Pre-Alpine fold interference patterns in the north-eastern Oetztal-Stubai-Complex (Tyrol, Austria)

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Abstract

Structural geological investigations near the base of the Brenner Mesozoic (BM) of the northeastern Oetztal-Stubai-Complex (OSC) showed evidence for polyphase Variscan folding of the OSC in comparison to Alpine structures of its Permomesozoic cover (BM). Local remapping in the Kalkkoegel area (northern BM) revealed an unknown first isoclinal folding event (D_1) with S- and NE-orientated folds that formed a penetrative schistosity subparallel to the bedding ($S_{0/1}$). Large-scale open folding with ESE- and WNW-trending fold hinges overprinted this $S_{0/1}$-foliation with the development of a S_2-foliation and resulted in regional-scale intermediate Type 2/Type 3 fold interference patterns. Reverse modelling of these “elongated mushroom-shaped” geometries led to a newly established simple pseudo-stratigraphy, with meta-igneous rocks intruding into paragneisses overlain by mica schists. After Permomesozoic strata (BM) were deposited discordantly on top of the OSC, both were overprinted by ductile and brittle deformation during the Cretaceous (D_3 & D_4) and Cenozoic (D_5 & D_6) Alpine orogenies. Brittle deformation is still ongoing, as displayed by seismic activity related to the northward movement of the Southalpine indenter.

Strukturgeologische Untersuchungen im Nordosten des Ötztal-Stubai-Komplexes (OSC) an der Basis des Brennermesozoikums (BM) lieferten neue Einblicke in die duktile polyphase variszische Deformationsgeschichte im OSC durch Vergleich mit alpidischen Strukturen in den überlagernden permomesozoischen Decksedimenten (BM). Durch lokale Kartierungen im Bereich der Kalkkögel (nördliches BM) konnte eine bisher unbekannte erste isoklinale Verfaltung (D_1) mit S- und NE-orientierten Falten festgestellt werden, die zur Bildung einer penetrativen Schieferung subparallel zur sedimentären Schichtung ($S_{0/1}$) führte. Eine zweite großmassstäbliche offene ESE-orientierte Verfaltung überprägte diese $S_{0/1}$-Schieferung mit der Bildung einer S_2-Schieferung und führte zur Bildung von intermediären Typ 2/Typ 3 Faltenüberprägungen. Reverse modelling ermöglichte die Interpretation dieser „gestreckten pilzförmigen“ Muster und die Einführung einer Pseudo-Stratigraphie mit metamorphen Plutoniten/Vulkaniten an der Basis, die in Paragneisse intrudiert sind und von Glimmerschiefern überlagert werden. Nach der diskordanten Sedimentation permomesozoischer Einheiten (BM) auf dem OSC, wurden beide Einheiten während beiden alpidischen Orogenesen in der Kreide (D_3 & D_4) und im Känozoikum (D_5 & D_6) duktil und spröd deformiert. Die anhaltende spröde Deformation durch die nordwärts-gerichtete Bewegung des Südalpinen Indenters ist anhand seismischer Aktivitäten immer noch feststellbar.

1. Introduction

The Oetzal Stubai Complex (OSC) of Tyrol, Austria, has been intensively investigated with a petrographic, petrological and geochronological focus (e.g. Miller, 1970; Hoernes and Hoffer, 1973; Veltman, 1986; Söllner and Hansen, 1987; Miller and Thöni, 1995; Klötzli-Chowanetz et al., 1997; Thöni, 1999, Tropper and Recheis, 2003). While petrological highlights such as migmatites, eclogites and mica schists attracted numerous scientists, the rather monotonous but widespread paragneisses of the OSC were hardly ever studied. Since the classic works on the OSC by Tollmann (1963, 1977), Schmidegg (1964), Purtscheller (1971, 1978), Thöni (1980, 1981) and Hoinkes et al. (1982) general structural investigations addressing the entire OSC are lacking. Structural and tectonic aspects of the OSC have mostly been carried out within relatively local geological contexts (e.g. Schmidegg, 1964; Förster, 1967; Förster and Schmitz-Wiechowski, 1970; van Gool et al., 1987; Fügenschuh et al., 2000; Sölva et al., 2005). In this investigation, we attempt to decipher the polyphase structural evolution of the northeastern part of the OSC, with a special focus on pre-Alpine folding events, and we compare our findings to those of Alpine deformation events recorded in its Permomesozoic cover, the Brenner Mesozoic (BM).

This study is based on regional mapping at the scale of 1:10,000 in the Kalkkoegel area in the northeasternmost portion of the OSC and BM. The post-Variscan age of the cover units enabled us to distinguish pre-Alpine structures in the OSC from Alpine structures in the BM. With the integration of a simplified version of the geological map sheet “Oetzthal” of Hammer (1929), local structural observations were interpreted in the context of a regional-scale fold interference pattern.

2. Geological Setting

The OSC is a basement complex located in the central Alps (Fig. 1) and is tectonically attributed to the Oetztal-Bundschuh-Complex, with the OSC west and the Bundschuh-Complex east of the Tauern window (Schmid et al., 2004). Although this attribution is still controversial (Neubauer et al., 2007), we assume that within the context of Alpine nappe stacking both complexes have similar positions. Following the palinspastic restoration of Frisch et al. (1998), both these complexes were connected prior to the eastern lateral escape of the Eastern Alps towards the retreating subduction zone in the Carpathians (Royden et al., 1982) in response to the northward movement of the Southalpine indenter and the Miocene exhumation of the Tauern window.

