1 Tectonothermal evolution in the core of an arcuate fold and 2 thrust belt: the southeastern sector of the Cantabrian Zone 3 (Variscan belt, NW Spain)

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7   **Abstract.** The tectonothermal evolution of an area located in the core of the Ibero-Armorican arc (Variscan belt)
8   has been determined by using the conodont color alteration index (CAI), Kübler index of illite (KI), the Árkai
9   index of chlorite (AI), and the analysis of clay minerals and rock cleavage. The area is part of the Cantabrian
10 Zone (CZ), which represents the foreland fold and thrust belt of the orogen. It has been thrust by several large
11 units of the CZ, what resulted in the generation of a large amount of synorogenic Carboniferous sediments. CAI,
12 KI and AI values show an irregular distribution of metamorphic grade, independent of stratigraphic position.
13 Two tectonothermal events have been distinguished in the area. The first one, poorly defined, is mainly located
14 in the northern part. It gave rise to very low-grade metamorphism in some areas and it was associated with a
15 deformation event that resulted in the emplacement of the last large thrust unit and development of upright folds
16 and associated cleavage (S_1).The second tectonothermal event gave rise to low-grade metamorphism and
17 cleavage (S_2) crosscutting earlier upright folds in the central, western and southern parts of the study area. The
18 event continued with the intrusion of small igneous rock bodies, which gave rise to contact metamorphism and
19 hydrothermal alteration. The second event was linked to an extensional episode due to a gravitational instability
20 at the end of the Variscan deformation. This tectonothermal evolution occurred during the Gzhelian-Sakmarian.
21 Subsequently, several hydrothermal episodes took place, in association with local development of crenulation
22 cleavage during the Alpine deformation.

23   1 Introduction

24 The Variscan belt defines an arc in the northwestern Iberian Peninsula (Ibero-Armorican Arc), whose core is
25 formed by the Cantabrian Zone (CZ), which represents the foreland fold and thrust belt of the orogen (Fig. 1).
26 This zone consists of Palaeozoic rocks in which two tectonostratigraphic units have been distinguished (Julivert,
27 1978; Marcos and Pulgar, 1982), whose limit is approximately located by the Devonian-Carboniferous
28 boundary. The preorogenic unit is formed of Cambrian to Devonian rocks consisting of alternating carbonate and
29 siliciclastic formations; they form a wedge that thins towards the foreland. The synorogenic unit is formed by
30 several clastic units, also thinning towards the foreland, which filled foredeep basins generated in the front of the
31 main thrust units of the CZ. In this zone, the Variscan deformation occurred during the upper Carboniferous and
32 gave rise to thin-skinned tectonics, with several large thrust units and associated folds (Fig. 1); the units were
33 emplaced in a sequence towards the foreland. The deformation occurred under shallow crustal conditions, so that
34 diagenetic conditions are dominant in the zone and absence of cleavage in the rocks is also dominant.
35 Nevertheless, there are several areas of the CZ where cleavage and very low- or low-grade metamorphism are

present. One of these areas is the southeastern sector of the Cantabrian Zone. It is a foreland basin that occupies the core of the Ibero-Armorican Arc, and has undergone, a complex history of sedimentation, deformation, metamorphism, and to a lesser extent, magmatism.

The present study aims to present a model of tectonothermal evolution for the core of the Ibero-Armorican arc based on conodont colour alteration index (CAI), the Kübler index (KI) of illite, the Árkai index (AI) of chlorite, the analysis of clay minerals and the rock cleavage development.

2 Geological setting

The southeastern sector of the Cantabrian Zone is made up of two units: the Pisuerga-Carrión unit (PCU) and the Valsurbio unit (VU) (Fig. 1). As a consequence of its location in the core of the Ibero-Armorican arc, the PCU has been thrust over successively by the VU, the Central Coal Basin, the Ponga unit and the Picos de Europa unit, (Fig. 1). Thus, the PCU operated as a foreland basin during a great part of the history of the Variscan deformation. This history involved the accumulation of a great thickness of synorogenic Carboniferous sediments, corresponding to clastic wedges associated with the exhumation of the thrust units.

The VU is located to the south of the PCU. Marine facies occur in both units, but from the Emsian, the sediments in the latter unit were deposited in deeper waters than those in the former unit. The VU represents an extension of the southern part of the CZ, and presents a Devonian succession comparable to that of that part (Koopmans, 1962). The units containing Silurian-Devonian rocks located within the PCU have been interpreted as transported from internal areas of the orogenic belt, specifically from the southeastern extension of the Westasturian-Leonese Zone, currently hidden under the Mesozoic-Cenozoic cover of the Duero basin, and located to the south of the VU (Frankenfeld, 1983; Marquínez and Marcos, 1984; Rodríguez Fernández and Heredia, 1987; Rodríguez Fernández, 1994). These units were named 'Palentine nappes' by Rodríguez Fernández and Heredia (1987). The PCU and the VU are separated by the León fault (also called 'Ruesga fault' in the study area) (Fig. 2), that traverses most of the Cantabrian Zone and whose meaning is controversial (de Sitter, 1962; Marcos, 1968, 1979; Julivert et al., 1971; Kullmann and Schönenberg, 1978, 1979; Heward and Reading, 1980; Aller, 1986; Rodríguez Fernández and Heredia, 1988; Nijman and Savage,1989; Rodríguez Fernández, 1991; Alonso et al., 2009, 2012).

