

Thermal history of the Northern Taiwanese slate belt and implications for wedge growth during the Neogene arc-continent collision

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| 4 | Chih-Tung Chen ^{1*} , Yu-Chang Chan ² , Olivier Beyssac ³ , Chia-Yu Lu ⁴ , Yue-Gau Chen ⁴ , |
| 5 | Jacques Malavieille ⁵ , Steven Kidder ⁶ , Hao-Cheng Sun ¹ |
| 6 | |
| 7 | ¹ Department of Earth Sciences, National Central University, Zhongli, Taoyuan, |
| 8 | Taiwan, R.O.C. |
| 9 | ² Institute of Earth Sciences, Academia Sinica, Taipei, Taiwan, R.O.C. |
| 10 | ³ Institut de Minéralogie et de Physique des Milieux Condensés, Université Pierre et |
| 11 | Marie Curie, CNRS UMR7590, Paris, France |
| 12 | ⁴ Department of Geosciences, National Taiwan University, Taipei, Taiwan, R.O.C. |
| 13 | ⁵ Université de Montpellier, Géosciences Montpellier, Montpellier, France |
| 14 | ⁶ City College New York, New York City, U.S.A. |
| 15 | |
| 16 | *corresponding author, e-mail: <u>chihtung@ncu.edu.tw</u> |
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39 Abstract

A significant issue in the study of orogenic systems concerns the roles played by frontal and basal accretion in the construction of orogenic wedges. These different accretion mechanisms result in different thermal histories, with underplated materials experiencing significant heating and deformation during tectonic burial. This work provides new thermal data from Raman spectroscopy of carbonaceous material (RSCM) in combination with structural and stratigraphic observations of the northern Taiwan slate belt to address these questions of wedge development. Sedimentary rocks of the Northern slate belt were deposited on the Chinese continental margin immediately before the onset of the Neogene Taiwan arc-continent collision. In the slates of the northern Hsüehshan Range, a large-scale pop-up structure on the prowedge of the Taiwan Orogen, syn-orogenic metamorphism has been investigated through analyses of peak temperatures and metamorphic field gradients. Results indicate underthrusting of the margin sediments to ~8 km depth with significant folding in two major duplexes occurring before underplating. Such basal accretion is considered responsible for the distinct culmination of the Hsüehshan Range in central Taiwan and its relative uplift with respect to the Backbone Range to the east along the Lishan Fault. A similar underthrusting scenario is also suggested for the Backbone Range Slate Belt. We propose that basal accretion is the predominant mechanism in the growth and evolution of the Taiwan orogenic wedge, and may have been achieved through inversion of a graben system on the ancient passive margin during continental subduction.

73 1. Introduction

74 Patterns of metamorphism recorded in mountain belts reflect the underlying 75 tectonic processes driving rock deformation and exhumation (e.g. Huerta et al., 1999; 76 Jamieson et al., 2002; Rosenberg et al., 2015). One important debate related to 77 orogenic belts is the dominant mechanism by which material enters an accretionary 78 wedge. Two endmember scenarios have been proposed: frontal accretion of material 79 at the toe of the wedge (e.g. Barr et al., 1991), or basal accretion by duplexing of 80 underthrust footwall rocks along the main detachment (van Gool and Cawood, 1994; 81 Gutscher et al., 1998; Kukowski et al., 2002). These contrasting wedge formation 82 mechanisms influence a mountain belt's structure as well as its metamorphic and 83 thermal configuration (e.g. Dahlen and Barr, 1989; Bollinger et al., 2004; Simoes and 84 Avouac, 2006; Bonnet et al., 2007; Chen et al., 2011; Konstantinovskaya and Malavieille, 2011; Beyssac et al., 2016; Molli et al., 2018). Terranes composed of 85 86 frontal-accreted sediments should reveal mainly static diagenesis and retain the 87 thermal signature of the original depositional setting; whereas basal-accreted rocks are 88 expected to exhibit a strong synorogenic (dynamic) metamorphic overprint acquired 89 during the underthrusting stage (Glodny et al., 2005; Konstantinovskaya and 90 Malavieille, 2005). Regional analyses of metamorphic terranes can provide insights 91 into the contribution of these contrasting accretion pathways. Given that it is a type 92 locality for the understanding of wedge mechanics (Davis et al., 1983; Barr et al., 93 1991), it is important to determine, as best as possible, the accretion mechanisms 94 involved in the active arc-continent collision system of Taiwan.

95 The Taiwan mountain belt results from oblique convergence between the Eurasian 96 and the Philippine Sea plates. Eastward subduction of the South China Sea beneath 97 the Manila Trench at a rate of ~8 cm per year (Yu et al., 1997; Lin et al., 2010; Fig. 1 98 inset) led to the oblique collision of the Chinese Continental Margin with the Luzon 99 Arc. This process has been ongoing since the late Miocene (Suppe, 1981, 1984; Teng, 100 1990). Given its well-constrained and relatively simple tectonic setting, Taiwan offers 101 the opportunity to determine how an extended continental margin covered with a thick 102 pile of sediments (Lin et al., 2003) transforms into an arc-continent collisional prism 103 (e.g. Simoes et al., 2007, 2012; Brown et al., 2012). The Taiwanese orogenic 104 wedge-in which both the basement and cover sediments of the continental margin are bulldozed against the indenting arc-was traditionally considered to have grown 105 106 by frontal accretion with only minor (<25%) underplating via basal accretion (e.g. 107 Dahlen and Barr, 1989). It follows that thrusts, and associated fault-bend folds (Suppe, 108 1980; Yue et al., 2005), are directly rooted into a low-dipping detachment fault 109 throughout the orogen (Carena et al., 2002). This inference generally holds true for 110 the outer fold-thrust belt of the orogeny in the Western Foothills. There, the deformed

111 sedimentary successions display burial diagenesis and an absence of syntectonic 112 prograde metamorphism (Chen and Wang, 1995). However, in the interior part of the 113 mountain belt, orogen-related heating and cooling is recognized in rocks of 114 prehnite-pumpellyite facies up to lower amphibolite facies (e.g. Ernst and Jahn, 1987; 115 Lo and Onstott, 1995; Liu et al., 2001; Fuller et al., 2006; Lee et al., 2006, 2015; 116 Beyssac et al., 2007; Chen et al., 2011, 2018). The distribution of metamorphic grades 117 across the orogen does not exhibit a linear increase from the west (foreland) to the 118 east (hinterland against the backstop) as envisaged in frontal-accreted wedges (Barr et 119 al., 1991). Instead, a high metamorphic temperature anomaly is present in the 120 prowedge Hsüehshan Range (Fig. 1 inset; Beyssac et al., 2007; Simoes et al., 2007, 121 2012). Further complexity arises from the subducted Chinese continental margin, 122 which possesses numerous Paleogene rifts and grabens related to the opening of the 123 South China Sea (Teng, 1991, 1992; Teng and Lin, 2004). Hence, we find it unlikely 124 that a single, continuous master detachment surface exists beneath the entire orogen. 125 Active shortening on the island as revealed by GPS measurements (Yu et al., 1997; 126 Lin et al., 2010) is localized along the western prowedge and eastern retro-wedge toes 127 rather than being evenly distributed across the entire system as predicted by classical 128 wedge theory (Avouac, 2003; Simoes and Avouac, 2006). Additionally, the vertical 129 displacement field across Taiwan suggests that tectonic underplating is responsible for 130 the spectacular uplift in the Central Range (Simoes et al., 2007; Ching et al., 2011). 131 The above arguments have prompted studies on the nature of accretion in the Taiwan 132 wedge with results advocating substantial tectonic underplating (Fuller et al., 2006; 133 Simoes et al., 2007, 2012; Malavieille, 2010; Chen et al., 2011, 2018; Brown et al., 134 2012). Other studies, however, have questioned the presence of Neogene prograde 135 metamorphism in the exhumed continental margin basement rocks, as well as the 136 overall significance of underplating (Yamato et al., 2009; Wintsch et al., 2011).