In the northernmost part of the BM, (i.e. the Kalkkoegel area) the stratigraphy encompasses a sequence of transgressive sediments deposited onto a paleo-weathering horizon of the uppermost OSC (Krois et al., 1990). The deposition was initiated by Permo-Triassic clastics of the Alpine Verrucano, followed by dolomites and marls of the Virgloria Formation, Reifling Formation and Partnach Formation that interfinger with dolomites of the Wetterstein Formation. These successions are sealed by the Raibl shales, and the Norian Hauptdolomite completes the sequence with dolostone units (Rockenschaub et al., 2003; Brandner et al., 2003; Brandner and Reiter, 2004). The hanging wall units of the W-dipping Brenner fault zone are tilted towards the east and thus preserved the parautochthonous BM (Köhler, 1978; Krois, 1989) from erosion in the vicinity of the fault zone in the eastern part of the OSC.

The OSC consists of metamorphic rocks with various protoliths: orthogneisses of plutonic origin, amphibolites of volcanic and plutonic origin, and metapelites and metapsammites of sedimentary origin. Paragneisses are considered to be the oldest rocks in the OSC with a mean crustal residence age of ca. 1.5 Ga (Schweigl, 1995) to 1.6 Ga (Thöni, 1999). Meta-igneous rocks, which intruded the meta-sediments, have protolith ages ranging from 520-530 Ma for mafic intrusions (Schweigl, 1995) and 485 ± 3 Ma for intrusions with granitic compositions (Bernhard et al., 1996). Paleogeographically, the OSC formed part of the Apulian margin and is tectonically attributed to the Upper Austroalpine nappe complex (e.g. Schuster and Frank, 1999; Schmid et al., 2004), which formed through WNW-directed thrusting during the Cretaceous (Eoalpine) orogeny (Frank et al., 1987; Froitzheim et al., 1994).

The OSC is entirely fault bounded, with the different faults having different kinematics and ages (Fig. 1). They consist of the Inntal fault to the north (Ortner et al., 2006), the Brenner fault zone to the east (Fügenschuh, 1995), the Schneeberg fault zone and Vinschgau shear zone to the south (Flöss, 2009; Speckbacher, 2009; Sölva et al., 2005), the Schlinig fault to the west (Schmid and Haas, 1989) and the Engadine line to the northwest (Schmid and Froitzheim, 1993).

3. Polymetamorphism

The OSC is well known for its polymetamorphic evolution (i.a. Hoinkes et al., 1982; Thöni, 1999) with the youngest metamorphic event (i.e. Eoalpine orogeny) being of Cretaceous age (Frank et al., 1987), which reaches lower greenschist facies in the NW corner of the OSC and increases towards the SE to upper amphibolite grade conditions, with eclogite facies conditions documented in the underlying Texel complex (Thöni, 1980, 1981; Hoinkes et al., 1991; Konzett and Hoinkes, 1996). This regional variation in metamorphism is reflected in an increase from 445 °C in the north to 530 °C in the south within the BM (Hoernes and Friedrichsen, 1978; Tessadri, 1981; Dietrich, 1983). Eoalpine metamorphism within the Kalkkoegel area reached temperatures of about 450 °C (Hoernes and Friedrichsen, 1978) whereas temperatures during the pre-Alpine metamorphic event were substantially higher (ca. 600 °C: Hoernes and Friedrichsen, 1978).

Purtscheller (1978) documented a Variscan metamorphic zonation in the OSC based on the distribution of aluminosilicates (Fig.1). However, this zonation is questioned by Hoinkes and Thöni (1993) and Tropper and Recheis (2003), who propose that although the zonation is consistent with the metamorphic gradient between the Variscan and Alpine orogenies, the aluminosilicates do not necessarily represent a single metamorphic event. Schuster et al. (2004) estimated the age of the Variscan thermal peak at 340 Ma and Variscan cooling ages at ca. 310 Ma (Miller and Thöni, 1995; Neubauer et al., 1999; Thöni, 1999) with maximum Variscan temperatures of 670 °C in the northern area of the OSC (Hoinkes et al., 1982).

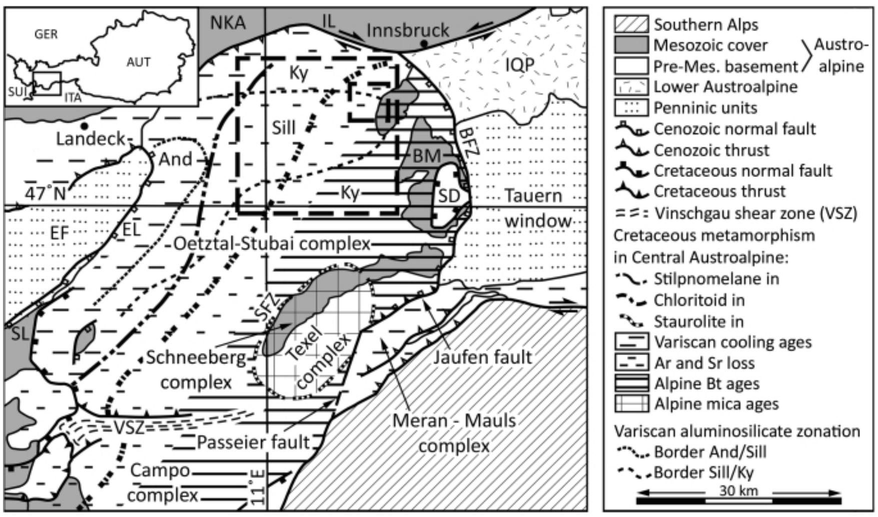

Figure 1: Tectonic overview of the Oetztal-Stubai-Complex showing the distribution of Cretaceous metamorphism (redrawn after Fügenschuh et al., 2000), small dashed rectangle: study area, large dashed rectangle: extent of geological map of Hammer (1929). BFZ = Brenner fault zone, BM = Brenner Mesozoic, EF = Engadine window, EL = Engadine line, IL = Inntal line, IQP = Innsbruck Quartzphyllite, NKA = Northern Calcareous Alps, SD = Steinach nappe, SL = Schlinig line, SFZ = Schneeberg fault zone, VSZ = Vinschgau shear zone. Variscan aluminosilicate zonation after Purtscheller (1969): And = Andalusite zone, Ky = Kyanite zone, Sill = Sillimanite zone.

