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(Article begins on next page)

### **Tectonophysics**

# Calcite fabric development in calc-mylonite during progressive shallowing of a shear zone: an example from the Annapurna Detachment zone (central Himalaya, Western Nepal) --Manuscript Draft--

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Corresponding Author:	Chiara Montomoli University of Turin ITALY				
First Author:	Laura Nania, PhD				
Order of Authors:	Laura Nania, PhD				
	Chiara Montomoli, PhD				
	Salvatore Iaccarino, PhD				
	Rodolfo Carosi, PhD				
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Suggested Reviewers:	Eugenio Fazio, PhD Associate Professor, University of Catania efazio@unict.it He is an expert for CPO methodology and meso and micro structural geology				
	Richard D Law, PhD Full Professor, University of Virginia Tech rdlaw@vt.edu Expert in Himalayan Geology, meso and microstructural analysis, South Tibetan Detachment System, crystallographic preferred orientation				
	Jean-Luc Epard, PhD Associate Professor, University of Lausanne Institute of Earth Sciences Jean-Luc.Epard@unil.ch Expert in Himalayan Geology snd structural geology				
	Sean Long Associate Professor, Washington State University sean.p.long@wsu.edu Expert of himalyan geology, South Tibetan Detachment System, microstructures				

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- Calcite fabrics superimposition can record crustal exhumation paths in shear zones
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- Protracted ductile shearing results in a lack of a brittle South Tibetan detachment
- Lithologies variation over the Himalayan strike influences the overall architecture

# Calcite fabric development in calc-mylonite during progressive shallowing of a shear zone: an example from the Annapurna

### <sup>3</sup> Detachment zone (central Himalaya, Western Nepal)

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5 6	Laura Nania <sup>a,b</sup> , Chiara Montomoli <sup>c,d, *</sup> Salvatore Iaccarino <sup>c</sup> , Rodolfo Carosi <sup>c</sup>						
7 8	a. Dipartimento di Scienze della Terra, Università di Firenze, Via Giorgio la Pira, 4, 50121, Firenze Italy						
9 10	b. Geological Survey of Canada, Natural Resources Canada, 601 Booth St, Ottawa, ON, K1A 0E8, Canada						
10 11 12	c. Dipartimento di Scienze della Terra, Università di Torino, via Valperga Caluso, 35, 10125 Torino, Italy						
12 13	d. Istituto di Geoscienze e Georisorse, CNR, Via Giuseppe Moruzzi, 1, 56127, Pisa, Italy						
14 15 16	L.N., <u>laura.nania@nrcan-rncan.gc.ca</u> ; C.M., <u>chiara.montomoli@unito.it</u> ; S.I., <u>salvatore.iaccarino@unito.it</u> ; R.C., <u>rodolfo.carosi@unito.it</u>						
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18	Corresponding author: Chiara Montomoli, chiara.montomoli@unito.it						
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#### 47 **1. Introduction**

48 Regional scale, orogenic wide, shear zones are regions of the Earth's crust that strongly affect orogens architecture through time. In the lower to the mid-upper continental crust, deformation often results in 49 mylonitic zones of variable width, depending on intrinsic and extrinsic (regional) deformation 50 51 parameters (e.g., see Ebert et al., 2008, 2009; Hunter et al., 2019; Cawood and Platt, 2021; Daczko and 52 Piazolo, 2022 with references). When marbles and carbonate-rich rocks are involved, meso-structural analysis can be a non-easy task (Nania et al., 2022b). The low contrast in competence between 53 54 rheological domains (e.g., primary or secondary foliations, see Passchier and Trouw, 2005) and the small 55 occurrences of micas in marbles prevent the formation of clear structures and kinematic indicators 56 typically used for a first characterization of the shear zones. To characterize the deformation of calc-57 mylonites, it is required to integrate mesostructural observations with microstructural analysis (e.g., Molli and Heilbronner, 1999; Molli et al., 2000; Leiss and Molli, 2003; Ebert et al., 2007; Oesterling et al., 58 2007; Herwegh et al., 2008; Molli et al., 2011; Rogowitz et al., 2014; Spanos et al., 2015; Sarkarinejad 59 and Heibati, 2017; Bauer et al., 2018; Negrini et al., 2018; Lacombe et al., 2021; Nania et al., 2022b). 60 61 However, given the difficulty of describing the mylonitic fabric at the mesoscale in marbles, there are

62 few works that engage in their study for tectonic investigations, and marble mylonites are still under-63 researched.

An example of this issue can be addressed in the case of the South Tibetan Detachment System (STDS) 64 65 in Himalaya. The STDS is a regional, syn-collisional, normal faults system, which has been largely the focus, over the last forty years, to define tectonic models for the exhumation of mid-crustal rocks in 66 67 collisional settings (Burg et al., 1983; Burchfiel and Royden, 1985; Hodges et al., 1992; Grujic et al., 1996; 68 Carosi et al., 1998; Beaumont et al., 2001, 2004; Vannay and Grasemann, 2001; Godin et al., 2006; Webb 69 et al., 2007; Kohn, 2008; Larson et al., 2015; Iaccarino et al., 2017). The STDS laterally involves different 70 regions of the Himalaya and, therefore, different lithotypes. Nonetheless, most of the tectonic (including 71 microtectonics) studies on STDS mostly concern areas where quartz-bearing rocks constitute the main 72 involved lithotypes, with minor attempts on calcite-rich tectonites in Central Himalaya (Parsons et al., 73 2016b, 2016d). Particularly in the Everest area, where spectacular outcrops of both quartz-rich and 74 carbonate-rich rocks are described (Carosi et al., 1998, 2002; Searle et al., 2003; Law et al., 2004, 2011; Waters et al., 2019), mesoscale observation, microstructure analysis, and geochronological 75 investigation have been adopted mainly on quartz-rich tectonites to define the picture of the 76 77 detachment system (Waters et al., 2019 with references), with fewer characterization of the carbonate 78 lithotypes in uppermost sections (Carosi et al., 1998; Corthouts et al., 2016; Larson et al., 2020). There, 79 the STDS is made up by (1) an older lower mylonitic zone, affecting mid-crustal rocks in the footwall 80 (essentially quartz-bearing lithologies) and marine metasediments in the hanging wall, and (2) an upper 81 younger discrete brittle fault, within the upward carbonate marine metasediments, with the same 82 kinematic and sense of shear of the lower detachment to which it laterally rejoints (Carosi et al., 1998; Searle, 1999; Searle et al., 2003; Law et al., 2004, 2011; Waters et al., 2019; Kellett et al., 2019; Larson 83 84 et al., 2020). The discrete upper younger brittle fault, however, is not documented in several areas along 85 the Himalaya (e.g., Cottle et al., 2007; Carosi et al., 2002, 2013; Kellett et al., 2019, with references), e.g., 86 when the km-thick mylonitic detachment involves carbonate-rich rocks (Nania et al., 2022b). Consequently, how the two elements of the STDS are related to a regional scale is still under debate 87 88 (Cottle et al., 2011; Carosi et al., 2013, 2018; Montomoli et al., 2017; Kellett et al., 2019, with references

therein; Nania et al., 2022b). It is therefore crucial to define how is the overall deformation of the
detachment system recorded in other areas of the belt, where lithotypes different from quartz-rich
rocks, such as marble mylonite, are involved.

92 This is the reason why we decided to characterize the STDS in a well-known region, the Kali Gandaki valley (Central Nepal, Fig. 1; Fuchs and Frank, 1970), where the calcite fabric of marble mylonites (the 93 94 main lithology) has been only little investigated until now (e.g., Parsons et al., 2016b, 2016d). The ductile 95 shear zone in the Kali Gandaki valley is known as Annapurna Detachment (Fig. 2a, b; Vannay and Hodges, 1996; Godin et al., 1999a; Godin, 2003; Waters, 2019; Pye et al., 2022, with references) hereafter 96 97 named as Annapurna Detachment zone. Marbles and limestones in the study area are spectacularly exposed (Colchen et al., 1986; Burchfiel et al., 1992, with references; Vannay and Hodges, 1996; Godin 98 99 et al., 1999a; Searle, 2010, with references; Parsons et al., 2016b, 2016c, 2016d) but only few kinematic 100 indicators occur (e.g., Carosi et al., 2014; Parsons et al., 2016b, 2016d). We, therefore, integrated optical, 101 crystallographic preferred orientation (CPO), and image analyses to identify new kinematic indicators 102 and to define the mylonitic fabric and the relative contribution and timing of the microstructures 103 development in the deformation regimes. We particularly focused on the type of the flow, the kinematic 104 vorticity conditions, and the strain rate. This allowed us not only to define the deformation evolution of 105 the local Annapurna Detachment zone in the Kali Gandaki area, but also to compare different 106 exhumation styles of the STDS for different and similar lithotypes along the belt and provides new 107 insights into a more general perspective concerning strain variation in marble mylonite during shear zones evolution. 108

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#### 110 **2.** Geological setting

111 2.1. Geological overview of the Himalaya

The Himalaya is an active collisional orogen linked to the collision and indentation of the Indian Plate into the Eurasia Plate started at c. 59-61 Ma (Hu et al., 2016; Parsons et al., 2020; An et al., 2021). From south to north, the main km-thick litho-tectonics units, accreted from the Indian northern margin, are (Fig. 1; see Hodges, 2000 for a review): the Siwalik Group (or Subhimalayan Sequence, consisting of Tertiary molasse sediments), the Lesser Himalayan Sequence (LHS) (subgreenschist facies to lowamphibolite facies rocks), the Greater Himalayan Sequence (GHS) (greenschist to high-temperature amphibolite facies rocks), and the Tethyan Himalayan Sequences (THS) (upper greenschist facies metamorphic rock to unmetamorphosed sedimentary rocks).