137 In this study, Raman spectroscopy of carbonaceous material is used as a 138 geothermometer (RSCM; Beyssac et al., 2002; Lahfid et al., 2010) on Paleogene to 139 Miocene continental margin metasediments exhumed in the Taiwanese Central Range 140 slate belt. Unlike the poly-phase metamorphosed continental margin basement now 141 exposed in the Tananao Complex of the eastern Central Range (Fig. 1 inset; Yui et al., 142 1988, 2009), these rocks are too young to have experienced tectonic events prior to 143 the late Cenozoic arc-continent collision. The highest recorded temperatures of the 144 late Tertiary rocks should therefore reflect either pre-orogenic burial in margin rift 145 basins, or heating during the Neogene orogeny (Beyssac et al., 2007; Chen et al., 2011, 2018). While most past efforts have focused on the central part of the island (e.g. 146 147 Dahlen and Barr, 1989; Fuller et al., 2006; Simoes et al., 2007), we present new 148 RSCM measurements in the northern part of the slate belt along a NW-trending

149 transect, the Northern Cross-Island Highway and the Taipingshan Road (Fig. 1). 150 According to the oblique convergence tectonic framework (Suppe, 1981), the northern 151 part of Taiwan has undergone a longer collisional process than further south, and 152 exposes rocks which were once located deeper within the orogenic wedge. Below, we 153 document the history of the northern slate belt using RSCM data on peak thermal 154 conditions together with structural, stratigraphic and published thermal history 155 constraints. This approach allows a quantitative estimate of synorogenic dynamic 156 metamorphism and sheds light on the extent of basal accretion. Structures responsible 157 for underplating are then evaluated, leading to a better understanding of basin 158 inversion atop the subducted continental margin, and a refined tectonic interpretation of the Taiwan arc-continent collision. 159

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161 2. <u>Regional background</u>

162 The slate belt of Taiwan outcrops on the prowedge side of the Taiwanese orogenic 163 wedge, and is sandwiched between deformed, un-metamorphosed continental shelf 164 sediments of the Western Foothills and the exhumed continental margin basement of 165 the Tananao Complex. The slate belt comprises two north-south trending zones: the 166 Hsüehshan Range and the Backbone Range Slate Belt (Fig. 1). The Hsüehshan Range 167 contains a continuous succession of Eocene to mid-Miocene continental margin 168 sediments (Teng, 1991) deposited in eastward-deepening rift grabens (Teng, 1992) 169 that were associated with the opening of the South China Sea (Teng and Lin, 2004). 170 The Backbone Range Slate Belt is found east of the Hsüehshan Range and comprised 171 mainly of Miocene pelagic sediments. The Hsüehshan Range and Backbone Range 172 are separated by the Lishan Fault in northern Taiwan. The Backbone Range Slates are 173 juxtaposed to the east against the schist-grade Tananao Complex along a basal 174 unconformity which was reworked as a major shear zone (Suppe et al., 1976; Fisher et al., 2002). 175

176 Following the cessation of Mesozoic paleo-Pacific subduction in southeastern 177 China, the continental margin remained subaerial until the early Eocene. Subsequently, 178 grabens developed in the middle part of the margin were drowned, allowing coastal to 179 shallow marine sediments to accumulate (e.g. Lin et al., 2003). During the Oligocene, 180 the grabens deepened and widened in response to the opening of the South China Sea 181 (Briais et al., 1993; Yeh et al., 2010) and thick successions of shelf and distal marine 182 deposits accumulated (Teng, 1992). Following the latest Oligocene, marine sediments 183 draped most of the continental margin. The rifted outer-most rim of the former margin 184 (Lu and Hsu, 1992; Shyu et al., 2005) was covered by minor Eocene deposits and 185 Miocene pelagic shale sequences (Teng, 1991). During the Neogene Taiwan Orogeny 186 some of the rift grabens situated in the distal portions of the continental margin were

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187 inverted as a large-scale pop-up structure forming the Hsüehshan Range (Clark et al., 188 1993; Tillman and Byrne, 1995; Brown et al., 2012). The Backbone Range Slate Belt 189 is made up of the deformed margin rim cover sediments. Due to the oblique nature of 190 the Taiwanese convergence, the northern parts of the ranges are considered more 191 evolved than their southern counterparts. Exhumed sedimentary successions are not 192 identical along the mountain belt (Teng, 1992) indicating the involvement of different 193 basins along the continental margin.

194 From bottom to top, the meta-sediments in the northern Hsüehshan Range 195 comprise the following units: Eocene Hsitsun Formation, containing phyllitic slate 196 and intercalated meta-sandstone layers (Fig. 2C); Eocene-Oligocene Szeleng 197 Sandstone, a thick quartzitic arenite to conglomerate; Oligocene Kankou Formation, a 198 massive slate with rare sandstone; Oligocene Tsuku Sandstone, a prominent arenite; 199 interbedded argillites and sandstones of the Oligocene Tatungshan Formation; and 200 uppermost Miocene Sule Formation in which argillite alternates with calcareous 201 sandstones (Fig. 2A and B). The Sule Formation is correlated with lower to middle 202 Miocene sequences found in the un-metamorphosed Western Foothills, suggesting 203 lateral continuation of deposition and facies associations in the Miocene. Occasional 204 volcanics, mostly basaltic, occur intercalated and conformable within the succession. 205 In the eastern part of the northern Hsüehshan Range, Oligocene sediments including 206 the Kankou, Tsuku and Tatungshan formations become indistinguishable as a result of 207 eastward-deepening of the rift basin, and are collectively mapped as the 208 predominantly slate Paling Formation. East of the Lishan Fault, the Backbone Range 209 Slate Belt is composed of massive slate usually assigned as the Miocene Lushan 210 Formation and the Eocene Pilushan Formation. It contains sheared lenses of limestone, 211 sandstone, conglomerates, meta-volcanics and mafic plutonic rocks. The presence of 212 an Eocene 'formation' in the Backbone Range Slate Belt remains disputed, as the 213 reported Eocene fossils (Chang, 1972) were from tectonic blocks embedded in the 214 Miocene pelagic mud matrix (Lu and Hsu, 1992).