The oldest sediments of the study area are Silurian (Wenlock-Pridoli) sandstones and lutites. The Devonian rocks consist of an alternation of carbonate and siliciclastic formations with the facies differences cited above. Mississippian rocks are mainly limestones in the lower part, especially in the VU; upwards a mostly turbiditic sequences appears with common olistoliths in the northern sector and some carbonate levels. The Pennsylvanian succession is dominantly siliciclastic, with a thickness of several thousand meters and synorogenic character. In relation to this synorogenic sedimentation, several syntectonic unconformities have been described, among which the Curavacas unconformity (early Moscovian) can be highlighted by its structural significance (Van Veen, 1965; Lobato 1977; Alonso and Rodríguez Fernández, 1983; Martín-Merino et al., 2014).

The first deformation events were pre-Curavacas (prior to or earliest Moscovian) and involved the emplacement of the Palentine nappes and the VU. Two generations of thrusts occurred during this episode (Rodríguez Fernández, 1994), which translated sequences northwards. Further, some back thrusts and normal faults occurred. Folds associated with the thrusting also formed during this episode.

74   The post-Curavacas deformation events involved the development of several generations of thrusts and
75   high-angle reverse faults, folds, cleavages and normal faults. Some thrusts have a trend approximately parallel to
76   the basal thrust of the Ponga unit and are probably related to the emplacement of this thrust unit (Rodríguez
77   Fernández and Heredia, 1987), which occurred during the late Moscovian. In the same episode that involved the
78   emplacement of the Picos de Europa thrust unit during the Kasimovian-Gzhelian (Merino-Tomé et al. 2009), N –
79   S shortening occurred in the study area, involving the development of thrusts, high angle reverse faults and the
80   reactivation of older faults with a movement dominantly southward (Maas, 1974). In addition, upright folds with
81   E – W axial trace developed; among them, the Curavacas-Lechada syncline is remarkable for its notable
82   dimensions (Savage, 1967; Lobato, 1977; Rodríguez Fernández, 1994).

83   Several cleavages have been recognised in the study area. Among them, a gently dipping cleavage
84   crosscutting folds is the most relevant and has been described by various authors (Van Veen, 1965; Savage,
85   1967; Lobato, 1977; Van der Pluijm and Kaars-Sijpsteijn, 1984;Van der Pluijm et al., 1986; Rodríguez
86   Fernández, 1994, 2001; García-López et al., 1999, 2007; Marín, 1997; Bastida et al., 2002).

87   A subsequent N-S shortening episode occurred during the Alpine deformation. It involved tightening of
88   folds, local development of crenulation cleavage and reactivation of some faults. It is responsible for the dome
89   geometry of the VU (Marín et al., 1995; Marín, 1997). Another post-Variscan structure of the study area is the
90   Ventaniella fault, which traverses the whole Cantabrian Zone in a NW – SE direction. It is essentially a dextral,
91   strike-slip fault with a net-slip of 4-5 km (Julivert et al., 1971) whose activity began in the Permian and
92   continues to the present (López-Fernández et al., 2002, 2004).

93   Outcrops of intrusive igneous rocks are common in the PCU, and their knowledge is important to
94   understand the tectonothermal evolution of this unit. Among these rocks, three granodioritic stocks (Pico Iján,
95   Peña Prieta and Pico Jano; Fig. 2) and many small outcrops can be distinguished. The latter are mainly
96   concentrated in the southern half of the unit.

97   The stocks of Pico Iján, Peña Prieta and Pico Jano dominantly have granodioritic composition and their
98   intrusion was favoured by the existence of faults (Suárez and García, 1974; Corretgé and Suárez, 1990), having
99   developed a notable aureole of contact metamorphism in the case of the Peña Prieta stock. The porphyroblasts in
100   this aurole are post-tectonic relative to the gently dipping cleavage (Gallastegui et al., 1990; Rodríguez
101   Fernández, 1994). The large number of small outcrops of igneous rocks that exist in the southern half of the unit
102   are concentrated in two areas (one in the eastern part and another in the western part) joined by a band whose
103   outcrops of igneous rocks have been related to the León fault (Corretgé et al., 1987; Suárez and Corretgé, 1987;
104   Corretgé and Suárez, 1990). All these southern outcrops have been in general related to fractures and appear as
105   small stocks, probably apophyses of bigger bodies in depth, and as sills or dikes. Their composition is varied and
106   ranges from granodioritic to gabbroic. In some cases, they developed ore bodies close to their contacts (Martín-
107   Izard et al., 1986).

108   Earlier metamorphic studies in the Cantabrian zone using CAI and/or KI methods have shown the existence
109   of areas with very low- or low-grade metamorphism in the PCU and the VU (Raven and van der Pluijm, 1986;
110   Keller and Krumm, 1993; Marín et al., 1996; Marín, 1997; Köberle et al., 1998; García-López et al., 1999, 2007,
111   2013; Bastida et al., 2002; Clauer and Weh, 2014; among others). Similar results have been obtained for the
112   southern part of the study area using coal rank and vitrinite reflectance (Colmenero and Prado, 1993; Llorens et
113   al., 2006; Colmenero et al., 2008; Clauer and Weh, 2014). This metamorphism has been described as associated

114 with a late-orogenic extensional event and to the corresponding cleavage (García-López et al., 1999, 2007, 2013;
115 Bastida et al., 2002). Analysis of ore deposits and of the tectonothermal evolution in the neighbouring unit of the
116 Picos de Europa has defined a subsequent hydrothermal episode during the Permian (Gómez-Fernández et al.,
117 1993, 2000; Bastida et al., 2004). Brime and Valín (2006) have suggested a hydrothermal origin for mineral
118 associations with chloritoid and pyrophyllite found in samples of pelitic rocks collected in the study area. K-Ar
119 dating of illite in samples collected in the southern part of the study area, identified four thermal episodes(Clauer
120 and Weh 2014), namely at: (1) 293 ±3 Ma (Cisuralian), (2) 268 ± 6 Ma (Guadalupian), (3)243 ± 5 Ma (middle
121 Triassic), and (4) 175 ± 6 Ma (early-middle Jurassic). From apatite fission tracks and zircon (U-Th)/He ages in
122 samples of Westphalian (Bashkirian-Moscovian) sandstones collected in the eastern part of the PCU, Fillon et al.
123 (2012) obtained ages of cooling below ≈180ºC of 37-39 Ma and below ≈110°C of 28-29 Ma. They inferred
124 Cenozoic erosion of a rock thickness between 6.4 and 8 km (assuming a steady-state geothermal gradient of
125 25ºC km-1).