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- 121

#### Figure 1

122 The GHS and the THS have undergone two main tectonic stages: (1) the main collision in the Eocene, 123 and (2) a (syn-collisional) southward exhumation (namely N–S extension) of the north-dipping units in the Late Oligocene-Miocene, characterized by nearly isothermal decompression followed by 124 retrogression (Hodges, 2000, with references). During the southward exhumation, the regional-scale 125 126 STDS juxtaposed the high-grade rocks of the GHS in the footwall against the low-grade metasediments of the THS in the hanging-wall (Caby et al., 1983; Burg and Chen, 1984; Burchfiel et al., 1992; Carosi et 127 al., 1998; Law et al., 2011; laccarino et al., 2017; Montomoli et al., 2017; Kellett et al., 2019, with 128 129 references). Especially in the northern part of the Himalaya, corresponding to the northern THS (i.e., see 130 the geographical/topographic division of Yin, 2006), and to a lesser extent at the top of the GHS, the 131 collisional tectonics resulted in crustal thinning through orogen-parallel E-W extension (Jessup et al., 2019; Larson et al., 2019; Pye et al., 2022, with references). The tectonic shift from N-S to E-W extension 132

occurred from the late Miocene (Nagy et al., 2015; Parsons et al., 2016a; Larson et al., 2019), and was 133 probably a progressive tectonic style changes occurred overtime after the end of the STDS activity 134 (Murphy et al., 2002; Chen et al., 2022, with references; Pye et al., 2022). The E-W extension developed 135 through N-S trending normal faults/grabens (Coleman, 1996; Colchen, 1999; Blisniuk et al., 2001; Jessup 136 137 et al., 2008, 2019; Mitsuishi et al., 2012; Larson et al., 2019) and a possible generalized re-heating and thermal relaxation (Nania et al., 2022a) along with granitic intrusions in several areas (Roger et al., 138 139 1995; Visonà and Lombardo, 2002; Streule et al., 2010; Mitsuishi et al., 2012; Visonà et al., 2012; Carosi et al., 2013; Zhang et al., 2020; Chen et al., 2022). From the late Miocene-Pliocene, especially in the 140 141 southern part of the belt, the Himalayan exhumation rate abruptly increased together with the extreme climate-induced erosion rates (Huntington et al., 2006; Garzanti et al., 2007; Gemignani et al., 2018; 142 Govin et al., 2020). In the Thakkhola region of Western Nepal, this tectonic and climate-induced 143 144 landscape modelling resulted into the Kali Gandaki valley (Fig. 1; Fig. 2a), a deep N-S gorge carved 145 normal to the Himalayan trend (Colchen et al., 1986; Carosi et al., 2014).

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#### 147 2.2. The upper Kali Gandaki valley

The upper Kali Gandaki valley, shown in Fig. 2a, exposes the crystalline rocks of the GHS and the nearly 148 149 continuous sequence of the THS along a natural and almost N-S trending cross-section (Colchen et al., 150 1980; Vannay and Hodges, 1996; Godin et al., 1999a; Godin, 2003; Carosi et al., 2014; Parsons et al., 151 2016c). Over the last years, the GHS exhumation in the Kali Gandaki valley has been recognized as 152 composite and diachronous for (at least) two main portions, divided by high-temperature tectonometamorphic discontinuities (e.g., the Kalopani shear zone, Fig. 2a, Carosi et al., 2014, 2016; and the 153 High Himalayan Discontinuity (HHD), Iaccarino et al., 2015). Hereafter, we define the GHS as made by 154 two sub-units: a lower GHS (GHS<sub>L</sub>) and an upper GHS (GHS<sub>U</sub>). The GHS<sub>U</sub> is the only represented in Fig. 2 155 156 and comprises the Units (once referred as Formations by Godin, 2003) II and III distinguished by Godin 157 (2003) and Searle (2010). Unit II is made by calc-silicate bearing gneiss and marble (Fig. 2a, b).





The upper portion of GHS<sub>U</sub> (Formation III of Godin, 2003 and Unit III, IV and V of Searle, 2010) is made by augen gneiss at the bottom, deformed by the top-to-the-S Kalopani shear Zone (41-36 Ma: Carosi et al., 2016), followed by calc-silicate at the top involved in the Annapurna Detachment zone Fig. 2; Pye et al., 2022, with references). It consists of garnet-tourmaline-bearing augen gneiss, rare foliated two-mica

leucogranite, small lenses of kyanite-bearing metapelite, and kyanite-bearing migmatite (Brown and 164 165 Nazarchuk, 1993; Vannay and Hodges, 1996; Carosi et al., 2016; Parsons et al., 2016c, 2016d). The 166 portions of the GHS<sub>U</sub> mainly involved in the detachment are those at the top, consisting of metapelite, white marble, and a 200 m-thick sequence of calc-silicate marble (Fig. 2; see also Brown and Nazarchuk, 167 168 1993; Coleman, 1996; Hodges et al., 1996; Vannay and Hodges, 1996; Godin et al., 1999a; Searle, 2010). 169 Coarse-grained marbles have a mineral assemblage of Cal+Qz+Bt+Ms+Kfs (Table 1; mineral 170 abbreviations after Whitney and Evans, 2010), where minor Cpx+Grt+Ves, Cpx+Scp, and Hbl±Grt 171 associations point out upper amphibolite facies metamorphic condition (Vannay and Hodges, 1996; 172 Parsons et al., 2016b, 2016d).

173 The THS base comprises the highly deformed Sanctuary Fm. (Pêcher, 1978), consisting of Proterozoic/Cambrian black schist, metasandstone, and metalimestone, located in the core of large 174 175 anticlinal fold nappe (e.g., Fang Antiform) (Fig. 2a, Bordet et al., 1971; Colchen et al., 1986; Hodges et al., 176 1996; Vannay and Hodges, 1996). The Annapurna Fm. is composed by 1000-1300 m of Cambrian coarsegrained marble and impure metalimestone (Fig. 2), with calcareous meta-psammite and metapelite 177 178 interbedded with phyllite (Pêcher, 1978; Hodges et al., 1996; Vannay and Hodges, 1996; Godin et al., 1999a, 2001; Godin, 2003; Crouzet et al., 2007; Searle, 2010; Parsons et al., 2016c). A mineral 179 assemblage of Cal+Qz+Ms+Bt±Ep defines a greenschist-facies within the biotite-zone for these rocks 180 (Table 1; Carosi et al., 2014; Parsons et al., 2016b, 2016d). Upward, the Nilgiri Fm., consists of Ordovician 181 182 micritic metalimestone (Bordet et al., 1971) with a main Cal+Qz+Ms low grade mineral assemblage (Table 1). The Nilgiri Fm. grades to the north into pink dolomitic sandstone and arenite ("North Face 183 quartzites" in Godin, 1999a). The Palaeozoic sequence of the THS continues with lower greenschist/sub-184 greenschist facies (Crouzet et al., 2007) Silurian-Devonian Sombre Fm., comprehensive of black shale, 185 limestone and arenaceous sandstone, capped by the Permian-Carboniferous Tilicho Lake Fm. and Thini 186 187 Chu Fm. (Fig. 2a, Garzanti and Pagni Frette, 1991). A continuous Mesozoic to Cenozoic (Eocene) metasedimentary to unmetamorphosed marine succession defines the upper portion of the THS (Bordet 188 189 et al., 1971; Colchen et al., 1980, 1986; Gradstein et al., 1991; Garzanti et al., 1994).

190 Within both the GHS<sub>U</sub> and the THS a poly-phase tectonic history has been documented (Brown and 191 Nazarchuk, 1993; Hodges et al., 1996; Vannay and Hodges, 1996; Godin, 2003; Parsons et al., 2016b). 192 Due to the lithological and rheological differences between each formation, and to the different 193 structural/crustal level position, the deformation history and the P-T conditions of deformation varies 194 from the GHS<sub>U</sub> to the THS (Godin, 2003, with references). Specifically, five deformational events have 195 been ascribed to the THS (Godin, 2003, with references): the D1, corresponding to the collisional 196 tectonic phase (with SW-verging isoclinal folds); the D2 phase for NE-verging folds; the Annapurna 197 Detachment shearing events (defined as D3 and Dt for the ductile extensional transposition onset and 198 the related high-strain zone development, respectively); the D4 for the post-peak metamorphic, associated to southwest-verging kink folds; and the D5 event, linked to the orogen- parallel E-W 199 extension locally recorded by the N-S trending Thakkhola graben and related system of normal faults. 200 201 In this work, we recognized these deformation stages as three main events consistently with the main 202 tectonic phases of collision, N-S extension, and orogen-parallel E-W extension (described by Vannay 203 and Hodges, 1996), resulting in the most evident structures in the study area (see Fig. 2). Hereafter, the 204 D2 phase indicates the tectonic stage responsible for the northeast-verging folding in the THS and in the 205 syn (to immediately later) >1500 m-thick ductile high-strain zone of the Annapurna Detachment zone 206 (Fig. 2), grouping the D2, D3 and Dt phases of Godin et al. (1999a, 1999b, 2001). An example of a F2 207 structure is the Nilgiri anticline, refolding the older Fang Antiform (Colchen et al., 1986; Godin, 2003). 208 We suggest that the axial plane of the dominant F2 northeast-verging mega-folds tends to parallelism 209 with the mylonitic foliation of the Annapurna Detachment zone, being dragged by the detachment 210 movement (Fig. 2b) as originally proposed by Burchfiel et al. (1992). We include in the Annapurna Detachment zone both the ductile "Annapurna detachment" in sensu strictu of Godin et al. (1999a) and 211 212 the "Dhumpu detachment" recently described by Pye et al., (2022).