215 The sedimentary succession in the northern Hsüehshan Range was thrust over the 216 Western Foothills along the Chuchih Fault and deformed into a series of NE-trending, 217 NW-vergent faults and folds including the Shihtsao Fault, the Loshan Fault, the 218 Chatienshan Anticline, the Tahan Fault, and the Hsitsun Anticline (e.g. Lin and Kuo, 219 1996; Fig. 1). The Hsüehshan Range is uplifted relative to the Backbone Range along 220 the subvertical E-vergent Lishan Fault which has a complex kinematic history (Lee et 221 al., 1997). Exhumation increases eastward within the Hsüehshan Range as Miocene 222 rocks are found on the flanks of the Chatienshan Anticline and are absent southeast of 223 the Tahan Fault, while Eocene strata crop out only in the core of the Hsitsun Anticline. 224 Degree of metamorphism also increases eastward within the Hsüehshan Range as

225 slaty cleavage is developed in all rocks southeast of the Junghua Dam on the 226 northwestern limb of the Chatienshan Anticline (the 'onset of cleavage' in Fig. 1) with 227 a NE-trending, inclined to subvertical attitude (Fig. 3A), while the Oligo-Miocene 228 indurate argillite in the exterior Hsüehshan Range possesses only pencil cleavage (Fig. 229 2). Metamorphic conditions of the Hsüehshan Range span from prehnite-pumpellyite 230 grade to greenschist facies in the Hsitsun anticline (Clark et al., 1993; Chen and Wang, 231 1995; Beyssac et al., 2007; Chen et al., 2011). Muscovite K-Ar and zircon and apatite 232 fission track ages from the greenschist facies rocks are younger than 5 Ma, indicating 233 cooling during the Taiwanese orogeny (Liu et al., 2001; Fuller et al., 2006; Beyssac et 234 al., 2007).

235 Based on its thermo-kinematic history, the Hsüehshan Range can be subdivided 236 into upper and lower nappe units (Chen et al., 2011). The current study area in the 237 northern Hsüehshan Range represents the lower nappe unit, in which synorogenic 238 metamorphism is interpreted to overprint earlier lower grade diagenesis (see below). 239 The central and southern Hsüehshan Range belong to the upper nappe unit which is 240 characterized by higher grade, statically-developed diagenetic metamorphism 241 (Beyssac et al., 2007). The cleavage developed in the northern Hsüehshan Range has 242 been constrained from syn-kinematic mica porphyroblasts found in concordant mafic 243 pyroclastic units within the Tatungshan Formation to have been acquired 6 - 2.5 Ma at 244 ~250°C (Chen et al., 2018).

245 The Backbone Range Slate Belt is more penetratively deformed, with nappe 246 stacks containing gently-dipping layer-parallel foliation (Fig. 3A) indicating the 247 dominance of top-to-the-northwest shear (Lu and Hsu, 1992; Tillman and Byrne, 1995; 248 Fisher et al., 2002). Overall the Backbone Range Slate Belt to the west of the Tananao 249 Complex consists of a synform or synclinorium (Yang and Lo, 1986). The 250 metamorphic grade in the Backbone Range Slate Belt increases eastward from 251 prehnite-pumpellyite facies at its western border with higher-grade Hsüehshan Range 252 rocks across the Lishan Fault, to biotite grade greenschist facies at its eastern 253 boundary with the schist basement (Ernst and Jahn, 1987; Beyssac et al., 2007). 254 Quantitative assessment of the peak metamorphic conditions, especially the maximum 255 temperature experienced by the rocks of the northern Hsüehshan Range and Backbone 256 Range Slate Belt, is lacking (Chen et al., 2011) and addressed here by applying the 257 RSCM method.

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259 **3.** <u>Thermometry</u>

260 **3.1.** The Raman spectroscopy of carbonaceous material (RSCM) geothermometer

261 Organic matter present in rocks is transformed into graphitic carbonaceous 262 material during metamorphism in a systematic process that can be used as a 263 geothermometer (e.g. Beyssac et al., 2002, 2019). The graphitization process is 264 irreversible, such that retrograde metamorphic reactions do not affect temperature 265 estimates. The RSCM thermometer is based on the quantification of the degree of 266 ordering of carbonaceous material and has an intrinsic calibration error of ~50 °C due 267 to the petrological data used for calibration. Relative accuracy is better at about 15 °C 268 (Beyssac et al., 2004). For temperatures below 330 °C, Lahfid et al. (2010) performed 269 a systematic study of the evolution of the Raman spectrum of carbonaceous material 270 in low-grade metamorphic rocks in the Glarus Alps (Switzerland). They showed that 271 the Raman spectra of lower temperature carbonaceous material has additional peaks 272 from the spectra observed at higher temperature, and established a quantitative correlation between the degree of ordering of carbonaceous material and temperature. 273 274 Although this calibration is not universal (see discussion in Lahfid et al., 2010), we 275 use it to obtain first-order insights regarding the thermal evolution of rocks in the 276 study area.

277 At low temperatures (<330 °C), Raman spectra of carbonaceous material exhibits 278 a graphitic band composed of the G and D2 bands, a defect D composite band made 279 up of the main defect band D1 with the D4 band shoulder, and a small D3 band in 280 between (Lahfid et al. 2010; Beyssac and Lazzeri, 2012). The structural organization 281 of carbonaceous material can be quantified using the R2 and RA1 parameters which 282 reflect the proportion of the D composite band area within the entire measured spectra 283 (R2=D1/[G+D1+D2], RA1=(D1+D4)/[G+D1+D2+D3+D4], where all variables284 represent band areas). A linear correlation of RA1 with metamorphic temperature is 285 established in the range 210-330°C (Lahfid et al., 2010).