126 **3 Methods**

127 **3.1 X-ray diffraction**

128 A total of 297 mudrocks from various localities (Fig. 2) were studied by X-ray diffraction (XRD) analysis in
129 order to determine their phyllosilicate mineralogy, Kübler Index (KI) of illites, and Árkai Index of chlorites (AI).
130 Preparation of samples and methods for XRD analysis follow the methods described in Brime et al. (2003)

131 Reaction progress in illitic minerals (*sensu* Środoń 1984) has been widely used to assess the evolution of
132 pelitic lithologies during diagenesis and low-grade metamorphism. Prograde changes can be identified by use of
133 the Kübler Index technique (illite "crystallinity", see Guggenheim et al. 2002) involving quantification of the
134 width of the 10Å peak of illite. This is an indirect measure of lattice reorganization and thickening of illite
135 crystals (Kisch 1983; Merriman and Peacor 1999) with increasing grade. The Kübler Index (KI) is expressed in
136 Δ^o 2θ to minimize variations caused by differences in recording conditions. For this study the KI was measured
137 using a laboratory procedure similar to that outlined by the IGCP 294 working group (Kisch, 1991). The
138 numerical KI value decreases with improving "crystallinity" and is expressed as small changes in the Bragg
139 angle Δ^o 2θ, using Cu Kα radiation.

140 The transient zone between diagenesis and metamorphism (the anchizone of Kübler, 1967) is defined by KI
141 values between 0.42^o and 0.25 Δ^o 2θ respectively (Kisch, 1991). The values obtained in our laboratory were
142 correlated with the Kübler scale using a calibration curve based on data obtained from polished slate standards
143 kindly provided by H. Kisch. The diagenetic zone has been subdivided in shallow (KI > 1 Δ^o 2θ) and deep (1.0 ^{o}
144 > KI > 0.42 Δ^o 2θ), using the terms proposed by the IUGS Subcommission on the Systematics of Metamorphic
145 rocks (Árkai et al., 2007). Following Merriman and Peacor (1999) we have divided the anchizone into low (0.42
146 < KI < 0.30 Δ^o 2θ) and high (0.30 < KI < 0.25 Δ^o 2θ).

147 Crystallinity index standards (CIS, Warr and Rice 1994) have also been used to compare the results thus
148 obtained with other published results using that scale. KI values obtained in this work could be converted to the
149 CIS scale by using the calibration equation:

150 $KI_{CIS} = 1.505 KI_{this\ work} - 0.046$ ($R^2 = 0.996$). $\hspace{2cm}$ (1)

151   For the problems involved in the use of the CIS scale to assess metamorphic grade see Brime (1999) and

152   Kisch et al. (2005).

153   The KI method does not allow temperature constraints to be placed on the upper and lower boundaries of

154   the anchizone and it is more likely to be a measure of reaction progress than of the thermodynamic equilibrium

155   achieved (Essene and Peacor,1995). However, this method, in combination with others such as fluid inclusions

156   or reflectance of carbonaceous material, indicates that the transition diagenesis–anchizone could be correlated

157   with a temperature of 230 ± 10 ºC whereas the limit anchizone–epizone would be at 300 ± 10°C (Müllis, 1979;

158   Frey et al., 1980; Frey, 1987; Von Gosen et al., 1991; Müllis et al., 1995; Merriman and Frey, 1999; Ferreiro

159   Mählmann et al. 2002; Müllis et al. 2002).

160   The presence of illite/smectite (I/S) and paragonite (Pg) and paragonite/muscovite (Pg/Ms) hampers

161   determination of KI of many samples. However, pro-grade changes can also be identified by use of the Árkai

162   Index technique (of chlorite "crystallinity", see Guggenheim et al., 2002) involving quantification of the width of

163   the 14Å or 7Å peaks of chlorite (Árkai, 1991, Meunier, 2005). The Árkai Index (AI) was determined in the (002)

164   peak using the same instrumental conditions as that for KI measurements. The anchizone limits have been

165   established using samples free of I/S, Pg-Pg/Ms in which the KI measured in the air dried samples has been

166   calibrated with Kisch standards. The AI was measured in the same samples and the regression line obtained

167   allowed the delimitation of the upper and lower anchizone limits using chlorite for AI values of 0.234 and 0.135

168   Δº2θ respectively. It should be noted that AI values are smaller than the corresponding KI and the method is

169   therefore slightly less sensitive than KI method.

3.2 Conodont colour alteration index (CAI).

171   Colour changes in conodont elements are related to the progressive and irreversible alteration of the amounts of

172   organic matter within their apatite composition. The CAI method is based on analysis of the colour changes that

173   the conodonts undergo in response of the organic matter to a temperature increase with time. These changes

174   permit construction of a scale of CAI values with eight units that allows the use of the conodonts as maximum

175   paleothermometers for a temperature interval of between 50 and 600ºC (Epstein et al., 1977; Rejebian et al.,

176   1987). It is used mainly in carbonate rocks. Besides colour changes, apatite textural alteration also takes place

177   and can provide complementary information about the thermal conditions. In the present paper the terminology

178   of Rejebian et al. (1987) and García-López et al. (1997, 2006) is used for the textural description of conodonts.

