the supra-Err slices in the outcrop south of the Ospiz.

Fig. 13. The tension gashes and their internal deformation in the supra-Err slices (in this case the Palombini shales) in the outcrop south of the Ospiz. They must be late as they show no folding and seem to cut all the previous structures, but I cannot tell whether they were coeval with the F_4 .

Fig. 14A. The origin of the F_1 structures of the supra Err slices. Whether these also include coaxial F_2 's as in the Ela Nappe (see below) I could not tell because of the intense shear. At this stage no tension gashes must have formed, as we see no folded examples (see B'). B. The F_2 folds probably formed during the top to the north deformation. Again at this stage probably no tension gashes formed.

Fig. 14C. The origin of the kink folds and kink bands (F_3) in the supra-Err slices.

Fig. 14D. The origin of the broad undulations (F_4) in the supra-Err slices.

Fig. 15. Orientation data from the structures in the Albula Nappe. A shows a hand specimen from the Upper Devonian pelagic shales and limestones in Istanbul showing the manner in which the limestone layers were disrupted during the deformation. B is a possible construction of similar blocks in the Middle Triassic dolomites of the Albula Nappe hypothesised to explain the aberrant strike orientations in a unit where there is no problem of telling where the bedding is.

Fig. 16. Mesoscopic structural styles from the Proliebas assemblage in the Albula Nappe. They are not entirely unlike those seen in the supra-Err slices.

Fig. 17. A. Orientation data for the structures observed in the Ela Nappe. B and C show the fold style shown by the deformed bedding-parallel stylolite layers. Especially C shows that there was a very considerable simple shear component in the origin of the Ela structures. D is a view of the Blaisun ridge from the east-northeast showing the mesoscopic folds of the F_1 generation. The thin brown lines represent the shales whereas the thicker grey bands stand for calcarenitic turbidite sections.

Fig. 18. Orientation data for the structures observed in the sub-Silvrettide lenses.

Fig. 19. Small thrust offsetting the main thrust contact between the lower and the upper slices of the sub-Silvrettide lenses.

Fig. 20A. Highly schematised cross sections showing the structural evolution of the Albula area shown in Fig. 2. A shows the situation in the early Jurassic when normal faults have extended the rocks exposed in my area. The red lines show the normal faults. I have drawn them planar here to follow up their evolution more easily, but they were most likely listric. All the colours and the numbers

correspond to the units shown in Fig. 2. In all the sections of Fig. 20, north is always to the right. B is the first phase of deformation and the formation of the first genre Elalbula Nappe with a south vergence. C and D are phases when the Ela and the Albula became separated from one another and the infra-Ela slices began to be individualised. All of this was most likely during the medial to late Cretaceous. In E, Ela itself become separated into at least two, but possibly more pieces. The main thrust fault accomplishing this cuts right across the Hauptdolomit, possibly along more than one surface, thereby isolating the future sub-Silvrettide lenses. This was still during the Cretaceous. In F, we are already in the Eocene and the north-vergent deformation is in full swing, bringing the Silvretta on top of the sub-Silvrettide Lenses and the lower half of the Ela and dragging them right across the Albula Nappe, which itself becomes dragged along the Err, trapping beneath some of the outermost layers of the old Elalbula Nappe, thereby creating the supra-Err slices. G shows in a schematised fashion the present stacking order and H the actual geometry the nappe stack now has after a major folding of the entire Alpine edifice, raising the "root zone" some 25 km in the Lepontine culmination and depressing the internal most parts of the Alps down to some 9 km in the Po Basin in Italy, forming the hinterland basin of the Alps (and the foreland basin of the Apennines as already pointed out by Suess in 1883). This is basement folding in a grand style in the sense of Argand (1924) and is the last act in the story of the Alps.

Fig. 21. Highly schematic cross-section showing the evolution of the Albula area at the scale of the entire Alpine edifice. Explanation is in the text.

STRATIGRAPHIC SYNOPSIS OF THE ALBULA PASS AREA

S

N

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Reduced bedding-perpendicular shortening
because of the resistance offered by the already folded calcite veins

S_1 surface:
Strike: 80
Dip: 60N

Larger scale second crenulations
(Pitch: 50W)

First crenulations
(Pitch: 0)

Antitaxial growth

Second generation tension gashes
(Str: 35, Dip: 50 SE)

First generation tension gashes
(Str: 170, Dip: about 90)

Fold hinge

A = A'

Calcite Filling (ataxial growth)
(milky white to orange-brown)

A

S

N

B

Ela Sedimentary Body carrying
the Silvretta on its back

D

Draft

SUBSILVRETTIDE LENSES

SSW

NNE

SILVRETTA NAPPE

Hauptdolomit (massive)

Kössen Beds

Greenish gneiss
(No good outcrops for
structural observations)

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