286 Raman spectra analyses were carried out at Tatung University in Taipei, and ENS, 287 Paris, both using a Renishaw (Wutton-under-Edge, UK) InVIA Reflex 288 microspectrometer equipped with a 514-mm Spectra Physics (20 mW) argon laser. A 289 DMLM Leica (Wetzlar, Germany) microscope with a $50 \times (100 \times \text{ at ENS})$ objective 290 (NA=0.90) was utilized to focus the laser on the sample. The laser power at the 291 sample surface was set to be around 1 mW using neutral density filters. The Rayleigh 292 diffusion was eliminated by notch filters, and to attain the optimal spatial resolution 293 the spectrometer entrance slit was closed down to 10 - 15 µm to reach a nearly 294 confocal configuration. The signal was dispersed using an 1800 gr/mm grating and 295 analyzed by a Peltier cooled CCD detector. Calibration of the spectrometer with a 296 silicon standard was carried out before each session. The analytical procedure 297 described in Beyssac et al. (2002, 2003) was adhered to prevent analytical pitfalls. 298 Measurements were done on polished thin sections cut perpendicular to the schistosity, 299 and the carbonaceous material was systematically analyzed below an adjacent 300 transparent mineral, usually quartz. At least ten spectra were recorded for each sample

301 to smooth out within-sample structural heterogeneity in the extended scanning mode (800 to 2000 cm⁻¹), thereby allowing clear imaging of the entire background signal 302 303 and proper defining of baseline. The acquired spectra were then processed using the 304 Peakfit software using a five-band solution with Lorentzian band profile, and the 305 parameter RA1 was calculated and utilized to estimate peak temperature (Lahfid et al., 306 2010). Spectra of the higher temperature sample NCIH-23 were analyzed using a 307 four-band solution following Beyssac et al. (2002), and the peak temperature was 308 inferred through the R2 parameter. As exhibited in Fig. 4 and described in detail 309 below, the carbonaceous material found within metasediments of the northern Taiwan 310 slate belt has a wide range of degree of graphitization similar to that reported in 311 Lahfid et al. (2010).

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313 **3.2.** Results from the northern Hsüehshan Range

314 A total of 24 samples from the Hsüehshan Range were collected along the 315 Northern Cross-Island Highway and analyzed. Sample spacing was about 1 km in the 316 northwest direction, i.e. perpendicular to the structural grain of the region. The dense 317 sampling was chosen in order to best constrain the thermal history of each strata unit 318 and structure along the transect. Most sampled rocks are massive slates and argillites, 319 along with a few thin slate layers from sandier units (e.g. the Szeleng Sandstone). 320 Representative field and microscopic images are shown in Fig. 2. Sample locations 321 and measurement results are listed in Table 1 and projected on a NW-trending profile 322 in Fig. 3B.

323 The Raman spectra of carbonaceous material of latest-Oligocene to Miocene 324 indurate argillites and siltite in the outer fringe of the Hsüehshan Range between the 325 Chuchih and Loshan faults (around Fushin and Kaopo villages) are indicative of 326 immature carbonaceous material suggesting peak temperature <200°C (samples 327 NCIH05 to 06-1). These low temperatures are consistent with the argillitic texture of 328 the samples (Fig. 2A and D). Immediately to the southeast of the Loshan Fault, the 329 temperature estimated using the RA1 parameter displays a relatively linear increase 330 from just above 200°C to around 270°C (NCIH07 to 11) at the core of the 331 Chatienshan Anticline (from Miocene Sule Formation down-section to the early 332 Oligocene Kankou Formation). The onset of slaty cleavage development corresponds 333 to an RSCM temperature of 245-265°C (NCIH07-1 and 09). The RSCM temperature 334 drops to 230 - 240°C in the slaty Sule Formation on the southern limb of the 335 Chatienshan Anticline (NCIH12 to 16; Chen et al., 2018) but the rocks still retain slaty cleavage (Fig. 2B and E). The temperature increases again to ~280°C (NCIH17, 336 337 18) in the Oligocene Paling Formation in the hanging wall adjacent to the Tahan Fault. 338 The degree of graphitization of carbonaceous material within the slates continues to

339 rise towards the core of the Hsitsun Anticline. The Oligo-Eocene Szeleng Sandstone 340 at Szeleng has an RSCM temperature of 300 - 310°C (NCIH21, 22 Fig. 2F), and the 341 Eocene Hsitsun Formation at Mingchih reaches 350°C (NCIH23; Fig. 2C). Further 342 southeast the estimated peak temperature is scattered between 260°C to 300°C 343 (NCIH25, 26, 29, and 31). Overall the RSCM temperature increases down-section 344 such that the older and deeper formations of the original sedimentary succession are 345 hotter than overlying units. RSCM temperatures also increase eastward within the 346 Sule Formation (between the limbs of the Chatienshan Anticline), and within the 347 Oligocene strata exposed in the Chatienshan Anticline and hanging wall of the Tahan 348 Fault (the Kankou-Tsuku-Tatungshan series and Paling Formation).

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0 **3.3.** Results from the northern Backbone Range Slate Belt

A total of six samples were collected from within the Backbone Range along the Taipingshan Road. The area is composed predominantly of massive slate which grade to phyllite near Cueifong Lake. A few, sometimes thick, lenses of meta-sandstone are also present. Sample locations and measurement results are listed in Table 1 and projected on the NW-trending profile shown in Fig. 3B.

356 The RSCM temperature of sample TPS01 is similar to that of samples NCIH29 357 and 31 on the southeastern end of Northern Cross-Island Highway, implying some 358 structural continuity with the Hsüehshan Range. A significant RSCM temperature 359 drop occurs in the low-hill area near the Lanyang River at Tuchang from ~270°C 360 (TPS01) to ~230°C (TPS02), coinciding with a microstructural change from 361 well-developed slaty cleavage to argillitic texture with incipient slaty cleavage (Fig. 362 2G and H). Further southeast the estimated peak temperature increases from ~230°C to ~240°C at Jhongjian (TPS04), and ~270°C to ~300°C at Taipingshan and the 363 364 Cueifong Lake (TPS08, 10 and 13) where phyllitic texture appears (Fig. 2I). The 365 increase of metamorphic temperature towards the southeast with increasing elevation 366 indicates the presence of an inverted metamorphic field gradient also seen in the 367 central Backbone Range (Beyssac et al., 2007) and likely resulting from intense 368 thrust-fold stacking of the Backbone Range slates (Lu and Hsu, 1992).

369

370 4. Discussion

4.1. New quantitative constraints on temperature and comparison with existing data

The peak temperature estimates using the RSCM method in the northern Hsüehshan Range fit well with patterns exhibited by other geological constraints (Fig. 374 3B). Zircon fission track ages are completely reset only in the Eocene and lowermost 375 Oligocene rocks along the highway; and, in contrast, younger formations yield 376 partly-reset and unreset ages (Liu et al., 2001). The Hsitsun Formation and Szeleng 377 Sandstone are thus both the oldest and highest-grade rock units in the region. The 378 trend of rising temperature between the Shihtsao Fault and the Chatienshan Anticline 379 axis is revealed in both the vitrinite reflectance (Chen et al., 2011) and RSCM data 380 sets. The metamorphic temperature assessed from the metamorphic assemblage of a 381 meta-basalt on the northwest limb of the Chatienshan Anticline (Shau and Yang, 1987) 382 is also close to RSCM and vitrinite reflectance-derived temperatures of nearby 383 samples. Illite crystallinity data (Chen et al., 1994) also exhibit a pattern that mirrors 384 the RSCM results: higher temperatures in the cores of the Chatienshan and Hsitsun 385 Anticlines relative to their flanks, and an increase in grade from hanging wall to 386 footwall across the Tahan Fault (Fig. 3B).