#### **3. Methods**

215 3.1. Calcite microstructure, paleopiezometry and finite strain analyses

216 Structural analysis was conducted on an almost N-S transect, oriented normal to the Annapurna 217 Detachment strike (Fig. 2a, b; Fig. 3). We selected eleven field-oriented marble samples from different 218 structural levels: ten specimens within the Annapurna Detachment zone and one immediately above it, in the Nilgiri Fm. for comparison (Fig. 2, Table 1). Standard thin sections were prepared on slides parallel 219 220 to the mineral lineation and perpendicular to the main foliation for microstructural analysis. Of these, six specimens from the GHS<sub>U</sub>, three from the Annapurna Fm., and one from the Nilgiri Fm. were selected 221 for image analysis to quantify the calcite volumetric abundance, aspect ratio, and grain size (see Table 222 223 1, Table 2). Large areas of interconnected calcite crystals were selected for each specimen to obtain a 224 representative sample of the microstructure. For each area, multiple images with the same pixel size 225 resolution of 5.08 µm (at least four different images for each area, rotated of 45°) were acquired under cross-polarized light with and without the gypsum plate inserted, to unambiguously identify calcite 226 227 grain boundaries. Grain boundaries were manually outlined as closed polygon using a vector-graphics 228 application (Adobe Illustrator, v. 27.5), discarding those grains cut by the image margins. Combining the representative areas, over 1000 interconnected crystals per specimen were selected to ensure 229 230 statistical reproducibility. Resulting maps were processed with ImageJ software (version 1.53t, by 231 Wayne Rasband, <u>https://imagej.nih.gov/ij/download.html</u>) to get the calcite volumetric abundance, 232 aspect ratio, and equivalent grain diameters. As second-phase minerals can affect the grain size (Ebert 233 et al., 2008) e.g., placing themself as obstacles to grain boundary mobility (enhancing pressure solution mechanisms and/or grain boundary sliding) and/or growing competitively (Busch et al., 1995; Herwegh 234 235 and Berger, 2004), we selected six specimens having the lowest percentage of second-phase minerals 236 (35-20%, Table 2), showing no evidence of static recrystallization or later further microstructures 237 affecting the grain size (see par. 4.1, 4.2). For each sample, the equivalent calcite grain size (expressed as the diameter of the equivalent circle, d) was calculated as the square root mean using GrainSizeTools 238 239 v2.0.2 (Lopez-Sanchez and Funez, 2015). Paleopiezometric estimates were used to get the differential 240 stress,  $\sigma$ , and the strain rate recorded by calcite deformed in the grain size insensitive (GSI) regime

(Renner et al., 2002, with references). The Barnhoorn et al. (2004) piezometer, designed for non-coaxial regime and calibrated for grain size data estimated through the linear intercept method, was adopted as  $Log\sigma = (-0.82 \pm 0.15) log(d) + 2.73 \pm 0.11$ .

As samples are affected by dynamic recrystallization, strain rate was calculated through the Renner et al. (2002) flow law, as  $\dot{\varepsilon} = A\sigma^n \exp \frac{\sigma}{\sigma_0} \exp \left(-\frac{Q}{RT}\right)$ , where A is the material constant (accounting for the chemical fugacity), n is the stress sensitivity (n=2),  $\sigma$  is the differential stress,  $\sigma_0$  is the resistance to glide, Q is the apparent activation energy for the process, R the universal gas constant, and T is the deformation temperature.

249 Twin density (D) was also measured in appropriately oriented grains using ImageJ software on 250 representative microphotographs (pixel resolution of 0.1 µm). Twins were analysed to estimate the 251 differential stress using the Rybacki et al. (2011) piezometer. Since, to date, it is not clear how the critical resolved shear stress of twins varies with the grain size (e.g., Parlangeau et al., 2018), we selected a pool 252 253 of crystals with homogeneous grain size. This choice does not affect the representativeness of the 254 sample, as it can be inferred later from the study of the grain size (unimodal grain size, par. 4.2). 255 Differential stress recorded by twinning was estimated through the equation  $\sigma_{(twin)} =$  $10^{1.29\pm0.02} D^{0.5\pm0.05}$  (Rybacki et al., 2011). Strain rates from twinning mechanisms were calculated 256 applying the exponential law by Rutter (1974) (see also Rowe and Rutter, 1990) as 257 Log  $\dot{\varepsilon}_{(twin)} = 5.8 - \left(\frac{250000}{2303 \text{BT}}\right) + 0.038 \,\sigma_{(twin)}.$ 258

259 Finite-strain estimation with the centre-to-centre method (Fry, 1979) was performed on microphotographs from six rock specimens (Table 1) using calcite crystals as strain marker. About 100 260 to 225 crystals for each specimen were measured. Fabric ellipses, Rxz, were achieved applying the 261 EllipseFit 3.6.2 W. 262 software (by Frederick Vollmer, available at https://www.frederickvollmer.com/ellipsefit/). For an objective shaping of the ellipses, we used the 263 264 exponential edge detection method (EED) of Waldron and Wallace (2007).

266 3.2. Analysis of Crystallographic Preferred Orientation (CPO)

267 Calcite and quartz crystallographic preferred orientations (CPOs) were investigated at the Geoscience 268 Centre of the University of Göttingen on five and three specimens, respectively. Data were acquired with an X-ray Texture Goniometer (model X'Pert Pro MRD\_DY2139 developed by PANalytical), specifically 269 270 designed for rock samples with relative coarse grain sizes (Leiss and Ullemeyer, 2006) (Table 1). 271 Measurements were made on rock's slides of c. 5 cm x 3.5 cm x 1-2 cm on the XZ planes (section parallel 272 to the mineral lineation, normal to the foliation plane) with five to six spots for each slide, each spot of c. 7 mm of diameter. Additionally, measurements were performed on the YZ plane of the finite strain 273 274 ellipsoid (section normal to the mineral lineation and macroscopic foliation) in those specimens where quartz CPO data were produced. Measuring CPOs on both planes allowed for a better pole figure 275 coverage and, therefore, for a better estimation of the orientation distribution functions of the less 276 abundant mineral phases (ODFs). To ensure the mineral phase composition at each slide, a 20 standard 277 278 diffraction pattern of 5-75° has been measured.

279 Complete pole figures were recalculated from ODFs exploiting the MTEX Toolbox (MTEX 5.4.0, 280 https://mtex-toolbox.github.io/) for Matlab of MathWorks (https://it.mathworks.com/products/matlab.html). For calcite cell parameters, we adopted a=b=4.988 281 Å, c=17.062 Å, and point group '312'; whereas for quartz we used a=b=4.913 Å, c=5.405 Å, and point 282 group '312'. To reduce artefacts linked to data acquisition of larger crystals (that provide more intense 283 284 peaks to the X-ray diffraction), the radially symmetric de la Vallee Poussin Kernel function (Schaeben, 1997) has been applied on ODFs calculation, with a halfwidth of 10° and a resolution of 5° (Hielscher 285 and Schaeben, 2008). 286

3D patterns of the main crystallographic elements were plotted on equal area lower hemisphere stereographic projection (pole figures). Calcite and quartz CPO intensities were calculated from ODF applying the texture index (or J-index) of Bunge (1982), and the M-index of Skemer et al. (2005) as comparison.

#### 292 3.3. Kinematic vorticity and shortening estimates

Estimating kinematic vorticity in naturally deformed rocks is often tricky as compositional and structural heterogeneities (typical in mylonite, e.g., SC-fabric, mica layering, etc.) are a prime cause of flow and strain partitioning (Handy, 1994; Jiang, 1994a, 1994b; Jiang and Williams, 1999; Kilian et al., 2011; Bhandari and Jiang, 2021). For this reason, we investigated the kinematic vorticity (sectional kinematic vorticity and mean kinematic vorticity, see Xypolias, 2010, for a review) and the resulting simple shear contribution only in the most homogeneous samples from representative areas of the detachment zone.

300 We applied two independents kinematic vorticity gauges on five suitable marble samples: the oblique 301 foliation method (Wallis, 1995), and the calcite CPO orientation (Wenk et al., 1987; see Table 1 and Table 302 3). The sectional kinematic vorticity (Wn) was estimated by measuring the  $\delta$  angle between the oblique foliation (Sb) and the mylonitic foliation (e.g., see Fig. 3). For each sample, the highest  $\delta$  angle (consistent 303 304 with the data distribution) was adopted (Wallis, 1995; Xypolias, 2010). From calcite CPOs, the simple 305 shear contribution was achieved by measuring the  $\omega$  angle between the main [c]-axes orientation and 306 the plane normal to the mylonitic foliation (Wenk et al., 1987). In this case, the mean kinematic vorticity 307 Wm (Passchier, 1997) can be derived from the obtained simple shear percentage assuming, for simplicity, a 2D flow (plane strain regime), in accordance with the flow regime recognized by Parsons 308 309 et al. (2016b) in the close areas, and by Law et al. (2004) further to the east, in the Everest Massif.

The oblique foliation method allows us to estimate the vorticity of a small increment of ductile deformation linked to the onset of the development of the shape preferred orientation (Xypolias, 2010). Alternatively, the calcite CPO can record the simple shear contributions occurring during a large segment of the deformation history (Wenk et al., 1987).

314

#### 315 **4. Results**

#### 316 4.1. Microstructures and their interpretation

The study samples from the Annapurna Detachment zone consist of calc-silicate-rich marble and white
marble belonging to the GHS<sub>U</sub>, and biotite-muscovite metapsammite and marble of the Annapurna Fm.