179   In agreement with Rejebian et al. (1987), well preserved conodonts and high CAI values with a wide dispersion

180   are indicative of contact metamorphism. Furthermore, coarse recrystallization and corrosion are related to

181   hydrothermal processes.

182   Samples were collected from the Pisuerga-Carrión and Valsurbio units. CAI values are based on 5 kg

183   samples of limestone that were treated, with 6% acetic acid solution. Unfortunately, recovery of conodonts from

184   Carboniferous rocks was hindered in some areas by their dominantly siliciclastic character. Sampling was

185   complemented with specimens from collections housed at the University of Oviedo (Spain) and those of the

186   National Museum of Natural History at Leiden (Netherlands) and Institut und Museum für Geologie und

187   Paläontologie in Göttingen (Germany) (Fig. 2). 213 positive samples, corresponding to an age interval from the

188   Pridoli to Ghzelian, were analyzed for CAI determination (Appendix 1 in the supplementary material).

189   The methodology involved in CAI determination can be found in García-López et al. (1997) and Bastida et
190   al. (1999). Several CAI values were obtained from most samples and the mean of CAI values have been
191   determined for each of them in order to tentatively contour CAI values. Samples with a range higher than 1.5
192   have not been used to obtain mean CAI values and temperatures. The interpretation of the results is mainly based
193   on the analysis of the CAI isograds and their relationship to the stratigraphic contacts and the main structures of
194   the study area.

195   For the metamorphic zonation from CAI data, we use the terminology described by García-López et al.
196   (2001), that involves a division in diacaizone (CAI < 4), ancaizone ($4 \leq$ CAI ≤ 5.5) and epicaizone (CAI > 5.5).

197   Temperature ranges of the CAI values were obtained from the Arrhenius plot presented by Epstein et al.
198   (1977) and Rejebian et al. (1987). The maximum heating time is the age of the rock. Nevertheless, it is possible
199   to place greater limits on this maximum time (García-López et al., 2013). According to the Arrhenius plot a
200   minimum temperature is required to obtain a specific CAI value. For example, in a rock with an age of 400 Ma
201   (Devonian), development of CAI = 5 requires at least 290ºC (point *A* in Fig. 3). Then, it can be assumed that a
202   temperature <290ºC does not contribute to generate a CAI = 5. Hence, to produce a given CAI, the maximum
203   time of heating begins when the rock reaches the minimum temperature necessary to produce that CAI, and it
204   ends when the rock cools down and the temperature becomes lower than this minimum value. According to
205   García-López et al. (2013), and in agreement with the age of the metamorphism analyzed below, we consider
206   that the main heating time corresponds to a late-Variscan period that began at the boundary Kasimovian-
207   Gzhelian and ended at the beginning of the Triassic (heating time of about 50 Ma). The hydrothermal post-
208   Variscan episodes described by Clauer and Weh (2015) and the most recent heating (mainly Cenozoic) described
209   by Fillon et al. (2012) probably lasted less than 50 Ma. If there are rocks within the area that underwent more
210   than one heating period, the period to be considered for the development of the CAI is the one which generated
211   higher temperatures. Anyway, due to the geometry of the CAI curves in the Arrehnius plot, for heating intervals
212   such as those involved in the present case, an error of a few Ma in the maximum time of heating has little
213   influence on the results. According to Patrick et al. (1985) and Rejebian et al.(1987) the minimum time of
214   heating assumed here is of 1 Ma.

3.3 Cleavage

216   Development of cleavage requires ~~ductile deformation of rocks and~~ mineralogical and microstructural changes,
217   involving mechanisms, such as pressure solution, which require a minimum temperature of about 200ºC for
218   cleavage development in pelitic rocks and 175ºC in limestones (Groshong et al. 1984). Thus widespread
219   presence of cleavage occurs below a certain crustal level (minimum overburden of 5-7 km; Engelder and
220   Marshak, 1985). Furthermore, the relations between folds and cleavages and the overprinting relations between
221   cleavages, play an important role in defining deformation events (Passchier and Trouw, 2005). In addition,
222   cleavage is also a key structure to establish chronological relations between metamorphic crystallization and
223   deformation. In the context of the study area, we call tectonothermal event to a deformational event with
224   cleavage development and associated metamorphic conditions, and use the term thermal event for metamorphic
225   conditions without cleavage development. Hence, cleavage is considered here as the main reference to ascertain
226   the tectonothermal events of the study area.

227 4 Results and interpretation

228   In order to facilitate description, the samples have been grouped in the following areas, namely Liébana,

229 Valdeón, Yuso-Carrión, Pisuerga, Riaño-Cervera and Valsurbio (Fig. 4; Martín-Merino et al., 2014).

230 4.1 Clay minerals

231 4.1.1 Clay mineral assemblages

232 Mineralogical analysis of the <2 μm fractions shows that dioctahedral K- rich mica-like structures (referred to as

233 illite or muscovite, I-Ms) is present in all the samples with the majority containing also chlorite (Chl) (Fig. 5).