387 Although the thermal history of the sedimentary succession in the northern 388 Hsüehshan Range had already been investigated using illite crystallinity (Chen et al., 389 1994), zircon fission track dating (Liu et al., 2001), and vitrinite reflectance (Chen et 390 al., 2011), the RSCM data reported here allows a complementary, locally more precise, 391 documentation of peak temperature which conforms well with geological 392 observations. Illite crystallinity data revealed a thermal trend similar to RSCM, but 393 failed to identify the increasing metamorphic temperatures in the Miocene rocks 394 southeast of the Chatienshan Anticline axis. An increase in temperature there is also 395 suggested by the occurrence of penetrative slaty cleavage absent from argillitic 396 counterparts to the northwest that contain only pencil cleavage (Fig. 3B). The vitrinite 397 reflectance (VR) data clearly indicate the greater extent of heating in the Miocene 398 strata southeast of the Chatienshan Anticline than in the exterior portion of the 399 Hsüehshan Range where peak temperatures are 200°C or lower. The diverse peak 400 heating conditions experienced by the rocks east of the Chatienshan Anticline axis 401 was therefore less discriminated in the VR data, and the peak temperatures from VR 402 are exaggerated west of the anticline axis (Fig. 3B). Such discrepancy between VR 403 and RSCM datasets probably arise because the VR calibrations (Barker, 1983; Barker 404 and Goldstein, 1990) applied in Chen et al. (2011) for all the VR data southeast of the 405 cleavage front were poorly constrained, with few constraints for Ro over 3. 406 Consequently, the VR-temperature calibration is saturated in this temperature range 407 (above ~200-250°C) and becomes insensitive. The new RSCM temperatures based on 408 the calibration of Lahfid et al. (2010) provide more realistic and reliable peak 409 temperature estimates that agree with structural observations (Fig. 2). In summary, 410 RSCM thermometry within the meta-sediments appears to sensitively reflect 411 variations of maximum temperature, and is therefore useful for estimating peak 412 burial/metamorphic temperatures in the range ~200°C to ~700°C (Beyssac et al., 2002; 413 Lahfid et al., 2010). When coupled with vitrinite reflectance data for rock units heated 414 under ~200°C, the carbonaceous material geothermometers form valuable tools in

415 understanding thermal metamorphism across orogens and basins.

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417 **4.2.** *Extent of synorogenic metamorphic overprint*

418 The eastward- and down-section-increasing pattern of peak temperature in the 419 Hsüehshan Range may be interpreted as a combination of static (basin burial) and 420 dynamic (orogen-related) metamorphism. This is best illustrated by the thermal 421 evolution of the Miocene strata. At the start of the Miocene, the Chinese continental 422 margin switched from rifting to drifting, which caused shelf-wide thermal subsidence 423 and facilitated a widespread draping of laterally-continuous margin sediments over 424 the more heterogeneous older units (Lin et al., 2003; Teng and Lin, 2004). The 425 Miocene sequence, which lacks abrupt lateral facies changes, can be correlated from 426 the Taiwan Strait all the way through the Coastal Plain, and Western Foothills, to the 427 Hsüehshan Range, then as thinner pelagic shale in Backbone Range Slate Belt (Teng, 428 1991). Considering the depositional configuration and timing of deposition just prior 429 to the inception of the Neogene arc-continent collision (foreland development at ~ 6.5 430 Ma; Lin and Watts, 2002), these c. 2 km thick Miocene strata most likely underwent 431 only minor diagenetic heating prior to structural disruption. For example, even where 432 their stratigraphic equivalents in the Coastal Plain and the Western Foothills were 433 later covered by foreland molasse sediments (Lin and Watts, 2002; Simoes and 434 Avouac, 2006), Miocene sediments were not heated above 100°C (e.g. Chen and 435 Wang, 1995; Kuo, 1997). The ~240°C peak temperature (~200°C temperature 436 increment) recorded by RSCM in the Miocene Sule Formation along the Northern 437 Cross Island Highway is therefore best explained by orogenic loading, since it is 438 equivalent to unmetamorphosed Miocene Mushan to Nankang formations in the 439 Western Foothills (e.g. Teng, 1991). Assuming a geothermal gradient of 440 25~30°C/km, ~8 km of overburden would be needed to match the observations. This 441 geothermal gradient is consistent with the change of 25-30°C/km in RSCM 442 temperature with stratigraphic depth in the central HR range (Beyssac et al., 2007), 443 the average thermal gradient in exploration wells in Taiwan (Zhou et al., 2003), and it 444 yields reasonable results in thermomechanical models (Simoes et al., 2007) and 445 thermobarometry (e.g. Kidder et al., 2013a).

446 Depositional patterns of the Eo-Oligocene strata in the Hsüehshan Range indicate 447 isolated eastward-deepening half-grabens with considerable regional variation (Teng 448 and Lin, 2004), much different from their Miocene counterparts. The eastward rise of 449 RSCM temperatures in these rocks may correspond in part to increased basin burial 450 depths (Fig. 3B), thus the relative roles of static and dynamic metamorphism cannot 451 be clearly discerned for these rocks.

452 Additional salient information regarding the extent of the synorogenic

453 metamorphic overprint in the northern Hsüehshan Range metasediments is the 454 anomalously low apparent geothermal gradient derived from RSCM data. Along the 455 investigated Northern Cross-Island Highway transect, the sediment succession from 456 the Eocene Hsitsun Formation to the Miocene Sule Formation is estimated to be about 457 6 km thick (Teng, 1991). The apparent paleothermal field gradient calculated from the 458 RSCM temperatures in the region is below 20°C/km for both the SE limb of the 459 Chatienshan Anticline (based on values of ~270°C in the lower Oligocene Kankou 460 Formation and ~240°C in the Miocene Sule Formation) and from Paling to the core of 461 the Hsitsun Anticline (using values of ~350°C in the Eocene Hsitsun Formation to 462 ~240°C in the Miocene Sule Formation; Fig. 3B). Such values are improbably small 463 for an actively rifting continental margin, since for example, the contemporary 464 geothermal gradient measured and integrated in the Qiongdongnan and Pearl River 465 Mouth basins along the northern border of the South China Sea is mostly above 466 30°C/km (Yuan et al., 2009). We conclude that the apparent geothermal gradient in 467 northern Hsüehshan Range has thus been reduced by about one-third its original value. 468

469 A likely mechanism for the low apparent geothermal gradient is folding that 470 predates peak metamorphism. This results in fold cores with temperatures more 471 similar to temperatures in their limbs than would occur if metamorphism predated 472 folding. We postulate that the presently-exposed continental margin sediment 473 sequence of ~6 km thickness was substantially folded while being underthrust to ~8 474 km depth where it experienced peak metamorphism, while the vertical distance 475 between the top and bottom of this segment (i.e. Miocene Sule and Eocene Hsitsun 476 formations) had been reduced to ~4 km among the anticline and syncline cores (Fig. 5 477 A-B) to produce the low, $\sim 20^{\circ}$ /km apparent geothermal gradient. A simple model of 478 fold limbs tilted at $\sim 45^{\circ}$ prior to peak metamorphism could reduce the apparent 479 geothermal gradient by one-third, i.e. from 30 to 20°/km. Such treatment corresponds 480 with ~33% shortening and ignores the effects of potential vertical thickening and 481 time-dependent aspects of heat diffusion related to the likelihood that folding and 482 metamorphism partly overlapped in time. Assuming that peak metamorphism roughly 483 coincided with peak burial and underplating, a minor amount of folding and 484 shortening followed underplating in order to bring fold limbs to their present 485 geometries ($\sim 55^{\circ}$ dip, Fig. 5C).