- from the THS (Fig. 2, Vannay and Hodges, 1996; Godin et al., 1999a). Interconnected calcite crystals are
- 320 the 65-75% of the bulk volume in the  $GHS_{U}$ , and over the 65-80% in the Annapurna Fm. (Table 2; Fig.
- 321 3a-j), defining an interconnected weak matrix (Handy, 1994).



Figure 3

324 Biotite, muscovite, and calcite shape preferred orientation (SPO) defines the main continuous foliation 325 (Sp). This fabric, correlated in literature to the mylonitic S2 foliation, develops from the GHS<sub>U</sub> to the 326 Annapurna Fm. (THS) (see also Hodges et al., 1996; Vannay and Hodges, 1996; Godin et al., 1999a; Carosi et al., 2014; Parsons et al., 2016b, 2016c, 2016d). Calcite grains have an aspect ratio of c. 2–2.5 (Table 2) 327 328 with a long axis ranging from parallel to high-angle with respect to the main foliation (Fig. 3c-f). In five 329 samples (K21-10-12, K21-10-13, K21-10-18, K22-10-19, and K29-10-52) calcite oblique foliations (Sb) 330 point toward the geographical north-direction. Calcite dynamic recrystallization is, therefore, syn-331 kinematic with the Annapurna Detachment top-to-the-north sense of shear (e.g., Fig. 3c-f).

At the bottom of the Annapurna Detachment zone, marbles belonging to the GHS<sub>U</sub> are coarse-grained (Fig. 3i-j). Straight grain boundaries, triple junctions, and a lack of undulous extinction in the lowermost sample (K31-10-65, Fig. 3j) are interpreted as indicative of static recrystallization or post-kinematic growth (e.g., Molli et al., 2000; Ohl et al., 2021), as also suggested by Brown and Nazarchuk (1993) and Vannay and Hodges (1996). Large white micas and biotite crystals, often overprinting the S2 foliation, show poikilitic structures (e.g., Fig. 3b, h, i), indicating a local post-mylonitic static recrystallization in the GHS<sub>U</sub>.

In the THS, within the Annapurna Fm. (e.g., K21-10-18; K22-10-19 at the base, K21-10-12, K21-10-13 339 above, Fig. 3b-f) calcite lobate grain boundaries and undulous extinction are typical microstructures for 340 intracrystalline deformation and dynamic recrystallization (Molli and Heilbronner, 1999; Molli et al., 341 342 2000; Barnhoorn et al., 2004). Calcite grain size distribution in these samples is unimodal (Fig. 4), whereas at the top (e.g., K21-10-12a) it becomes slightly bimodal, with rare calcite ribbons and 343 porphyroclasts. Unimodal grain size, lobate grain boundaries and little undulous extinction in calcite in 344 marbles at the base can be interpreted as resulting from the superimposition of grain boundary 345 migration (GBM) recrystallization regime on subgrain rotation recrystallization (SGR) regime (e.g., 346 347 Busch et al., 1995; Molli et al., 2000; Piazolo and Passchier, 2002; Stipp et al., 2002; Ulrich et al., 2002; Rogowitz et al., 2014, 2016). Type II and Type I twins overprint most calcite host grains (Fig. 3, Fig. 5a). 348 349 Thin twins often develop parallel to thicker tapered twins (Fig. 3g, 4b).

350 Only in few specimens, close to the GHS<sub>U</sub>, intercrystalline fractures, calcite veins and fluid inclusions 351 trails at a high angle to the main foliation point out partly healed post-mylonitic brittle deformation, without affecting the recognizability of the original grain size, developed by dynamic recrystallization. 352 Hot-cathodoluminescence images reveal evidence of minor overgrowths on calcite and thin 353 354 recrystallized (bright red) rims indicating Mn<sup>2+</sup> rich fluids recrystallization in calcite (Boggs and 355 Krinsley, 2006) e.g., in sample K21-10-18, where calcite shows straight grain boundaries (see Fig. 3e). 356 Micas (non-luminescent phases in Fig. 3e) are undeformed and often interstitial between the calcite rims in such specimens. Such features should be likely due to a fluid circulation linked to the Dhumpu 357 358 Detachment late reactivation (Pye et al., 2022).

Within small asymmetric lens-shaped quartz aggregates, crystals are incipiently deformed (e.g., samples
K21-10-12a, THS, and K21-10-18, at the boundary with GHS<sub>U</sub>). Out of the aggregates (e.g., Fig. 3c, j),
quartz occurs as rounded strong clasts within the weak calcite matrix.

362

#### 363 4.2. Paleostress and paleotemperature estimations

364 To avoid overestimation of the grain size due to the (local) annealing, six representative purer specimens with no evidence of static recrystallization have been selected. This choice does not affect the 365 366 representativeness of the grain size results, as static recrystallization does not seem to have particularly 367 affected the grain size of the annealed samples (e.g., Fig.2e, par. 4.1). Calcite mean grain size in samples 368 deformed by GBM varies from bottom to top (Fig. 4), with equivalent diameters from 770  $\mu$ m, in the 369 GHS<sub>U</sub>, to 390 µm in the Annapurna Fm. Recrystallized grains in sample at the top (K21-10-12a), 370 deformed by SGR, have a mean grain size of 250±30 µm (Table 2). Applying Barnhoorn et al. (2004) piezometer for these samples, differential stress values are in the range of 4-19 MPa (Table 2). The 371 Renner et al. (2002) relation for strain rates was applied once deformation temperatures were inferred 372 (Table 2). For the GHS<sub>U</sub> marbles (K29-10-51, K29-10-52, K31-10-64), a deformation temperature of c. 373 500°C (773.15 K) was adopted, as it is in the range proposed by Parson et al. (2016b) through quartz 374 375 and dolomite microstructures for the basal part of the ductile detachment in the Kali Gandaki and Modi 376 Khola valleys (300-600°C, Parson et al., 2016b, and up to 700°C probably during the previous stages,

see Parsons et al., 2016d) and, for similar structural levels, by Schneider and Marsh (1993) based on the
metamorphic assemblage and petrological insight (500-530°C for the GHS<sub>U</sub> involved by the Chame
Detachment, the prosecution of the STDS in the Marsyandi valley).



380 381

#### **Figure 4**

A deformation temperature average of c. 400°C (673.15 K) was adopted for THS rocks (samples K21-10-12, K21-10-12a, K22-10-19) following the main estimations for similar structural levels, close to the study area (300-500°C, Parsons et al., 2016b), and in the nearby Marsyandi valley (440-370°C, Schneider and Marsh, 1993). The corresponding strain rates for these deformation temperatures are of 8.1x10<sup>-10</sup>-1.3x10<sup>-9</sup> s<sup>-1</sup> in the GHS<sub>U</sub>, and of 1.1 x10<sup>-11</sup>-3.1x10<sup>-11</sup> s<sup>-1</sup> in the THS marbles (Table 2).

Type II *e*-twins have a thickness of c. 3-4 μm and a mean twin density of c. 40-55 (normalized to 1 mm length) (Fig. 5a, b; Table 2). According to Ferrill et al. (2004), the comparison of the mean twin width and the mean twin density is correlated to the deformation temperature. In our specimens, this ratio implies that the main twins' development occurred at temperatures below 300°C, likely of c. 200-250°C 391 (Fig. 5c). Crystals with thicker and spaced tapered twins, on which finer twins are superimposed, may





393

394

#### Figure 5

395 According to Rybacki et al. (2011) piezometer, twin density developed for differential stress of 118-154 MPa (Table 2). As e-twins typically dominate on the other deformation mechanisms at T<400°C 396 397 (Groshong, 1988; Burkhard, 1993), and the mean twin width vs the mean twin density ratio indicates deformation temperatures down to 200-250°C (Ferrill et al., 2004), we adopted an average temperature 398 of 250°C (523.15 K) for estimating strain rates linked to the shifting from dynamic recrystallization to 399 400 twinning as dominant deformation mechanism accounting that the twins induced by the non-coaxial deformation continued to be generated at the lower temperatures proposed by the graph. The resulting 401 402 strain rates are in the range of  $4.4 \times 10^{-15} - 2.2 \times 10^{-14} \text{ s}^{-1}$  (Table 2).

403

#### 404 4.3. Crystallographic preferred orientation (CPO) data and interpretation

405 Quartz, occurring in small asymmetric lenses or as isolated stronger clast within calcite matrix (two 406 samples belonging to the THS, one from the GHS<sub>U</sub>), have been analysed for the CPOs. Quartz has weak 407 CPOs intensity, defined by J-index of 1.16 and M-indexes of 0.01 (almost close to a total random 408 distribution, see Skemer et al., 2005), with multiples of uniform distribution in the range 0.7–2.2 409 (expressed as min-max in Fig. 6a-c). From bottom to top, the [*c*]-axes on [0001] pole figures define 410 clockwise asymmetric single girdle distributions suggesting dextral non-coaxial shear (Fig. 6a, Lister,

- 411 1977; Schmid and Casey, 1986). Couples of maxima are close to the Y-direction of the finite strain
- 412 ellipsoid (Fig. 6a-c).



Figure 6

We interpret that the wide girdle distribution resulted by a mix of rhomb<a> and prism<a>  $\pm$ 415 basal/ $\pi$ '<*a*> slip (e.g., Toy et al., 2008; Morales et al., 2011, 2014, with references), with further 416 417 mechanical grains rotation attenuating the CPO intensity (Stallard and Shelley, 1995). Alternatively, we propose that the broad peripheral [c]-axis distribution derive from a large contribution of dislocation-418 419 induced grain boundary sliding and subordinate dislocation glide (Kilian and Heilbronner, 2017), 420 determining the weak CPO strength (Graziani et al., 2020). In both cases, we interpret quartz CPOs as 421 due to intracrystalline deformation under a non-coaxial flow. Comparing the asymmetry of the fabric 422 (with respect to the orientation of the foliation and the lineation) with the original geographic 423 orientation, the dextral non-coaxial shear is consistent with the top-to-the-north ductile shearing of the 424 Annapurna Detachment zone (Fig. 6a-c).