234 Chlorites have poorly developed 14 Å peaks, indicating high iron content suggestive of chamositic compositions

235 (Moore and Reynolds, 1997). Other phases such as ordered illite/smectite (I/S) or paragonite (Pg) and mixed

236 layers paragonite/muscovite (Pg/Ms) are also common. Asymmetry of peaks and sample behaviour after

237 glycolation indicate the presence of ordered illite/smectite (I/S). Absence of random I/S indicate that zone 3 of

238 Eberl (1993) has been reached, suggesting that temperatures exceeded 100ºC. Kaolinite (Kln) and pyrophyllite

239 (Prl) may also be present in some samples and are abundant in a few samples. In addition, chloritoid (Cld) is also

240 present and widespread in the study area. However it is more abundant in samples close to the intrusions and/or

241 faults in which case is found together with chlorite and Pg + Pg/Ms. Only in samples to the E (Pisuerga Area),

242 where it is not as abundant, it may be associated with Prl, Kln or I/S. Finally chlorite/vermiculite mixed

243 layer(C/V) and stilpnomelane (Stp) have been found, in small amounts, in a few samples, and are restricted to

244 samples to the E Pisuerga and E Yuso-Carrión areas.

245   The I/S is more abundant to the N and E (Valdeón, Liebana, and eastern part of the Pisuerga and Yuso-

246 Carrión areas) and Kln presence is almost restricted to samples to the E (Pisuerga and eastern part of the Yuso-

247 Carrión areas). Prl is common in samples from the central and northeastern parts (Liebana and Pisuerga areas,

248 and eastern part of the Yuso Carrión area) (Fig. 5). Quartz, calcite, feldspars and goethite were accessory phases

249 recognized in some samples.

250   In general the assemblages found are, in order of abundance of the most frequent phase besides illite (n =

251 291):

252 I + **Chl** (n = 215) ± Pg ± Pg/Ms ± Cld ± I/S ± [C/V] ± [Prl] ± [Kln] ± [Stp]

253 I + **Pg** + **Pg/Ms** (n = 162) ± Chl ± Cld ± [I/S] ± [Prl] ± [C/V] ± [Kln] ± [Stp]

254 I + **I/S** (n = 112) ± Chl ± Kln ± Prl ± C/V ± Pg ± Pg/Ms ± C/V± [Cld] ± [Stp]

255 I + **Cld** (n = 93) ± Chl ± Pg ± Pg/Ms ± I/S ± [C/V] ± [Prl] ± [Kln] ± [Stp]

256 I + **Prl** (n = 56) ± I/S ± Chl ± Pg ± Pg/Ms± C/V± [Cld] ± [Kln] ± [Stp]

257 I + **C/V** (n = 48) ± I/S ± Chl ± Pg ± Pg/Ms ± Kln ± Prl ± Cld ± [Stp]

258 I + **Kln** (n = 39) ± I/S ± Chl ± C/V± [Prl] ± [Cld] ± [Pg] ± [Pg/Ms] ± [Stp]

259 4.1.2 Kübler Index

260 As mentioned above, determination of KI has been hampered by the presence in some samples of certain types

261 of I/S and big amounts of Pg or Prl, in relation to the amount of illite, that interfere with the 001 peak of illite,

262 therefore rendering their KI values useless for grade determination using KI, even in the glycolated state. As a

263 result, 23 (indicated in *italics* in Appendix 2 in the supplementary material) of the 291 samples studied yielded

264 doubtful KI values (Figs. 6, 7; Appendix 2 in the supplementary material). They are ~~given,~~ nevertheless, as they

265 indicate maximum value of the KI.

266 Grade ranges from deep diagenetic to epizonal, but deep diagenetic and mainly low anchizonal metapelites

267 are predominant in most of the areas (Figs. 6 and 7). Expandability of the 10 Å peak is only lost at the high

268 anchizone to epizone boundary. Deep diagenetic areas can be found to the north (Liébana and Valdeón areas).

269 The Riaño-Cervera area is mainly low anchizonal with a few samples being diagenetic or deep anchizonal. In the

270 Pisuerga area, the grade ranges from deep diagenetic to low anchizonal (Figs. 6, 7). Higher grade (epizonal)

271 samples may appear in any formation and they are more abundant in the western part of the Yuso-Carrión area,

272 where a big size intrusion is located, and in the Devonian of the Valsurbio area (Fig.7). In both cases, it is in

273 those high grade samples where chloritoid is more abundant (Fig.6).

274 Work in progress on the variation of the chemical composition of the phyllosilicates of the study area

275 allowed estimation of temperatures using Battaglia's (2004) approach based in the variation in the chemical

276 composition of illites. Temperatures obtained are in the range 230-280°C, consistent with the anchizonal KI

277 values of the analyzed samples. The observed deficit in layer charge (Brime and Valín, 2006) is characteristic of

278 anchizonal K white micas (Hunziker et al 1986; Livi et al. 1997; Merriman and Peacor 1999; Árkai 2002; Árkai

279 et al 2003).

4.1.3 Árkai Index

281 It, has been measured in 118 samples, most if not all of them, containing various amounts of I/S and/or Pg,

282 Pg/Ms. Of them, 40 yielded diagenetic values, 63 low anchizonal values, 14 high anchizonal values and just one

283 epizonal value. Distribution of these values can be seen in Fig. 6. In those cases in which KI and AI have been

284 determined, the correlation of the grade indicated by both ~~indices,~~ is good ($r = 0.65$; significance level 0.1% $r_{60} =$

285 0.41) suggesting that both phases were formed under the same conditions, and supporting the reliability of the AI

286 as indicator of grade in those cases in which it is the only index available (Árkai et al., 1995).

287 The existence of some discrepancies between KI and AI may be caused by the presence of small amounts

288 of I/S or Pg that alters the width of the illite peaks. However, those discrepancies are always small and are

289 usually in samples at the boundary between metamorphic grade zones (Appendix 2 in the supplementary

290 material).