We find it unlikely that post-metamorphic thickening significantly contributed to lowering the apparent geothermal gradient because of the absence of structural or microstructural evidence for this process. For example, there does not appear to be a greater abundance of minor faults or internal penetrative deformation in the North relative to the central Hsüehshan Range, where the apparent geothermal gradient is 491 similar to the interpreted ancient one (Beyssac et al., 2007). Undetected penetrative thickening is, in any case, less likely in the North due to low temperatures that 492 493 inhibited deformation mechanisms such as dislocation creep that facilitated 494 penetrative strain in the central Hsüehshan Range (e.g. Kidder et al., 2012). This 495 interpretation of folding prior to metamorphism in the Northern Hsüehshan Range 496 differs from what is interpreted in the central Hsüehshan Range, where a ~30°C/km 497 gradient is preserved, and folding and faulting were interpreted to postdate peak 498 heating (Beyssac et al., 2007).

499 The RSCM results in the northern Backbone Range Slate Belt along the 500 Taipingshan Road transect are also indicative of dynamic metamorphism overprinting 501 basinal diagenesis. As argued above for the Miocene sediments of the Hsüehshan 502 Range, the Miocene Lushan Formation was incorporated into the Taiwan orogenic 503 pile soon after its deposition, giving little time for diagenetic heating prior to 504 deformation. Peak metamorphic temperatures range from ~230°C in 505 Tuchang/Jhongjian to over 290°C around Cueifong Lake. These temperatures also 506 indicate heating beneath ~8 km of tectonic overburden if a geothermal gradient of 507 ~30°C/km is assumed. The Lushan formation originated as pelagic sediments 508 covering the rifted/drifted Chinese continental margin basement, and sedimentary 509 criteria indicate that the present thickness of the original Lushan sedimentary 510 sequence was likely thinner than the modern 2 km thickness (e.g. Teng, 1992), thus 511 the Lushan sediments are interpreted as including several duplexes. The presence of 512 an inverted metamorphic field gradient in an accretionary setting can also be well 513 explained by a significant component of deformation concurrent with peak 514 metamorphic conditions (e.g. Kidder et al., 2013b).

515

4.3. Nappe stacking structure of the Taiwanese slate belt and implications for wedge kinematics

518 The new RSCM and structural data reported in the present study, along with 519 existing geothermometric, thermochronological, and structural data of the entire slate 520 belt (e.g. Clark et al., 1993; Liu et al., 2001; Fuller et al., 2006; Beyssac et al., 2007; 521 Brown et al., 2012; Simoes et al. 2012; Chen et al., 2011, 2018) permit a 522 comprehensive evaluation of the geological structures in this part of the orogenic 523 wedge and their potential origin. The Lishan fault-which moved the Hsüehshan 524 Range up relative to the Backbone Range—is poorly exposed in the study area, but its 525 presence is substantiated by an RSCM temperature gap of more than 40°C in the 526 studied transect (Fig. 3B), as well as by structural analyses (Lee et al., 1997; Brown et 527 al., 2012). The enhanced exhumation on the Hsüehshan Range side of the Lishan 528 Fault is consistent with inversion of the Hsüehshan Trough basin controlled by the 529 geometry of the underlying basement. In this scenario, the basement high east of the 530 Hsüehshan Trough (inherited from Paleogene rifting related to the opening of the 531 South China Sea) pushed the Hsüehshan Range sequences from the east when it 532 subducted beneath the accretionary prism, and evolved into the Tananao Complex 533 now exposed in eastern Central Range. We interpret that the basement-Hsüehshan 534 Range contact developed into a major backthrust, the Lishan Fault, while the Tananao 535 Complex is not yet exposed in the footwall due to being covered by the Lushan 536 Formation (Fig. 6B).

537 The synorogenic origin of peak metamorphism hypothesized in the previous 538 section strengthens the argument that the northern Hsüehshan Range entered the Taiwan orogenic wedge via basal accretion (as the 'lower nappe' of Chen et al., 2011) 539 540 rather than frontal accretion. The present peak metamorphic temperature 541 configuration indicates that the lower nappe was underthrust beneath the then 542 accretionary prism (to ~8 km depth) and substantially folded before been incorporated 543 into the orogenic wedge and exhumed. By combining RSCM temperatures, structural 544 data (from km-scale faults and folds, to preliminary microstructural observations), and 545 stratigraphic data, this basal-accreted terrain can be subdivided into two thrust 546 duplexes/nappes (Fig. 6A): the exterior duplex being is the hanging wall of the 547 Shihtsao Fault; the interior one the hanging wall of the Tahan Fault. The exterior 548 duplex consists of relatively younger and more proximal margin sediments (the early 549 Oligocene Kankou Formation to the Miocene Sule Formation) and exhibits generally 550 lower RSCM temperatures (Fig. 3B). This exterior duplex has a simple, uniformly 551 SE-dipping slaty cleavage that likely originated in the inner/proximal portion of the 552 Hsüehshan Trough half-graben system (Teng, 1991). In contrast, the interior duplex is 553 composed of the older and more distal Eo-Oligocene sequence from the Hsitsun 554 Formation to the Paling Formation (the temporal equivalent of the more-proximal 555 Kankou, Tsuku and Tatungshan formations). The interior duplex also has higher 556 RSCM temperatures, and stronger cleavage development and overall deformation. 557 These lower nappe duplexes were basally accreted into the wedge at depths of about 8 558 km, as determined above based in part on their RSCM temperatures.

559 The Backbone Range Slate Belt can be characterized as a synclinorium (Yang and 560 Lo, 1986), and nappe stack in which the original sedimentary succession has been 561 multiply folded and transposed, with severely sheared and boudinaged tectonic blocks 562 of various lithologies locally defining broken formations (Lu and Hsu, 1992). 563 Synorogenic prograde heating occurred when these pelagic sediments were 564 underthrust along with the margin basement beneath the accretionary prism. The rapid 565 southeastward rise of peak temperature determined by RSCM, along with the 566 transition from slate to phyllite, indicates the progressive exhumation of more deeply

buried units to the east. These were brought up by reverse faulting and shortening
related to the duplexing of the Tananao Complex to the east (the purple unit in Fig. 6B;
Fisher et al., 2002; Beyssac et al., 2007; Simoes et al., 2007; Malavieille, 2010;
Malavieille et al., 2019), where a second, deeper underplating zone is likely located
(Simoes et al., 2007). The rise of RSCM temperature along the Taipingshan Road
likely corresponds to cryptic reverse faults which root into the basal thrust (Fig. 6A).