Calcite CPOs strength, defined by the J-index of 1.09-1.39 and the corresponding M-index of 0.02-0.04, 425 426 is quite constant between samples, with multiples of uniform distribution in the range 0.5-3.5 (Fig. 6d-427 h). The clear calcite CPO patterns support that calcite grains have been strongly reoriented during 428 deformation despite of structural anisotropies and/or to the second-phase mineral 429 amount/distribution (Olgaard, 1990; Hippertt, 1994; Tullis and Wenk, 1994; Herwegh and Berger, 2004; Graziani et al., 2020). Broad [c]-axis point maxima are close to the Z-axis of the finite strain (almost 430 431 normal to the foliation), while  $\langle a \rangle$ -axes form girdles sub-parallel to the XY plane (Fig. 6d-h). In the uppermost sample, K21-10-12a, where the [c]-axis point maxima are strongly inclined, the poles to the 432 rhombic planes  $\{10\overline{1}4\}$  are weakly focussed on point maxima normal to the foliation plane, while poles 433 to the  $\{e\}$ -planes (on  $\{01\overline{1}8\}$  pole figure) define couples of strong asymmetric maxima inclined toward 434 435 the left (Fig. 6d). In all other analysed samples, the poles to the rhombic planes  $\{10\overline{1}4\}$  and  $\{01\overline{1}2\}$  define 436 weak small circles distributions (Fig. 6e-h). Point maxima of the  $\{e\}$ -planes poles are inclined toward 437 the left with respect to the foliation pole (Fig. 6e-h). For all specimens, a top-left asymmetry (Fig. 6d-h) 438 is antithetic to the calcite oblique foliations orientations measured on the same XZ-plane (Fig. 3c-f, h), 439 and to the quartz CPOs asymmetry (Fig. 6a-c).

The interpretation of calcite CPOs in terms of slip systems is not straightforward (Ohl et al., 2021, with
references). In all study samples, calcite CPOs can be due to the high-temperature basal<*a*> slip (during

grain boundary migration mechanisms) or to the coupled activity of rhomb< a > slip and e-twinning, or 442 to both (HT basal<*a*> slip followed by LT rhomb<*a*> slip and *e*-twinning). In the sample at the top (K21-443 444 10-12a), the peripheral asymmetric couples of maxima for the [c]-axis, the  $\{e\}$ -planes and the  $\{r\}$ -planes let us lean towards the coupled activity of rhomb<*a*> slip and *e*-twinning. Indeed, *e*-twinning can 445 446 strongly rotate calcite [*c*]-axes against the sense of shear (Wenk et al., 1987; Lacombe, 2010; Tripathy 447 and Saha, 2015), also favouring the slip along the  $\{r\}$ -planes (Oesterling et al., 2007). Concerning the 448 other samples, the observed CPO can be potentially related to the activity of any calcite slip system. 449 However, the strong maxima of the [c]-axes and the  $\{e\}$ -planes, and the absence of peripheral maxima 450 in the pole figures of the  $\{r\}$ - and  $\{f\}$ -planes, are consistent with a CPO balanced by *e*-twinning or 451 basal<a> slip alone. Different degrees of [c]-axis maxima inclination in most samples point out heterogeneous deformation (Kern and Wenk, 1983; Wenk et al., 1987), while the symmetry of the [c]-452 and {e}-maxima in sample K21-10-12 (Fig. 6e, about 2° of asymmetry to the left) indicates a CPO 453 454 equilibration under dominant pure shear conditions (Wenk et al., 1987). In general, calcite CPOs top-toleft asymmetry is consistent with an antithetic orientation linked to intracrystalline deformation 455 456 accommodated under a top-to-right non-coaxial flow. The top-to-right non-coaxial flow observable on 457 the XZ-plane of the finite strain is geographically consistent with the top-to-the-north sense of shear of 458 the Annapurna Detachment zone.

459

460 4.4. Kinematic vorticity and shortening of the flow: results and interpretations

461 The Wenk et al. (1987) method (Fig. 7a, Table 3) provided the simple shear contribution in four samples (K21-10-12, K21-10-18, K22-10-19, K29-10-52). We discarded specimen K21-10-12a as the [c]-axis 462 463 maxima do not fall on the primitive circle (Fig. 6d) and its CPO is, therefore, over inclined for the Wenk 464 et al. (1987) method. Vice versa, we kept in the sample the specimen K21-10-18 (despite we previously 465 interpreted as affected by little static recrystallization) since the superimposition of incipient annealing 466 should not have affected the orientation of the fabric but its intensity (Barnhoorn et al., 2005; Herwegh 467 et al., 2008), and it does not seem affected by the late brittle deformation associated to the Dhumpu Detachment defined by Pye et al. (2022). Among the sample, the  $\omega^{\circ}$  angle between the [*c*]-axis maxima 468

and the pole to the mylonitic foliation indicates a contribution from c. 30% to c. 50% of simple shear for
specimens K22-10-19, K29-10-52, and K21-10-18 (Fig. 7a, Table 3). A lower simple shear is deduced for
the specimen K21-10-12, where almost symmetric [*c*]-axis maxima are displayed (Fig. 6e). Assuming a
plane strain deformation, as inferred through quartz CPO studies on the regional structure of the STDS
from Eastern Nepal to Western Nepal (e.g., Law et al., 2004, 2011; Parsons et al., 2016b), the percentages
of simple shear of specimens K22-10-19, K29-10-52, and K21-10-18 correspond to a mean kinematic
vorticity (Wm) of 0.45-0.71 (Table 3, Fig. 7b).



477

476

Figure 7

The Wallis (1995) method for the sectional kinematic vorticity (Wn) was applied on four samples (K21-10-12, K21-10-13, K22-10-19, and K29-10-52, Table 3) where oblique foliations (Sb) are more evident and statistically developed (e.g., Fig. 3c-f). The  $\delta^{\circ}$  angles between the Sb and the main foliation decrease up-section from c. 26 to 20°, corresponding to Wn=0.79-0.64 (Table 3). For the same assumptions used for the previous method (i.e., plane strain deformation), the sectional kinematic vorticity range indicates a contribution of c. 40-50% of simple shear (Table 3, Fig. 7b).

With the exclusion of the uppermost samples (K21-10-12), we observe that both oblique foliation and
twinning-influencing calcite CPOs recorded comparable and quite consistent subsimple shear flow
conditions (Fig. 7b) even considering intrinsic limitations of vorticity gauges (e.g., the assumption of a
nearly plane strain dominated deformation, Iacopini et al., 2008; Xypolias, 2010; Fossen and Cavalcante,
2017).

489 In general shear, the coaxial component of the deformation is connected to the shortening perpendicular 490 to the flow plane (layer parallel extension) as a function of the finite strain (Wallis et al., 1993). The finite-strain equivalent ellipse ratios (R<sub>XZ</sub>) of six specimens (K21-10-12a, K21-10-12, K22-10-19, K29-491 492 10-52, K31-10-64, and K31-10-65; Table 1) range between 1.24 to 1.58, and slightly increase down-493 section within the Annapurna Detachment zone vertical profile (Fig. 8, Table 3). Shortening values of 494 0.89-0.92 and of 0.86-0.87 (Table 3) are then obtained combining the finite-strain ratio (R<sub>XZ</sub>) with the kinematic vorticity estimates after Wallis (1995) method and the Wenk et al. (1987) method, 495 respectively. 496

497



498

499

#### Figure 8

#### 500 **5. Discussion**

501 5.1. Deformation style of the Annapurna Detachment zone

A concept typically adopted for studying shear zones is "self-similarity", i.e., the attitude of rocks to produce consistent structures from the large scale to the microscale. In marble mylonites of the Annapurna Detachment zone, calcite grains define the weak interconnected matrix, accounting over 505 65% of the bulk volume in each specimen (*Table 2*). Calcite constitutes the main weak phase 506 accommodating the overall deformation (e.g., Handy, 1994) and, as this trait occurs in the whole 507 sampling, we can scale up the picture we have from the microanalysis to infer large-scale information 508 for the detachment zone. Moreover, marble mylonites involved by the Annapurna Detachment zone 509 show remarkable lithological affinities with the nearby STDS in the Dolpo region to the west (Fig. 1, see 510 Carosi et al., 2002, 2007) and in the Modi Khola and Marsyandi valleys (close to the Manaslu range in 511 Fig. 1) to the east (Schneider and Masch, 1993; Searle, 2010; Parsons et al., 2016b, 2016d; Carosi et al., 512 2023). This allows us to exploit the literature database for detachment as external constraints to add to 513 our microstructural investigation.