4.1.4 Mineral distribution in relation with grade

292 Kaolinite is present in deep diagenetic samples and also in some low anchizonal ones. Maximum stability of Kln

293 is 270°C, according to laboratory experiments (Velde, 1992) and therefore in agreement with its presence in the

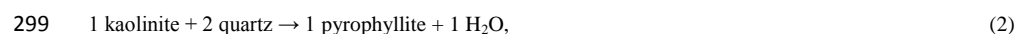
294 anchizonal samples. Paragonite, apart from its presence in deep diagenetic samples, is more frequently present in

295 the anchizone and some epizonal samples (Figs. 5 and 6). Pyrophyllite is more abundant in samples from the

296 anchizone but it can also be present in diagenetic samples.

297 The widespread occurrence of Kln and quartz in the diagenetic rock samples may provide the starting

298 material for the formation of Prl by the reaction

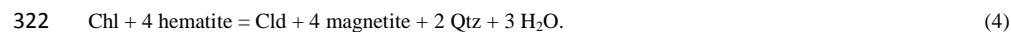
299 1 kaolinite + 2 quartz → 1 pyrophyllite + 1 H_2O,   (2)

300 as suggested in the Glarus Alps by Frey (1978), who considered Prl an indicator of anchizonal regional
301 conditions. In fact of the 53 samples in which Prl is present, Kln is found, and in very small amounts, in only 8
302 of them. However, the stability field of Prl is strongly influenced by water activity, and thus the formation
303 temperature could be notably lower (Thompson, 1970; Winkler, 1979; Hemley et al.,1980). Its presence in
304 diagenetic samples is not uncommon and could due to the influence of magmatic fluids (Hosterman et al., 1970;
305 Kisch, 1987). According to Kisch (1987) Prl, appears in regional terrains only in the anchizone but in areas of
306 intrusive activity it may appear in lower grade zones. Therefore presence of Prl in samples with diagenetic KI/AI
307 values, as in the eastern part of the Liebana and Yuso Carrión areas, could be regarded as evidence for high
308 geothermal gradients or magmatic heating.

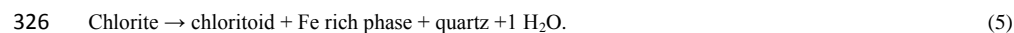
309 Chloritoid is abundant in samples from the high anchizone to epizone (southeastern Valsurbio area and
310 western Yuso Carrión area), but it can also be present, in smaller amounts, in low anchizonal (1, 3W, 4E, 4W,
311 5E, 5W) and even diagenetic samples (eastern Liébana, Pisuerga and Riaño-Cervera areas). Cld is Fe rich. The
312 average Mg/(Fe+Mg) found is < 0.12 (Brime and Valín, 2006) similar to that of pelites subjected to intermediate
313 P/T conditions. It is noteworthy that when this phase is present in the samples (a total of 93 samples have Cld),
314 Prl is absent, or in very minor amounts in a few samples (ten in total), indicating that it could have been formed
315 according to the reaction originally proposed by Zen (1960), which is generally accepted for the formation of
316 Cld during metamorphism of aluminous pelites (Theye et al. 1992):

317 Pyrophyllite + chlorite → chloritoid + quartz +1 H_2O. (3)

318 The absence of Cld in the eastern part of the Yuso Carrión area where Prl is abundant, together with Chl,
319 could indicate that the temperature required for its formation by reaction (3) has not been reached.

320 Presence of some Fe oxides has been detected in samples from the study area. Therefore more Cld could be
321 produced by the reaction suggested by Bucher and Frey (1994):

322 Chl + 4 hematite = Cld + 4 magnetite + 2 Qtz + 3 H_2O. (4)

323 However, it has been observed in thin sections that occurrence of Prl is almost restricted to veins and
324 fracture zones, suggesting that Cld could have been formed during the hydrothermal alteration of the pelites
325 following the reaction proposed by Phillips (1988):

326 Chlorite → chloritoid + Fe rich phase + quartz +1 H_2O. (5)

327 Presence of Cld in the anchizone has been discussed by Kisch (1983), who concluded that Cld cannot
328 unequivocally be regarded as an indicator of the beginning of the epizone, as previously suggested, because there
329 are occurrences in the anchizone (Árkai et al., 1981). In the study area, Cld is more abundant in the epizonal
330 samples of the Valsurbio (6) and W Yuso-Carrion (3W) areas, but it is also present in low anchizonal and a few
331 deep diagenetic samples from the Riaño Cervera (5E), Pisuerga (4E) or Valdeón (2) areas, thus corroborating the
332 conclusion of Kisch (1983).

333 Chloritoid and Prl are widespread in virtually all rock types and grade conditions (Appendix 2 in the
334 supplementary material). This occurrence could be related to basin wide alteration by infiltrating hydrothermal
335 fluids (Phillips, 1988; Brime and Valín, 2006). Late to post Variscan fluid flow events have been described in the
336 other areas of the Cantabrian Zone (Ayllón et al., 2003; Gasparrini et al., 2003) and related to a more general
337 Variscan event (Boni et al., 2000).

4.2 Conodont colour alteration index (CAI).

The CAI values are independent of the stratigraphic position of the samples (Fig. 8). These values vary widely, ranging from 1.5 to 7.5, corresponding respectively to intervals of temperatures of < 40-60ºC and 550-590ºC (see Appendix 1 in the supplementary material). However, values equal to or lower than 2 are unusual, being limited to the southeastern sector of the Pisuerga-Carrión unit. Some samples with conodonts having high CAI values and a range of one and a half units, or more, are indicative of contact metamorphism and/or hydrothermal processes. Values ≥ 6 are usually found close to outcrops of igneous rocks. The upper boundary of the ancaizone (CAI = 4) corresponds to a temperature range of 190 - 225°C, while the lower limit (CAI = 5.5) corresponds to the range of 340-375ºC.