For the eclogites in the central Oetztal, PT conditions of 27 kbar and 730 °C, together with an age of 370-340 Ma, were determined by geothermobarometry and Sm-Nd mineral isochrons (Miller and Thöni, 1995). The oldest metamorphic event within the OSC is represented by Ordovician/Silurian migmatites of the Winnebach area, for which Klötzli-Chowanetz et al. (1997) and Thöny et al. (2008) determined ages ranging between 490 ± 9 Ma (U/Pb zircon age) and 441 ± 18 Ma (U-Th-Pb electron microprobe monazite age).

The metamorphic evolution of the OSC is well-documented, whereas structural investigations are rather scarce (e.g. van Gool et al., 1987; Sölva et al., 2005). Although a clear attribution of structural elements to the different tectonometamorphic events has not been proposed yet, it is our understanding that differences in metamorphic grades were essential to classify observed structures as Alpine or Variscan deformational events.

4. Deformation stages

The OSC is dominated by large-scale E-W-striking open folds. However, the exact location of their major axial traces were not clearly mapped at this stage, partly because of numerous non-cylindrical parasitic folds and a discontinuous appearance of lithologies due to boudinage. Despite these difficulties structural observations and regional map patterns can be interpreted in the following proposed deformation sequence.

4.1 Deformation stage D_1

D_1 is defined by a penetrative schistosity (S_1) formed by biotite and muscovite in amphibolite facies conditions and by the occurrence of garnet, staurolite and kyanite. D_1 is related to large amplitude (up to 100m scale) isoclinal folding and the main foliation can be addressed as a composite $S_{0/1}$ planar feature because S_1 is subparallel to the observed bedding. Main observations related to D_1 are (1) isoclinal and often intrafolial folds, (2) slightly S- or NE-plunging fold hinges (F_1), (3) mainly NE- and S-trending mineral stretching lineations (L_1) (sub-) parallel to the fold hinges and (4) slightly ESE- and WNW-dipping $S_{0/1}$-axial planes (Fig. 2, left).

Due to the parallelism of F_1 and L_1, D_1 folds could be interpreted either as sheath folds (Alsop and Carreras, 2007) or as isoclinal folds originally formed parallel to the stretching direction during intense shearing (Grujić and Mancktelow, 1995). The latter interpretation is favoured because the NE-SW trending stretching lineations point to shearing along this direction, but no unequivocal shear sense could be determined with respect to the D_1 stretching lineation. Further investigations are necessary to verify the geodynamic meaning of D_1.

4.2 Deformation stage D2

This deformation phase forms the dominant structures within the OSC and refolds the $S_{0/1}$ foliation (Fig. 3a, b). It is characterized by (1) sub-horizontal ESE- and WNW-plunging non-cylindrical parasitic folds (F_2) of different orders associated with open folds with kilometer-scale amplitudes, (2) sub-horizontal ESE- and WNW-orientated mineral stretching lineations L_2 and (3) steeply N- and S-dipping S_2 foliations with a similar mineral assemblage as S_1 (Fig. 2, left).

This superimposed folding leads to a strong parallelism of F_1 and F_2 that results in subparallel $S_{0/1}$- and S_2-foliations. These foliations can only be clearly identified in D_2 hinge zones where F_1 and F_2 remain more or less in their original orientation. D_1 and D_2 are responsible for the penetrative foliation generally observed throughout the whole OSC. Due to their great overlap in terms of metamorphic grade and geometry of structural elements, D_1 and D_2 are proposed to represent ongoing deformation during one protracted Variscan