573 The pre-convergence setting of the Chinese continental margin is illustrated in Fig. 574 6C, in which the difference between Paleogene (Eo-Oligocene) and Miocene 575 depositional patterns is highlighted. In the initial stage of collision, the rifted 576 continental margin sliver east of the Hsüehshan Trough half-graben (Lu and Hsu, 577 1992; Shyu et al., 2005) and its overlying Miocene pelagic sediments, were 578 underthrust and buttressed against the leading edge of the Philippine Sea Plate. This 579 caused intense deformation and heating of the Miocene sequence (the present-day 580 Backbone Range slate). This couplet of deformed proto-Tananao Complex and 581 metamorphosed cover sequence then bulldozed the more inboard graben sediments as 582 convergence continued. The deepest and easternmost distal succession within the 583 trough was driven towards the foreland, forming the 'upper nappe' unit (Chen et al., 584 2011) now exposed in south-central Hsüehshan Range. This foreland-ward transport 585 of the statically metamorphosed (Beyssac et al., 2007) 'upper nappe' unit was 586 synchronous with tectonic underplating and duplexing of the underthrust proximal 587 Hsüehshan Trough sediments that make up the 'lower nappe' units described in this 588 study. The timing of this event was $\sim 6 - 2.5$ Ma (Chen et al., 2018), and it culminated 589 with the uplift and exhumation of the present day Hsüehshan Range.

590 This description implies the presence of multiple active detachment faults in the 591 formation and growth of the Taiwanese accretionary wedge (Malavieille, 2010; 592 Malavieille et al., 2019). The dynamically metamorphosed lower nappe unit is able to 593 crop out in the northern Hsüehshan Range because the northern region has 594 experienced a longer history of uplift and denudation due to oblique and 595 southward-propagating convergence (Suppe, 1981). The upper nappe unit in the north, 596 which presumably resembled the central Hsüehshan Range rocks, has eroded away. 597 Based on the thermokinematic history of the lower nappe unit rocks described above, 598 it is likely that shortening of the Hsüehshan Range rocks likely continues at depth 599 within the orogenic wedge. Such shortening is completed before reaching the surface, 600 since the GPS velocity field (e.g. Yu et al., 1997; Lin et al., 2010) demonstrates that 601 the Hsüehshan Range is extensional at the surface. The Backbone Range slate, 602 however, comprising rift basement cover sediments, continues to be faulted, folded 603 and uplifted together with the thrust stacking and doming of the Tananao Complex 604 (Fisher et al., 2002; Malavieille, 2010; Yui and Chu, 2000).

605 The temperature-time and structural evolution of the Taiwanese slate depicted 606 above suggests that the Neogene Taiwan arc-continent collision involved significant 607 dynamic heating, particularly by tectonic overloading of underthrusted rocks. Since 608 the cover sediments are found to have experienced prominent synorogenic 609 metamorphism, structurally lower basement rocks are likely to have as well. From this 610 perspective, we find the hypothesis that the Taiwan Orogen experienced little 611 prograde metamorphism, particularly in the Tananao Basement Complex (Yamato et 612 al., 2009; Wintsch et al., 2011), unlikely. In addition to inferences from the 613 metamorphism of overlying sedimentary sequences, heating associated with the 614 Neogene orogeny in the Tananao Basement Complex has been firmly documented in 615 analyses of thermochronologic records (e.g. Lo and Onstott, 1995; Fuller et al., 2006) 616 and is supported by thermal metamorphic modeling (Beyssac et al., 2007; Simoes et 617 al., 2012). Our results also suggest that basins atop highly extended continental 618 margins are able to subduct as a whole to upper-mid crustal depths before being 619 inverted and exhumed. This may have implications for the rheological and tectonic 620 evolution of subducting continents in arc-continent and continent-continent collision 621 zones.

622

623 5. Concluding remarks

624 New RSCM temperature data in northern Taiwan provide evidence of synorogenic 625 metamorphism related to the recent arc-continent collision. Using thermal, structural, 626 and stratigraphic data, we postulate that synorogenic metamorphosed slates in the 627 Hsüehshan Range were underthrust to about 8 km depth and thickened via folding 628 before being basal accreted into the orogenic wedge as two major duplexes. The slates 629 in the northern Backbone Range Slate Belt were tectonically loaded under similar 630 conditions and developed an inverted metamorphic field gradient while being 631 exhumed with the basement Tananao Complex. The inferred deformation was likely 632 accommodated along multiple, possibly simultaneously active, detachment levels. 633 Tectonic underplating is a key mechanism, along with frontal accretion and surface 634 erosion, in the growth, uplift and deformation of the Taiwanese orogenic wedge. 635 Prograde heating is an integral part of the Neogene orogeny and should be considered 636 in kinematic reconstructions.

637

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871 Figure captions

- Fig. 1: Geologic strip map along the investigated Northern Cross-Island Highway
 (Hsüehshan Range) and Taipingshan Road (Backbone Range Slate Belt) transect.
 Inset shows tectonic framework of Taiwan with the slate belt highlighted.
 Locations of the RSCM samples are marked in yellow and listed in Table 1.
- 876

877 Fig. 2: (A~C) Macroscopic occurrence of outcrops: (A) sand-shale interbedded 878 argillite with only faint pencil cleavage in the Miocene series northwest of 879 cleavage front (site NCIH05-1), width of view ~1 m; (B) meta-siltstone with 880 spaced pressure-solution foliation near NCIH15 in the Miocene series on the 881 southeastern limb of the Chatienshan Anticline; (C) Eocene slate with strong 882 penetrative cleavage at NCIH23. (D~I) Microscopic observation of samples, with 883 1 mm white scale bars: (D) non-foliated latest Oligocene argillite/siltstone 884 (NCIH07); (E) latest Oligocene argillite/slate with initial cleavage development 885 through pressure solution (NCIH12); (F) Eocene meta-sandstone with sutured 886 quartz grains packed between well-developed cleavage domains (NCIH22, 887 cross-polarized view); (G) Oligocene slate with penetrative slaty cleavage 888 (TPS01); (H) Miocene slate east of the Lishan Fault with patchy pressure solution seams in initial slaty cleavage development (TPS02); (I) Miocene phyllitic slate 889 890 with dense foliation, growths of quartz porphyroblasts and mica-chlorite 891 aggregates (TPS13).

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Fig. 3: (A) Geologic profile of the studied transect along the Northern Cross-Island
Highway (NCIH) and Taipingshan Road (TPS) (location marked in Fig. 1),
distribution of the RSCM samples (projected), and structural data acquired in this
study. (B) Calculated maximum temperature from RSCM measurements of the
transect profile, along with published data of vitrinite reflectance (Chen et al.,
2011), illite crystallinity (Chen et al., 1994), zircon fission track (Liu et al., 2001)
and igneous petrology (Shau and Yang, 1987).