514 Calcite and quartz fabric and CPOs have clear patterns related to a plastic deformation, like those 515 identified in the area by Parsons et al. (2016b). Microstructural and calcite CPOs data allowed us to 516 recognize two dominant deformation mechanisms that we can use to picture the main deformation 517 parameters: (1) dynamic recrystallization by GBM/SGR, determining the grain size distributions and the 518 oblique foliations; (2) type II twinning of calcite crystals. From our analyses of the kinematic indicators 519 at the microscale (e.g., Fig. 3c-f, h), and by pole figures data interpretation (Fig. 6), both intracrystalline processes accommodated a ductile deformation under a top-to-the-north non-coaxial flow (Fig. 7b) 520 consistent with the Annapurna Detachment zone shearing. Kinematic vorticity estimates based on the 521 oblique foliation method (related to dynamic recrystallization) and the CPO (balanced by twinning) are 522 523 consistent with only one outlier (Fig. 7b) for a sample at the detachment external limit (Fig. 2). A main 40-50% of simple shear is constrained for Wn=0.64-0.79 through the oblique foliations (Wallis, 1995), 524 that is consistent with the simple shear range of 30-50% derived from calcite CPOs (Wenk et al., 1987; 525 Fig. 7b). The presented data are a semi-quantitative result because the reference frame adopted is the 526 527 mylonitic foliation instead of the exact shear plane. In natural shear zones, it is recurrently assumed that 528 the mylonitic foliation is in close parallelism with the shear plane. In our case, this assumption does not compromise the result as our samples belong to a high strain zone, where the pervasive mylonitic 529 530 foliation accommodated a huge amount of strain (Godin et al., 1999a). A detachment-parallel transport magnitude of 25-170 km has been estimated e.g., by Law et al. (2011) for the regional prosecution of the 531

532 detachment system in the Everest area, confirmed toward the east by Long et al. (2019) in NW Bhutan. 533 This extreme magnitude of tectonic transport allows to assume that the mylonitic foliation had 534 neglectable variations in angle with the shear plane from a regional perspective (Fossen, 2016). Moreover, the kinematic vorticity estimates that we propose for the Annapurna Detachment zone (0.64-535 536 0.79) fit with the one estimated for the STDS in the Everest area, where kinematic vorticity data of 0.67-537 0.98 are reported by Law et al. (2004and Larson et al. (2020) through the porphyroclasts-based (Wallis 538 et al., 1993; Simpson and De Paor, 1997) and the quartz [c]-axis based (Wallis, 1992) vorticity gauges. Similar kinematic vorticity values (Wm = 0.74–0.91), supportive for a simple shear dominated flow with 539 540 a critical component of pure shear, have been found also in the sheared limestone and marble of the THS 541 by Jessup et al. (2006) in the same area. The minor differences between our estimates (up-to Wm=0.79) 542 and previous proposed kinematic vorticity data possibly depend on the strain partitioning of the 543 complex heterogeneous STDS from area to area. This can be due to the different involved lithologies and 544 the different strain memory of the analysed structures.

The centrepiece of comparing the two kinematic vorticity gauges adopted methods (oblique foliation and CPOs) is that the same reference frame has been used. Therefore, regardless of how quantitative the result may be, our data indicate that both structures, deriving from the two intracrystalline deformation mechanisms, developed not only for the same kinematics but with comparable simple shear contribution during the non-coaxial flow.

550

5.1.1. Evolution of the deformation of the marble mylonite of the Annapurna Detachment zone 551 To unravel whether dynamic recrystallization and twinning reflect a single step of the long-lasting 552 553 shearing (active together) or a progressive change in the plastic regime during the exhumation/cooling path (an early and a late stage of shearing), hereafter we focus on the differential stress, the deformation 554 555 temperature, and the strain rate estimates. We will later compare these results with the literature database for the Nepalese areas to visualize the differences and similarities of the detachment system 556 when involving marbles instead of quartz-bearing lithologies. There are several works showing that 557 558 grain boundary mobility and twinning can develop together for the same deformation conditions, at T<400°C, and for low differential stress (e.g., Schmid et al., 1987; De Bresser and Spiers, 1993; 1997). Twinning produces high-angle boundaries, in which reticular defects and the stored elastic energy accumulate, triggering twin/grain boundary mobility mechanisms at MT-HT conditions, or pressure solution/solution transfer at LT conditions (Lafrance et al., 1994). The main tool to verify in which regime the deformation has been accommodated is the comparison of the differential stress and deformation temperature.

565 Our data indicates that the differential stress, the deformation temperature, and the strain rate 566 estimable for GBM/SGR and twinning are different (Table 2). We propose differential stress values of 4-567 19 MPa for the grain size development (adopting Barnhoorn et al., 2004, paleopiezometer) and of 118-154 MPa for twinning (after Rybacki et al., 2011 paleopiezometer; Table 2). For the dynamic 568 569 recrystallization mechanism, equivalent results have been reported for the Everest area (Law et al., 2011; Waters et al., 2019). Adopting the Stipp and Tullis (2003) quartz-based piezometer, Law et al. 570 571 (2011) and Waters et al. (2019) provided differential stress records of 10–15 MPa at the base of the STDS, and strain rates of 10<sup>-12</sup>-10<sup>-15</sup>s<sup>-1</sup> from the base of the detachment to almost 600 m of vertical 572 distance. Concerning the twin density paleopiezometer, differential stress values as those estimated in 573 574 this work for twinning in calcite (>100 MPa) are not excessively high when compared with data from 575 other Low-Angle Normal Faults (e.g., the Whipple detachment in South-eastern California, Axen, 2004, 576 2019) and are consistent with results obtained in the Lower Dolpo region for the same structure in the 577 STDS marbles (Nania et al., 2022b). Nevertheless, the inferred differential stress to produce twins in calcite is about one order of magnitude higher than that required for dynamic recrystallization. 578

The comparison between the deformation temperatures for the two intracrystalline mechanisms is more complicated and requires an interdisciplinary approach. During shearing at mid- to upper-crustal levels, slow grain boundary migration in calcite controls grain boundary morphology for a wide temperature range (Lafrance et al., 1994) and different nature of intercrystalline fluid (Schenk et al., 2005). Furthermore, equivalent CPOs can form at temperatures above 500°C down to T<150°C (e.g., Molli et al., 2010; Verberne et al., 2013; Bauer et al., 2018; Sly et al., 2020, with references; Lacombe et al., 2021). There are currently no calibrations for calcite slip system activation as geothermometer that 586 consider the role of fluids and the abundance and composition of second-phase minerals (e.g., Ohl et al., 587 2021). For this reason, calcite features must be compared with those of the metamorphic minerals that 588 are typically indicators of the deformation temperature. Calcite shape preferred orientation (SPO) is consistent with that of calc-silicates and main metamorphic minerals (Fig. 3). This indicates that calcite 589 590 recrystallization is syn-tectonic with the metamorphic assemblage indicative of the metamorphic facies. 591 Amphibolite facies assemblage of  $GHS_U$  prosecutes throughout the Annapurna Fm. base (THS), 592 decreasing abruptly to greenschist facies in few hundreds of meters (see also Garzanti and Pagni Frette, 1991; Garzanti et al., 1994; Hodges et al., 1996; Vannay and Hodges, 1996, Crouzet et al., 2007; Parsons 593 594 et al., 2016b). These two metamorphic facies let us consider a main temperature range of 500-600°C for the GHS<sub>U</sub>, and of at least 400°C for the base of the THS that is involved by the detachment zone. This 595 temperature range is consistent with the one proposed in close areas e.g., by Parsons et al. (2016b) for 596 597 two transects in the Kali Gandaki and Modi Khola valleys through quartz and dolomite microstructures, 598 especially when compared with the deformation temperature associated to the GHS<sub>U</sub>. It is also 599 consistent with the petrological constraints (i.e., calcite/dolomite geothermometer) of Schneider and 600 Masch (1993) and of Crouzet et al. (2007) in close areas, where the temperature proposed from the top of the GHS<sub>U</sub> to the biotite-zone of the THS are of 520-390°C. Similar estimates for the GHS<sub>U</sub> (c. 500°C) 601 602 have been proposed through the analysis of quartz microstructures combined with petrological 603 constrains by Nagy et al. (2015) and Soucy La Roche et al. (2018) for the upper Karnali valley of Western 604 Nepal (520°C for structural levels comparable with those of our samples) and in the Everest area in Eastern Nepal (Law et al., 2004, 2011; Cottle et al., 2011; and Waters et al., 2019). Such data rely on 605 different geothermometers, such as quartz CPO and opening-angle thermometry (Law et al., 2004, 606 607 2011), Raman spectroscopy on carbonaceous material (Cottle et al., 2011), and petrological constraints 608 (Waters et al., 2019). Especially with reference to Law et al. (2011), we highlight that the range of 609 deformation temperatures for the rocks at the base of the detachment (GHS<sub>U</sub>) has been precisely associated with the mechanisms of dynamic recrystallization and the oblique foliation, as in our case for 610 611 the syn-kinematic recrystallization of calcite. Therefore, we hypothesize that the part of the detachment

shearing accommodated by dynamic recrystallization and oblique foliation development occurred at
mid-temperatures conditions of 500/550-400°C (from the base to the top of the involved volume).

By contrast, our temperature estimates for twins are significantly lower (Fig. 5c). According to the semiquantitative thermometer, e-twins developed down to (at least) 250°C and, in any case, twinning starts to be a dominant mechanism able to re-orient the CPOs at deformation temperatures below 400°C (Groshong, 1988; Burkhard, 1993). Therefore, not only the differential stress values but also the deformation temperatures associated with the dynamic recrystallization and twinning are significantly different.