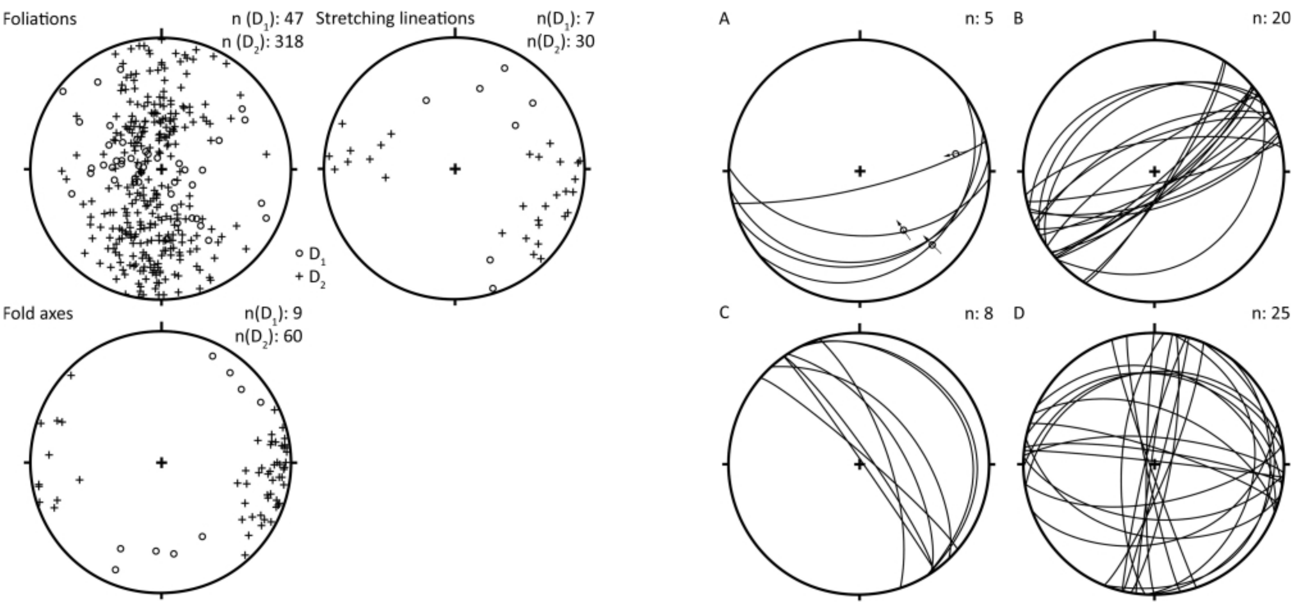

Figure 2: Structural data of the investigation area: (left) comparison of D_1 and D_2 foliations, stretching lineations and fold hinges; (right) A: NNW- and W-orientated D_5 thrusts, B: brittle faults related to the Stubaital fault, C: brittle faults related to the Halsl fault, D: brittle faults in the hanging wall of the Brenner fault.

tectonometamorphic event. D_1 and D_2 are clearly pre-Alpine since their axial traces and foliations are not observed within the BM and are discordant to the stratigraphic base of the BM (Krois et al., 1990).

4.3 Deformation stage D_3

This deformation stage is manifested by rarely observed W to NW-directed thrust faults and is interpreted to be related to Cretaceous thrusting (i.e., Trupchun phase, Froitzheim et al., 1994). D_3-structures are only hardly observable due to later overprinting and, according to Purtscheller (1978) and Fügenschuh et al. (2000), are restricted to the top of the OSC.

4.4 Deformation stage D_4

Subhorizontal to slightly SE-dipping foliations are developed along shear zones with top-to-the-SE directed shearing. Related folds have meter-scale amplitudes and show similar orientations as kilometer-scale D_2 folds in the OSC. However, due to differences in scale and qualitative determination of forming temperatures from thin section inspections (e.g. green biotite in BM: greenschist facies; brown biotite in OSC: amphibolite facies) these structures are instead attributed to a late Cretaceous extension event (Ducan-Ela phase, Froitzheim et al., 1994), which affected the uppermost part of the OSC close to the BM contact, whereas most of the deformation was taken up by the Mesozoic cover sequences. Moreover, planar SE-orientated normal faults led to the SE-dipping foliation subparallel to the bedding in the BM (Rockenschaub et al., 2003) and equivalent SE-orientated crenulation lineations or fault propagation folds in the OSC (Fig. 3c). The contact between the OSC and its cover (BM) is termed parautochthonous (Köhler, 1978; Krois, 1989) because only minor amounts of shearing occurred at this boundary.

4.5 Deformation stage D_5

Small-scale NNW- to SSE-vergent brittle thrusts were observable throughout the field area (Fig. 2A) but due to their limited extent do not show up on map scale. Mica-rich units of the OSC show a weak crenulation (Fig. 3d) with ENE-trending fold hinges, in accordance with observations made by Langheinrich (1965). Both their orientations (folds and thrusts) and kinematics (thrusts) are consistent with Cenozoic thrusting of Austroalpine units upon Penninic units (Rockenschaub et al., 2003), i.e. the Blaisun phase of Froitzheim et al. (1994).