The lack of carbonate rocks prevents in some areas the construction of a complete map of CAI isograds; however, it is possible to observe that they crosscut the trend of the Variscan structures. CAI data allow the distinguishing of the following sectors in the study area (Fig. 8):

(a) Northern sector (Liebana and Valdeón areas).This is an ancaizonal area that passes without thermal discontinuity through the basal thrust of the Picos de Europa unit. Inside this unit, the boundary ancaizone/diacaizone appears and the CAI decreases northwards (Bastida et al., 2004; Blanco-Ferrera et al., 2011).

(b) Central-eastern sector (eastern part of the Yuso-Carrión area) with dominance of ancaizonal conditions. Here, in the areas where CAI data exist, a remarkable homogeneity of CAI values appears, mainly in the Devonian rocks of the area located to the east of the Curavacas-Lechada syncline.

(c) Central and western sectors, (western part of the Yuso-Carrión and Riaño-Cervera areas) where the limited CAI data available indicate that epicaizonal areas coexist with ancaizonal areas.

(d) Southern sector. It corresponds to the VU and presents a wide area with epicaizonal conditions (García-López et al., 2013).

(e) Southeastern sector (eastern part of the Yuso-Carrión and Pisuerga areas). In this area, diacaizonal conditions are dominant, but ancaizonal and epicaizonal areas also appear. The latter areas appear adjacent to outcrops of igneous rocks. The greater variation of CAI values is found in this area, with a range from 1.5 to 7.

The apparently chaotic distribution of CAI isograds is probably due to a heat from subsurface intrusions at depth resulting in isotherms having complex geometry. This pattern may be related to a crustal thinning during an extensional episode and to the subsequent intrusion of igneous bodies that emplaced at different levels, generating contact metamorphism and hydrothermal fluids.

In general, conodonts of the study area are undeformed and well preserved. Most of them have granular texture due to apatite recrystallization under high temperature. The apatite crystals do not show preferential orientation and their size increases with the temperature. The granular texture is commonly incipient for CAI values between 4 and 4.5 and is widespread for CAI values ≥ 5. Furthermore some conodonts with CAI values between 6 and 7 have coarse recrystallization, corrosion and loss of ornamentation, so that in a few cases they have lost their original morphology ("ghost conodonts"). These alterations of conodonts with CAI values ≥ 5 are indicative of contact metamorphism and hydrothermal processes. Some conodonts with CAI ≥ 4.5 present sets of parallel microfissures, probably related to the rock cleavage. Conodonts with CAI ≤ 4 have occasionally unaltered surfaces, but most of them present a sugary texture (dull, frosted or pitted surfaces).The surfaces of

378 these conodonts show several types of overgrowth of apatite crystals, mainly developed independently of the

379 CAI value (or thermal changes under diacaizonal conditions); they were usually a result of apatite solution and

380 crystallization processes (Blanco-Ferrera et al., 2011).

4.3 Cleavage development and tectonothermal events

382 Two main cleavages have been found in the study area (Fig. 9). Cleavage S_1 is a rough foliation associated with

383 upright folds and trends approximately E-W; it appears mainly in the northern half of the study area. This

384 cleavage affects latest Carboniferous rocks (Kasimovian-Gzhelian age). Cleavage S_2 dips gently, crosscuts

385 earlier upright folds and is associated with the development of meter-scale open cascade folds. It is especially

386 well developed in the Curavacas-Lechada syncline and in the VU. Both cleavages appear in different areas and

387 have not been observed superimposed. The existence of two cleavages with different age suggests that two

388 tectonothermal events took place in the southeastern sector of the CZ.

389 The first event, associated with the S_1 cleavage (sub-vertical), is not well defined, since it developed in an

390 area where presence of I/S and Kln and KI and AI values indicate deep diagenesis or low anchizonal conditions,

391 and CAI values dominantly show ancaizonal conditions. According to the age of the latest rocks affected, this

392 event probably occurred during the late Gzhelian. It resulted in crustal thickening produced by the N - S

393 shortening that gave rise to the emplacement of the Picos de Europa unit, the last major Variscan thrust unit that

394 was generated in the foreland fold and thrust belt, and it disassociated with the development of other thrusts and

395 folds to the south of this unit.

396 The second event produced cleavage S_2 that crosscuts the upright folds in the southern part of the study

397 area (Curavacas-Lechada syncline and the VU). Under the microscope, this cleavage shows evidences of

398 pressure solution and crystallization of oriented muscovite and chlorite with formation of chlorite-muscovite

399 porphyroblasts. X-ray diffraction indicates that the rocks affected by this cleavage present Chl + Pg-Pg/Ms

400 assemblages; KI and AI values indicate high anchizonal to epizonal conditions and CAI values indicate similar

401 conditions. The gently dipping attitude of S_2 suggests that this event was associated with an extensional

402 deformation and the corresponding crustal thinning as a consequence of a late orogenic gravitational

403 readjustment of the orogen in the core of the Ibero-Armorican arc. This interpretation agrees with the kinematic

404 evolution of the arc proposed by Gutiérrez-Alonso et al. (2004, 2011). The event culminated with the intrusion

405 of igneous rocks that ~~traversed~~ the rocks with cleavage and generated a contact metamorphism associated with

406 hydrothermal processes, as indicated by: (1) the widespread presence of Cld and Prl (Fig 5; Brime and Valín,

407 2006) and low KI and AI values, (2) recrystallization in conodonts (granular texture), (3) high CAI values in

408 samples close to outcrops of intrusive rocks, (4) wide range of CAI in some samples, and (5) irregular variation

409 of the CAI through the area and strong corrosion in some conodonts. The cleavage S_2 developed earlier than the

410 porphyroblasts (biotite, andalusite and chloritoid) formed during the contact metamorphism associated with the

411 granodioritic stock of Peña Prieta (Gallastegui et al., 1990; Rodríguez Fernández, 1994), whose age is Cisuralian

412 (292+2/-3 Ma after Valverde-Vaquero et al., 1999). These data suggest the development of a thermal event that

413 took place near the boundary Carboniferous-Permian and that progressed during the Cisuralian with the intrusion

414 of many small igneous bodies that rose along faults (Suárez and García, 1974; Corretgé and Suárez, 1990).