620 For the temperature and the differential stress ranges that we obtained, the strain rate required for 621 dynamic recrystallization of calcite (c. 10<sup>-11</sup> s<sup>-1</sup>) is greater than the results for twinning development (of. C. 10<sup>-15</sup>-10<sup>-14</sup> s<sup>-1</sup>) (Table 2). The methods used to calculate strain rates for GBM (Renner et al., 2002) are 622 probably not effective for acquiring quantitative estimates of impure marbles. Overall, unlike the case 623 of twinning, for little temperature variations in the calculation of the strain rates for GBM, the resulting 624 values change by several orders of magnitude. All of it make us suspicious that our estimates for the 625 626 strain rate of calcite recrystallization are underestimated. However, even without considering the results for the strain rate for GBM quantitatively, we suggest that for higher temperatures and lower 627 differential stresses the strain rate required for plastic deformation must be faster than that necessary 628 629 for the combination of lower temperature and higher differential stress, necessary for twinning. 630 Therefore, we propose that marbles recorded a (not yet quantifiable) slowing of the deformation during 631 the Annapurna Detachment zone ductile shearing, in addition to the lowering in deformation temperature and the increase of differential stress. 632

For a normal continental geothermal gradient of  $25-40^{\circ}$ C/km, in accordance with the typical lithospheric strength profiles, a differential stress of  $\leq 15$  MPa at temperatures of 400-500°C (recorded by the syn-kinematic dynamic recrystallization) occurs in the middle-upper crust under ductile deformation conditions. *Vice versa*, a high differential stress at low temperature conditions for viscouslike deformation (recorded by twins) usually occurs in the brittle-ductile transition (Fig. 9). Combining all the microstructural data with, CPOs, metamorphic facies, and temperatures from the literature

database, we propose that the two deformation mechanisms reflect two stages of progressive ductileshearing of the Annapurna Detachment zone in the Kali Gandaki valley (Fig. 9):

- D2<sub>early</sub>, recorded by calcite (and quartz) intracrystalline deformation, and characterized by a
   down-section lowering of differential stress (range of 4-19 MPa) and increase of deformation
   temperature (from 400°C in the THS to at least 500°C in the GHS<sub>U</sub>) within the vertical Annapurna
   Detachment profile, under "fast" strain rates.
- 645 D2<sub>late</sub>, recorded by twinning in calcite, defined by high differential stress (118-154 MPa) and low
   646 deformation temperatures (down to 250°C), under lower strain rates (of c. 10<sup>-14</sup> s<sup>-1</sup>).



#### 648



We therefore suggest that both dynamic recrystallization and twinning are not simply linked to multiple 649 650 activation pulses of the ductile Annapurna Detachment zone. The ductile flow within the Annapurna 651 Detachment zone protracted up to shallow crustal levels, close to the epizone/anchizone. After the 652 cessation of the plastic shearing, at temperatures below 200°C, marbles experienced minor brittle 653 deformation, little documented by intercrystalline fractures, calcite veins, and fluid inclusions trails at a 654 high angle to the main foliation. To which event this brittle deformation is connected is not immediately 655 observable from our sampling. The onset of a brittle deformation toward the north and at the base of 656 detachment has been already identify in the Kali Gandaki valley (Hurtado et al., 2001; Godin, 2003, with 657 references). An example of this is the late brittle deformation that reactivated the Dhumpu Detachment,

during a distinctive episode of late deformation in the Pliocene (Hurtado et al., 2001; McDermott et al.,
2015; Pye et al., 2022).

660 With regard to the STDS *in sensu strictu*, we link the D2<sub>early</sub> deformation temperatures, strain rates and 661 differential stress to the main detachment shearing and, more precisely, to the extensional juxtaposition 662 of the hot GHS<sub>U</sub> with the cold THS, and to the development of the pervasive mylonitic foliation and 663 oblique foliation (Fig. 9). The following stage, where twinning dominated in the CPO reorientation 664 (D2<sub>late</sub>), occurred with a strain rates of c. 10<sup>-15</sup> s<sup>-1</sup>, which is still compatible to the strain rates observed 665 though quartz paleopiezometry for the Everest area even in the main mylonitic zone (Law et al., 2011). 666

- 667

# 5.2. Tectonic implications for the Annapurna Detachment zone and the South TibetanDetachment System

A temporal variation of the STDS internal deformation and rheology has been described for marbles and 670 metalimestone in the Lower Dolpo Region (Nania et al., 2022b) and for quartz-mylonite in Eastern 671 672 Himalaya (Long et al., 2019; Zhang et al., 2022). In central to eastern Himalaya, it has been suggested that this temporal variation occurred with a later localization of a brittle fault at shallower levels, with 673 674 the same kinematics, merging into the ductile shear zone (Carosi et al., 1998; Searle et al., 2003; Searle, 675 2010, with references). The Qomolangma and Lhotse Detachment in the Everest area represent the best-676 preserved structures for this architecture (Carosi et al., 1998; Searle et al., 2003; Schultz et al., 2017). As 677 in our case, an increase of differential stress overtime within the STDS has been documented implicitly in similar lithologies by Law et al. (2011). Values of 25–35 MPa are reported for a younger upper brittle 678 679 segment of the STDS, whereas values of 10-15 MPa are documented for the older lower ductile shear 680 zone in the Everest area.

The younger ductile-brittle to brittle normal-sense fault, however, crops out only in few other areas, especially in eastern Himalaya (e.g., Sikkim and Zhergerand, see Kellett et al., 2013; Montomoli et al., 2017, with references), lacking in several parts of the belt (Cottle et al., 2007; Kellett and Grujic, 2012; Carosi et al., 2013; Kellett et al., 2019 with references). Despite the occurrence of later normal-sense faults truncating (or close to) the mylonitic zone, like below the Phu Detachment in the Marsyandi valley,
there is no conclusive evidence for the later localization of a brittle STDS segment where carbonatebearing rocks are dominant, as in the Kali Gandaki valley, the Lower Dolpo region (Carosi et al., 2002,
2013), and the Manaslu range in Mid-Western Nepal.

For the Lower Dolpo region, the lack of the brittle fault at the top of the marble mylonite in the detachment zone has been recently correlated to the ability of calcite to deform plastically even at shallow crustal levels (Nania et al., 2022b). At temperatures below c. 300°C, quarzitic rocks would deform in the brittle regime, and the increasing strain hardening within the plastic-to-brittle shear zone would explain the migration and the new localization of the detachment, which is not required when marble mylonite are involved.

695 The structural data for the marble mylonites in the Annapurna Detachment zone support the same idea of a progressive evolution of the STDS without the localization of the upper branch (prior of possible 696 697 late re-activations). The two main differences between the deformation path of the Annapurna Detachment zone and the one in Lower Dolpo concern the kinematic vorticity and the strain rate 698 699 occurring during the D2<sub>late</sub>. In the Lower Dolpo region, the strain rate remains constant from D2<sub>early</sub> to 700 D2<sub>late</sub>, while the kinematic vorticity recorded during the D2<sub>late</sub> is lower, locally suggesting a decelerating strain path (Nania et al., 2022b). We do not document, here, the same pattern in the Kali Gandaki valley, 701 where the kinematic vorticity remained constant even when the plastic deformation rates decreased. 702 703 Adopting our kinematic vorticity results, the shortening estimates of c. 13-14% (as a minimum estimate due to the used calcite crystals as strain markers) are still comparable with the values reported for the 704 705 Everest area by Law et al. (2004) for the lower detachment (10-30%) and by Larson et al. (2020) for the upper detachment (14–26%). The variations between the Kali Gandaki valley, the Lower Dolpo region, 706 707 as well as the other Himalayan areas, highlight the lateral variability of the regional structure, and stress 708 the need for further investigations of other areas of the belt, through microstructural and interdisciplinary analyses on the mylonitic zone before building large-scale tectonic models. 709

#### 711 **6.** Conclusion

Combining calcite microstructural analysis with regional-scale information, we reconstructed the
evolution of the ductile Annapurna Detachment zone, representing the local segment of the Himalayan
STDS in the Kali Gandaki valley. We documented:

A progressive shallowing of the Annapurna Detachment zone occurred through two consecutive
 stages of ductile shearing (D2<sub>early</sub> and a D2<sub>late</sub>), recorded by calcite microstructures. Mylonitic
 foliation, syn-kinematic mineral and calcite grain size developed during the D2<sub>early</sub>, whereas
 calcite twinning, crosscutting most calcite grains and reorienting the calcite [*c*]-axes against the
 shear sense, occurred during the D2<sub>late</sub>.

The Annapurna Detachment zone suffered a cooling from the D2<sub>early</sub> (at least 500/550-400°C
 traced up-section) to the D2<sub>late</sub> (T≤250°C) under constant kinematic vorticity in a general shear
 flow. Strain rates probably decreased overtime from the D2<sub>early</sub> to the D2<sub>late</sub>.

We interpret these two stages of shearing as representative of a shallowing of the shear zone.
 Cooling of rocks at almost constant kinematic vorticity for little decelerations enhanced the
 increase in the differential stress and the strain hardening, accommodated by carbonates in the
 ductile regime.

Compared the Annapurna Detachment zone to other segments of the STDS, we suggest that the
 regional-extended discontinuity did not experience an equal evolutionary history all along the
 Himalaya, with strain partitioning due to the different lithologies and local features. Protracted
 ductile shearing in carbonate-bearing rocks may be the cause of the lack of the upper brittle
 STDS in several Himalayan transects.

From a broader point of view, our work highlights how the behaviour of marbles in shear zones
 can determine complex and composite histories, which can be deconvolved using calcite
 microfabrics.

735

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- 1458
- 1459 Figure and Table Captions

1460

Fig. 1 – Geological map of the Nepal Himalaya, modified after Corrie and Kohn (2011). The location of
the Kali Gandaki valley is highlighted. Abbreviations: HHL, High Himalayan Leucogranite; THS, Tethyan
Himalayan Sequence; GHS, Greater Himalayan Sequence; LHS, Lesser Himalayan Sequence; SHS,
Subhimalayan Sequence; MCT: Main Central Thrust; STDS: South Tibetan Detachment System; MBT:
Main Boundary Thrust; MFT: Main Frontal Thrust.