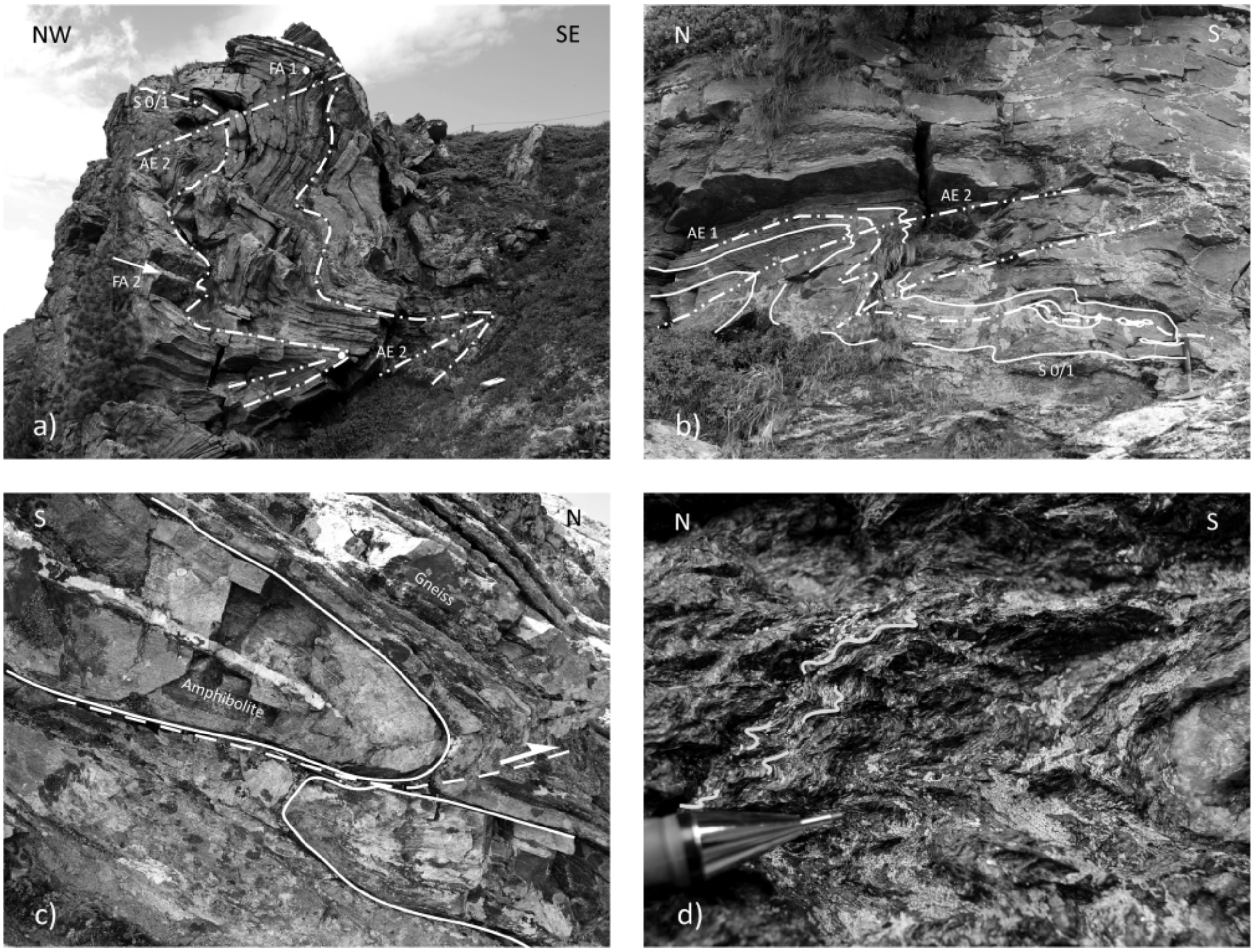

Figure 3: Outcrop scale structures: (a) D_2 hinge zone with refolded $S_{0/1}$, (b) south-vergent D_2 parasitic fold refolding isoclinal D_1 fold, (c) thrusting of originally (D_1) boudinaged amphibolite in a D_2 fault propagation fold, (d) D_5 crenulation in mica schist.

Independent evidence for this interpretation is pending.

4.6 Deformation stage D_6

Both from overprinting criteria and geometrical considerations, the youngest phase of deformation is related to deformation of the OSC and BM in the hangingwall of the Brenner fault zone (Fügenschuh et al., 1997). This fully brittle stage is evidenced by NE- (e.g. Stubaital fault, Fig. 2B) and NW-trending (e.g. Halsl fault, Fig. 2C) strike-slip faults together with N-S-trending normal faults (e.g. Seejoechl fault, Fig. 2D). Brittle deformation started essentially during the Neogene and has remained active until recent times (Reiter et al., 2005). The relative timing of activity along these faults, as proposed by Rockenschaub et al. (2003), cannot be verified because of a lack of exposure of cross-cutting relationships.

5. Model

The OSC is a polyphase folded complex, a fact already known and described by several authors (e.g., Schmidegg, 1956; Tollmann, 1963; 1977; Purtscheller, 1971; 1978; van Gool et al., 1987). A qualitative description of D_2 folds (in our notation) dates already back to Hammer (1929), but was not elaborated in a more genetic context. Only van Gool et al. (1987) investigated structures in the SW of the OSC, which show some similarities to the observed structures in the NE. The main problem with correlating the findings of van Gool et al. (1987) from the southernmost OSC with the structures in the north is the substantially stronger Cretaceous overprint in the south, where Eoalpine metamorphism reached amphibolite facies conditions. On the contrary, structures of the southernmost OSC are more likely related to Cretaceous tectonics than by Variscan deformation (Pomella et al., 2010).

Firstly, the observed two-phase folding is displayed by the two different sets of fold hinges and stretching lineations and the mutual fold interference in outcrops. Outcrop-scale fold interference patterns are largely of Type 3 (Ramsay, 1967), whereas the incorporation of the geological map of Hammer (1929) provides a more regional perspective showing a mixed Type 2/Type 3 fold interference pattern. The difference between Type 2 and Type 3 fold interferences is the orientation of initial folding and can be seen in Fig. 4. Type 2 superimposed folding leads to folding of both initial fold hinges and axial planes in contrast to Type 3 folding, which results in subparallel fold hinges without folding of the pre-existing fold hinges. Pure Type 2 fold interferences show mushroom-shaped outcrop patterns (Fig. 5, 90° between fold hinges and axial planes) and with decreasing angle between the two fold hinges the mushroom pattern becomes elongated (Fig. 5, 45° and 20°). The evidences described above indicate that the angle is likely 20-30° between D_1 and D_2 fold hinges, whereas the initial orientation of D_1 and D_2 axial planes is estimated as perpendicular, but according to Odonne and Vialon (1987)

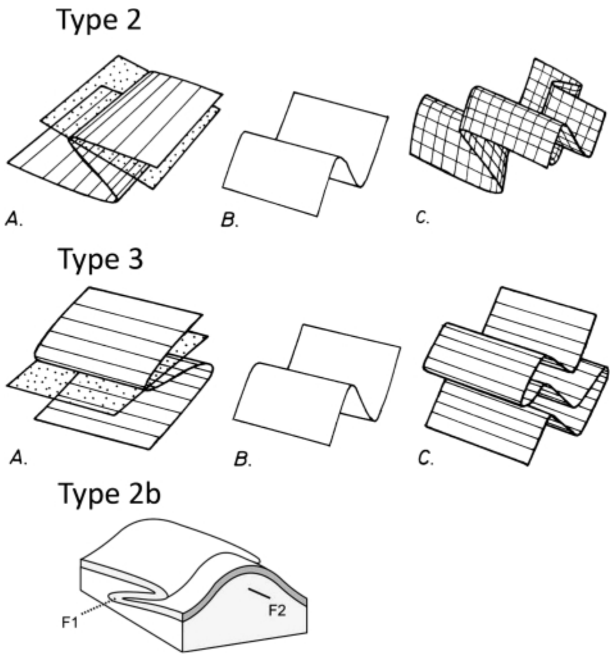

Figure 4: Type 2 and Type 3 fold geometries (Ramsay and Huber, 1987): (A) initial folding, (B) superposed folding, (C) resulting fold geometry and Type 2b fold geometry (Simón, 2004): (F1) initial fold axis of isoclinal folds, (F2) overprinting fold axis of open folds.