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Fig. 4: Raman spectra of CM obtained in the Hsüehshan Range along the Northern
Cross-Island Highway (NCIH-06, 07, 09, 17, 20, 21, 23) and the Backbone Range
Slate Belt (TPS-02, 04) with the D and G composite bands indicated. The RA1/R2
ratio and corresponding RSCM temperature calculated for each of the depicted
spectrum are specified.

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Fig. 5: Cartoon of the proposed thermokinematic history of the 'lower nappe unit'metasediments between the Shihtsao and Lishan faults along the Northern

Cross-Island Highway in the northern Hsüehshan Range: (I) pre-orogenic setting of the strata in a ~6 km deep basin on the rifted continental margin; (II) peak thermal/metamorphic state when the continental margin was incorporated in the orogenic system and the northern Hsüehshan units were overridden by the accretionary prism, presumably the 'upper nappe unit' now found in the central-southern Hsüehshan Range; (III) retrograde exhumation after basal accretion with continued shortening. Please refer to section 4.2 for details.

Fig. 6: (A) Interpreted structural profile of the northern Taiwan slate belt along the studied road transects. (B) Proposed structural architecture across the northern Taiwan mountain belt. Location is shown in Fig. 1 inset. Detailed duplex structure and thrust stacking of the Hsüehshan and Backbone ranges are enlarged and illustrated in inset. (C) Pre-orogen configuration of the involved sediment successions on the Chinese continental margin (dotted pink denotes margin basement) and the inferred material transport paths of these units (after Teng and Lin, 2004; Simoes et al., 2007; Chen et al., 2011, 2018).

947 <u>Table</u>

Table 1: RSCM temperature data obtained along the Northern Cross-Island Highwayand the Taipingshan Road. Temperatures are calculated following Beyssac et al.

950 (2002) and Lahfid et al. (2010).

| Sample | Longitude | Latitude | Altitude,m | Spectra # | RA1/R2 ratio | sdv | T,°C | 1σ, °C | se, °C |
|-------------|---------------------------------|------------|------------|-----------|--------------|-------|------|--------|--------|
| Northern Cr | oss Island H | Highway ti | ransect | | | | | | |
| NCIH05 | 121.3647 | 24.7900 | 389 | 9 | 0.50 | 0.011 | <200 | | |
| NCIH05-1 | 121.3580 | 24.7822 | 363 | 7 | 0.49 | 0.008 | <200 | | |
| NCIH06 | 121.3603 | 24.7721 | 487 | 12 | 0.52 | 0.008 | <200 | | |
| NCIH06-1 | 121.3474 | 24.7592 | 499 | 11 | 0.52 | 0.011 | <200 | | |
| NCIH07 | 121.3494 | 24.7530 | 510 | 12 | 0.55 | 0.002 | 211 | 3 | 0.87 |
| NCIH07-1 | 121.3564 | 24.7504 | 587 | 17 | 0.57 | 0.004 | 244 | 5 | 1.21 |
| NCIH09 | 121.3567 | 24.7286 | 582 | 10 | 0.59 | 0.004 | 263 | 5 | 1.58 |
| NCIH10 | 121.3551 | 24.7151 | 557 | 17 | 0.59 | 0.004 | 273 | 5 | 1.21 |
| NCIH11 | 121.3622 | 24.7088 | 606 | 15 | 0.59 | 0.003 | 268 | 4 | 1.03 |
| NCIH12* | 121.3605 | 24.6998 | 632 | 14 | 0.57 | 0.004 | 246 | 5 | 1.34 |
| NCIH13 | 121.3437 | 24.6991 | 654 | 14 | 0.56 | 0.004 | 225 | 5 | 1.34 |
| NCIH15 | 121.3812 | 24.6804 | 563 | 18 | 0.56 | 0.004 | 236 | 5 | 1.18 |
| NCIH16 | 121.3923 | 24.6747 | 676 | 10 | 0.57 | 0.004 | 247 | 5 | 1.58 |
| NCIH17 | 121.4045 | 24.6747 | 685 | 16 | 0.60 | 0.004 | 277 | 5 | 1.25 |
| NCIH18 | 121.4076 | 24.6717 | 730 | 12 | 0.60 | 0.004 | 283 | 5 | 1.44 |
| NCIH19 | 121.4098 | 24.6534 | 903 | 11 | 0.59 | 0.003 | 271 | 4 | 1.21 |
| NCIH20 | 121.4173 | 24.6453 | 919 | 14 | 0.61 | 0.004 | 292 | 5 | 1.34 |
| NCIH21 | 121.4328 | 24.6486 | 1170 | 15 | 0.62 | 0.004 | 309 | 5 | 1.29 |
| NCIH22 | 121.4418 | 24.6454 | 1112 | 11 | 0.62 | 0.003 | 302 | 4 | 1.21 |
| NCIH23 | 121.4517 | 24.6404 | 1121 | 15 | 0.65 | 0.022 | 351 | 10 | 2.58 |
| NCIH25 | 121.4838 | 24.6424 | 1203 | 13 | 0.61 | 0.003 | 298 | 4 | 1.11 |
| NCIH26 | 121.4818 | 24.6316 | 1203 | 13 | 0.61 | 0.004 | 293 | 5 | 1.39 |
| NCIH29 | 121.4942 | 24.6057 | 651 | 14 | 0.59 | 0.002 | 266 | 3 | 0.80 |
| NCIH31 | 121.5126 | 24.5980 | 412 | 14 | 0.60 | 0.003 | 281 | 4 | 1.07 |
| Taipingshan | Road trans | ect | | | | | | | |
| TPS01 | 121.4945 | 24.5768 | 394 | 17 | 0.59 | 0.003 | 269 | 4 | 0.97 |
| TPS02 | 121.5000 | 24.5600 | 504 | 24 | 0.56 | 0.004 | 227 | 5 | 1.02 |
| TPS04 | 121.5127 | 24.5458 | 739 | 17 | 0.57 | 0.004 | 239 | 5 | 1.21 |
| TPS08 | 121.5242 | 24.5110 | 1844 | 12 | 0.60 | 0.004 | 275 | 4 | 1.15 |
| TPS10 | 121.5372 | 24.4963 | 1915 | 12 | 0.61 | 0.004 | 291 | 5 | 1.44 |
| TPS13 | 121.6075 | 24.5110 | 1883 | 11 | 0.61 | 0.004 | 296 | 4 | 1.21 |
| 951 | *: data from Chen et al., 2018. | | | | | | | | |

Figure 1.



Figure 2.



Figure 3.



Bedding – Hsuehshan Range







Cleavage – Hsuehshan Range

Cleavage – Backbone Range

Epizone

Figure 4.



Figure 5.

(A) Pre-collision setting

(B) Peak state

Figure 6.