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Fig. 2 – (a) Upper Kali Gandaki valley geological sketch map (modified after Godin, 2003 and Carosi et
al., 2016) showing the originally mapped Annapurna Detachment (AD) as solid line, the inferred highstrain zone boundaries by the dashed lines, and study sample's location as yellow dots. The GHS<sub>U</sub> Unit
definition follows Godin (2003). (b) S-N geological cross section from Lete toward Marpha (A-A' trace
in a).

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1474 Fig. 3 – Micro photos (crossed nicols) of representative samples referred to their structural position 1475 with respect to the S-N geological cross section. The red solid line (AD) indicates the Annapurna 1476 Detachment's position according to Godin et al. (1999a). (a) Massive limestone from the Nilgiri Fm. 1477 (THS), with serrated calcite grain boundaries typical of pressure solution processes. (b) Impure marble 1478 with oblique SPO, lobate boundaries, and twins in calcite. Coarse-grained biotite crosscuts the main 1479 foliation. (c-d) Marbles with calcite lobate grain boundaries, oblique SPO, and Type II e-twins. (e) Photos 1480 of sample K21-10-18 under optical microscope and in hot-cathodoluminescence (insert). (f-i) Coarse-1481 grained marbles within the GHS<sub>I</sub>. (i) Straight grain boundaries in the coarse-grained calcite within the 1482 GHS<sub>U</sub>.

Fig. 4 - Key images of part of the grain size distribution maps acquired for the analysed sample. Images
are only part of the grain size distribution maps to allow an immediate comparison between samples
for the same scale. Cal% is the measured 2D abundance of calcite in each sample. The vertical scalebar
on the left of each map correspond to the equivalent radius of each grain calculated by ImageJ software.
White hole are the second-phase minerals. The main grain size distribution histograms are plotted for
representative samples. See Table 2 for details.

Fig. 5 – Examples of analysed twin sets in calcite crystals used for paleopiezometry and
paleothermometry estimations. (a) annealed boundaries of untwined crystals are indicated. (b)
Example of high-angle boundaries due to twinning, leading to pressure solution processes. The white
arrow points tapered twins. (c) Mean twin width vs mean twin density plot after Ferrill et al. (2004).
The dashed blue and orange curves indicate paths of increasing strain for temperatures below 170°C
and above 200°C, respectively. The main twin sets point out final temperature conditions of 170-200 °C,
while the preserved tapered and thicker twins support deformation temperatures above 200 °C.

Fig. 6 – Main quartz (a-c) and calcite (d-h) pole figures (the specimens are listed from top to bottom
along the profile, reference frame is displayed at the top left). Pole figures shows that both quartz and
calcite CPO are well-defined, supporting asymmetric fabrics. See text for further details.

Fig. 7 - (a) Wenk et al. (1987) diagram for estimating simple shear contribution in marbles from calcite
CPOs. (b) adapted Law et al. (2004) graph for kinematic vorticity estimates obtained from both applied
methods. Relationship between kinematic vorticity and relative components of pure and simple shear
for instantaneous 2D flow is given. See Table 3 for details.

- Fig. 8 Fry plot with interpreted fabric ellipse. Low-density areas (vacancy field ellipse) approximate
  the finite strain equivalent ellipse. Voids are defined through the exponential edge detection method of
  Waldron and Wallace (2007). Rxz = Finite strain axial ratios for XZ sections of finite strain ellipsoid.
- Fig. 9 Schematic illustration for the Annapurna Detachment zone tectonic evolution, compared for a
  crustal strength/differential stress profile (not to scale). The black solid line refers to the strain rate at
  the D2<sub>early</sub>, whereas the dashed line refers to the D2<sub>late</sub>. Two main stages of shallowing and cooling, D2<sub>early</sub>
  (a) and D2<sub>late</sub> (b) are suggested for decreasing strain rates under subsimple shear, supporting a strain
  hardening and an increase of the differential stress.

- 1524 Table 1

Sample	Latitude, longitude	Formation (Unit)	Mineral Assemblage	Calcite deformation	Analysi s	Fabric
K21-10-05	28.735472, 83.677528	Sombre Fm. (THS)	Cal+Qz+Wm	pressure solution	0	Isotropic coarse-grained limestone
K21-10-12A	28.699472, 83.625917	Annapurna Fm. (THS)	Cal+Dol+Qz+Wm+Bt±Chl	SGR + twinning	٠	Continuous foliation (Sp), Calcite SPO
K21-10-12	28.699472, 83.625917	Annapurna Fm. (THS)	Cal+Dol+Qz+Wm±Rt	GBM + twinning	٠	Continuous foliation (Sp), Calcite SPO
K21-10-13	28.699472, 83.625917	Annapurna Fm. (THS)	Cal+Dol+Qz+Wm+Bt±Chl	GBM + twinning	0	Continuous foliation (Sp), Calcite SPO
K21-10-18	28.666917, 83.590472	Annapurna Fm. (THS)	Cal+Dol+Qz+Kfs±Bt±Wm± Chl	GBM + twinning	O	Continuous foliation (Sp), Calcite SPO
K22-10-19	28.672250, 83.597306	Annapurna Fm. (THS)	Cal+Dol+Qz+Bt+Wm	GBM + twinning	•	Continuous foliation (Sp)
K29-10-51	28.650333, 83.626722	Unit III (GHS <sub>U</sub> )	Cal Dol+Qz+Bt	GBM + twinning	•	Continuous foliation (Sp)
K29-10-52	28.650333, 83.626722	Unit III (GHS <sub>U</sub> )	Cal+Dol+Qz+Bt	GBM + twinning		Continuous foliation (Sp)
K29-10-53	28.650333, 83.626722	Unit III (GHS <sub>U</sub> )	Cal+Dol+Qz+Bt	GBM + twinning	•	Continuous foliation (Sp)
K31-10-64	28.666306, 83.588639	Unit III (GHS <sub>U</sub> )	Cal+Dol+Qz+Bt	GBM + twinning	٩	Continuous foliation (Sp), Calcite SPO
K31-10-65	28.666306, 83.588639	Unit III (GHS <sub>U</sub> )	Cal+Dol+Qz+Bt	GBM + static recrystallization	٩	Continuous foliation (Sp), Calcite SPO
<i>Legend</i> : O microstructural analysis; O paleopiezometry; O Finite-strain analysis; O CPO.						

**Table 1 –** Summary of the studied samples structural positions, features, and related type of analysis.
Abbreviations: Bt – biotite; Cal – calcite; Chl – chlorite; Dol – dolomite; Kfs – feldspar; Qz – quartz; Rt –
rutile; Wm – white mica; GBM – grain boundary migration; SGR – subgrain rotation recrystallization; Sp
– main foliation; SPO – shape preferred orientation.

- 1531
- 1532 Table 2

Sample	% Cal	AR	RMS (μm)	σ (MPa) Barnhoorn et al. (2004)	έ (GBM) (s <sup>-1</sup> ) (T=400- 500°C)	Mean twin width (µm)	Mean twin density (n/mm)	σ <sub>twin</sub> (MPa) Rybacki et al. (2011)	Ė (twin) (s <sup>-1</sup> ) (T=250°C)
K21-10-12A	65	2.24	250±30	13.7±4.8	3.1E-11	-	-	-	-
K21-10-12	65	2.45	390±180	10.2±3.6	1.7E-11	3±1	43	128±8	5.0E-15
K22-10-19	80	1.98	530±170	8.3±2.9	1.1E-11	3±1	48	135±9	9.4E-15
K29-10-51	78	1.96	490±160	8.7±3.1	1.4E-09	-	-	-	-
K29-10-52	73	2.13	440±140	9.4±3.3	1.6E-09	2±1	55	145±9	2.2E-14
K31-10-64	70	2.48	710±150	6.8±2.4	8.1E-10	4±1	42	126±8	4.4E-15
K31-10-65	75	2.41	770±200	-	-	-	-	-	-

1533

**Table 2** – Results from grain size and twin analyses. %Cal = abundance of calcite in sample; AR = aspect ratio; RMS= root mean square calcite crystal equivalent diameter;  $\sigma$  (MPa) Barnhoorn et al. (2004) = differential stress calculated from the RMS.  $\dot{\epsilon}$ (GBM) = strain rate from dynamic recrystallization mechanisms (Renner et al., 2002).  $\sigma_{twin}$  (MPa) = differential stress based on the twin density from Rybacki et al. (2011) paleopiezometer.  $\dot{\epsilon}$  (twin) = strain rate from Rutter (1974).

- 1539
- 1540 Table 3

Sample	Rxz	δ° (Sb angle)	simple shear %	Wn (Sb)	S (Rxz, Sb)	ω° CPO	simple shear % (CPO)	Wm (CPO)	S (Rxz, [ <i>c</i> ]-axis)
K21-10-12A	1.242	ND	ND	ND	ND	ND	ND	ND	ND
K21-10-12	1.322	20	41	0.64	0.92	2	5	0.08	0.87
K21-10-13	1.270	23	46	0.72	0.91	ND	ND	ND	ND
K22-10-18	ND	ND	ND	ND	ND	10	30	0.45	0.86
K22-10-19	1.403	25	49	0.77	0.90	10	30	0.45	0.86
K29-10-52	1.463	26	50	0.79	0.89	15	50	0.71	0.87

- **Table 3 –** Results from finite-strain and CPO analyses. Rxz = Axial ratios from the ellipse voids (Waldron
- and Wallace, 2007);  $\delta^{\circ}$  = angle between the oblique foliation (Sb) and the main foliation (Sp);  $\omega^{\circ}$  = angle
- 1544 between the main [c]-axes orientation and the plane normal to the foliation; Wn = sectional vorticity
- 1545 number; Wm = mean kinematic vorticity number; S = shortening.