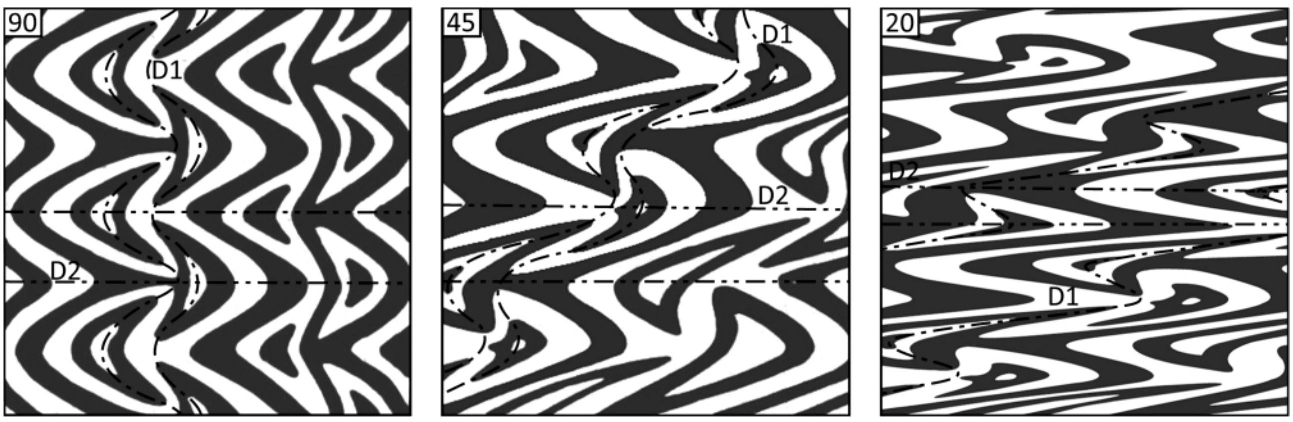

Figure 5: Fold interference pattern after Ramsay and Huber (1987): (90) typical Type 2 interference pattern with perpendicular fold hinges and axial planes, (45) intermediate Type 2/Type 3 interference pattern with an angle of 45°, (20) intermediate Type 2/Type 3 interference pattern with an angle of 20° illustrating differently stretched mushroom-shaped pattern.

Figure 6: Digitized geological map sheet “Oetzthal” of Hammer (1929) (1:175,000). For legend see Fig. 8.

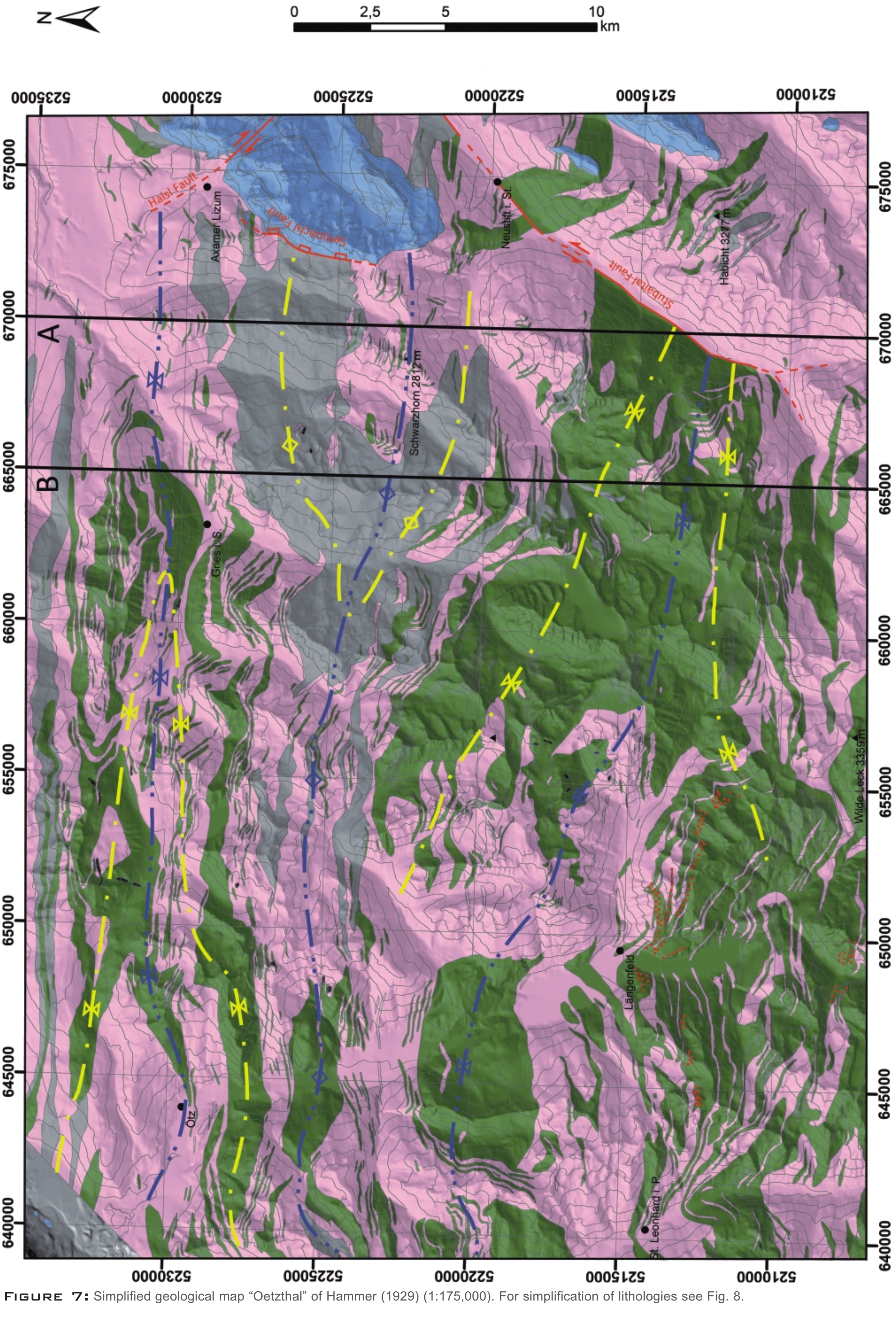

Figure 7: Simplified geological map "Oetzthal" of Hammer (1929) (1:175,000). For simplification of lithologies see Fig. 8.

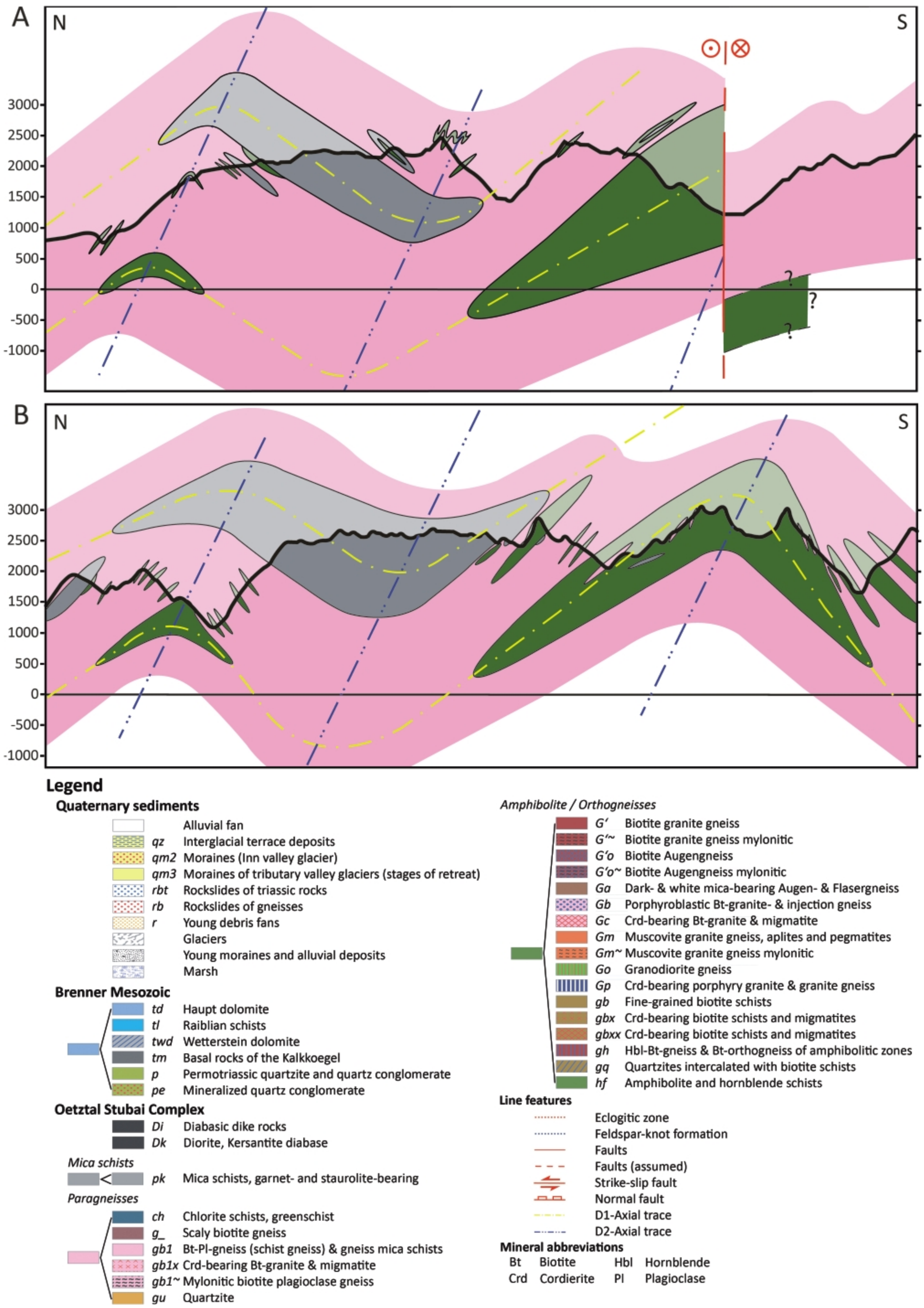

Figure 8: top: Cross sections A and B, axial traces indicated in Fig. 7; bottom: Legend to Fig. 6 to Fig. 8, (right squares indicating original lithological classification of Hammer, 1929 and left squares summarized in parentheses indicating reduced lithologies of Fig. 7 and Fig. 8, top).

directions of fold axes in areas of superposed folding cannot be used to determine compression directions.

Ghosh (1974) and Skjernaa (1975) distinguished, on the basis of physical experiments, that overprinting folds are cylindrical if the first folding event is tight or isoclinal, but low-cylindrical when first folds are rather open. Ghosh et al. (1992) also determined that fourth mode superposed buckling (Type 2b; Simón, 2004; Fig. 4) in contrast to third mode (Type 2a; Simón, 2004) consists of isoclinal D_1 folds with folded hinges and axial planes because of refolding by D_2. Although the cylindricity of D_1 folds, local thickness variations and viscosity contrasts of lithologies were neglected, these theoretical assumptions are consistent with the pattern observed in the investigated part of the OSC.

Ghosh (1974) and Grujić (1993) investigated Type 1/Type 2 fold interferences and determined that "strong shear strains develop along the hinge zones of those early folds that are located at the limbs of the later folds. This will rotate any lineation, which was not parallel to the first fold hinge line into parallelism with it." Transferred to Type 2/Type 3 fold interferences this parallelization of the D_1 hinges and D_2 hinges results in the observed outcrop-scale Type 3 fold interference patterns in D_2 limb zones and intermediate Type 2/Type 3 fold interference pattern could only be observed in D_2 hinge zones.

Type 2 interference patterns become much more evident after redrawing and simplifying Hammer's map (Fig. 6) and combining of the originally depicted lithologies to essentially three lithological groups (Fig. 7): (1) Paragneisses, (2) micaschists and (3) magmatics (felsic and mafic). Although this simplification might ignore local lithological details and variations, we insist on the regional coherence. The combination of felsic and mafic magmatic rocks was mainly done due to their rather discontinuous pattern, a feature resulting from their relatively higher competency during deformation and, consequently, their appearance as boudins. Moreover it has to be stated that the distinction of D_1 antiforms from synforms could not be carried out unequivocally. We based our interpretation on a pseudo-stratigraphy only for working purposes, which is made up of (from bottom to top) (1) magmatic rocks, (2) paragneisses and (3) micaschists. Based on this stratigraphy micaschists are found in the cores of synforms while amphibolites and orthogneisses form the cores of antiforms (Fig. 8, top). The effect of lateral thickness variations can be observed due to the initial isoclinal D1 folding and can be assumed for the entire northeastern part of the OSC (Fig. 8, top).

6. Discussion

The geodynamic setting of the investigated area is difficult to determine due to the lack of detailed investigations on the relation between metamorphism, deformation and related geochronological data. For the eclogites and migmatites of the central OSC, excellent petrological and geochronological data are available, but no correlation with different stages of deformation has been worked out at this point.

While meta-igneous rocks are associated with paragneisses and could then be syn- or postdepositional to them, qualitatively micaschists are almost exclusively free of magmatic rocks and could represent a late to post-magmatic formation.

The OSC was substantially affected by the tectonometamorphic overprint related to the Variscan orogeny (Maggetti and Flisch, 1993; von Raumer and Neubauer, 1993), which led to a still preserved metamorphic zonation, especially in its northern part (cf. Fig. 1), as well as a dominant E-W striking composite penetrative foliation. The D_1 structures are either related to this Variscan cycle, or could date back to an even earlier Ordovician-Silurian tectonism, coeval to migmatites of the central OSC (Schweigl, 1995; Thöni, 1999). Here, we assume a Variscan age for both our folding stages, since structures related to D_1 fit well with the known Variscan metamorphic temperatures of 469 - 630 °C and 4.2 – 7.3 kbar for the northern part of the OSC (Tropper and Recheis, 2003). Furthermore, the orthogneisses in the OSC, which are dated as Cambrian to Silurian, are affected by D_1 and D_2 (Thöni, 1999). The first phase (D_1) formed isoclinal folds with sub-horizontal fold axial planes. The second phase (D_2) developed kilometer-scale relatively open folds with subvertical, recently inclined fold axial planes. The resulting mixed Type 2/Type 3 interference patterns can stem from two independent and geodynamically distinct events, or can alternatively be formed during a single event, as shown by Forbes et al. (2004). The mixed Type 2/Type 3 interference pattern changes to a pure Type 3 pattern towards the west, whereas the eastern continuation of the OSC is unknown due to the tectonic boundary of the Brenner fault. Nevertheless future work should focus on the comparison of the OSC to the Bundschuh complex east of the Tauern window, a unit which is not unequivocally correlated with the OSC (Schmid et al., 2004; Neubauer et al., 2007).

7. Conclusion

Detailed structural mapping allows the determination of a relative chronology of deformation phases and the recognition of fold interferences in the OSC. Yet on a more regional scale, these small scale observations might not hold true because of lithological controls on folding (rock type, thickness variations, etc.), mainly non-cylindrical folds and variable amplitudes depending on their location in the polyphase deformed geometry. Our combination of local mapping and incorporation in a regional geological map, revealed a mixed Type 2/Type 3 interference pattern for the pre-Alpine two-stage folding. The unequivocal pre-Mesozoic character of the first two phases of deformation becomes clear when comparing their structural elements to those of the unconformably overlying Permomesozoic cover in this area. Four known younger deformation phases (ductile and brittle) affected both the basement (OSC) and its cover (BM) and range from Cretaceous to sub-recent.

Reducing the number of originally mapped lithologies by summarizing them to reasonable entities clarifies the structural framework of the OSC and reveals the main outcrop pattern of the region. Although we are aware of possible over-simplifications, we are confident that the main lithological elements, structural evolution and, eventually, the structural geometry can be inferred by this approach.

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