415 In some locations, S_2 is gently folded with local development of crenulation cleavage. This may be a result

416 of the Alpine deformation, which is the only post-Variscan compressional deformation described in the area

417 (Gallastegui, 2000 and references therein), and involves a ductile deformation that required a moderate
418 temperature and gave rise to a dome shape in the Valsurbio unit (Marín, 1997).

419   The features described above refer to late-Variscan events related to the development of penetrative
420 structures or to the intrusion of igneous rocks. However, the existence of such events does not preclude the
421 development of others subsequently, such as the hydrothermal post-Variscan episodes described by Boni et al.
422 (2000), Muchez et al. (2005), Gasparrini et al. (2006) and Clauer and Weh (2014).

423   5 Discussion

424 The distribution of the different grade indicators used in the study area shows, in general, an acceptable
425 correlation between them, although some discrepancies have been observed. All methods coincide in pointing
426 out the location of the areas with higher metamorphic grade, being the CAI and the AI the indicators that tend to
427 give the highest and the lowest grade respectively. All indices point to an irregular distribution of the areas with
428 very low-grade metamorphism. Diacaizonal areas are limited to the southeastern sector. A discrepancy is
429 observed in the eastern Yuso-Carrión area between clay mineral and CAI data. A notable number of CAI values
430 systematically indicate ancaizonal conditions, whereas clay assemblages with abundant I/S and Kln, KI and AI
431 indicate diagenetic conditions.

432   The correlation among the different indicators that can be used to establish the metamorhic grade is
433 challenging due to several factors, such as the different kinetics of the processes that modify the colour of the
434 conodonts and the transformation of the clay minerals and the different influence of fluids in limestones and
435 pelitic rocks. These processes could explain the discrepancies observed.

436   In the context of the Cantabrian Zone, the low-grade extensional metamorphism of the PCU extends
437 westwards and allows an elongated area to be defined that can be followed up to the Central Coal Basin (Fig. 1)
438 (Aller, 1981, 1986; Brime, 1985; Aller et al., 1987, 2005; García-López et al., 2007). The biggest width of this
439 zone is in the study area, coinciding with the core of the Ibero-Armorican arc. In the Central Coal Basin, this
440 metamorphism is linked to the occurrence of cleavage associated with crosscutting folds and it has been related
441 to the possible existence of igneous bodies in depth (Aller, 1986) or to the rise of fluids along faults, especially
442 the León fault (Aller et al., 2005). In the metamorphic southern part of the study area (VU), the metamorphism
443 disappears westward, so that the adjacent western unit of the Esla nappe region is not metamorphic (García-
444 López et al., 2013; Valín and Brime unpublished data).

445   The existence of hydrothermal alteration has been suggested by Brime and Valín (2006) and Clauer and
446 Weh (2014). The common occurrence of Cld and Prl and the irregular distribution of CAI values, the wide range
447 in the high CAI values in some samples and textural alterations of conodonts agree with the occurrence of
448 hydrothermal fluids and possible subsurface igneous bodies. As for the thermal events dated by Clauer and Weh
449 (illite K-Ar; 2014), the first has an age(293 ±3 Ma),comparable to that of the Peña Prieta granitoid (292+2/-3 Ma
450 after Valverde-Vaquero et al., 1999, Cisuralian), and the others (Guadalupian, middle Triassic and early-middle
451 Jurassic) are probably related to the crustal extension associated with the Basque-Cantabrian basin. Specific
452 structural evidences of these three later events have not been found and their temperatures were probably lower
453 than those of the episode linked to the late-Variscan extensional episode, which, being related to igneous
454 intrusions, probably gave rise to the paleothermal peak.

455 The zircon (U-Th)/He ages obtained by Fillon et al. (2012) indicate that the Westphalian (Bashkirian-
456 Moscovian) rocks involved in the dating had probably a temperature above 180ºC up to 37-39 Ma ago (Eocene).
457 This is consistent with the occurrence of Alpine ductile deformation, which resulted in a local crenulation
458 cleavage.

459 **6 Conclusions**

460 Study of the low-grade metamorphic rocks and associated structures in the south-eastern sector of the CZ has
461 allowed a model for the tectonothermal evolution of the core of an arcuate orogenic belt to be developed. This
462 core, although it is a part of a foreland fold and thrust belt where diagenetic conditions are dominant, portrays a
463 complex evolution due to its special location inside the belt. It was thrust during the Carboniferous by large units
464 from south, west and north, which resulted in a great accumulation of syntectonic sediments, and development of
465 unconformities and structures. The latter arose as a result of compression in different directions that also
466 generated a crustal thickening. The emplacement of the last thrust unit (Picos de Europa unit) produced a N – S
467 shortening that generated thrusts and associated folds. Shortly afterwards, upright folds with E – W trend and
468 associated cleavage (S_1) developed, mainly in the northern part of the sector. The ductile deformation occurred
469 under thermal conditions which reached the anchizone and the ancaizone in some areas. This represents the first
470 tectonothermal event registered in the southeastern sector of the CZ.

471 At the end of the Variscan deformation, gravitational instability gave rise to an extensional episode and the
472 corresponding crustal thinning. During this event an increase in the thermal gradient enabled ductile deformation
473 and low-grade metamorphism to take place in some areas (second tectonothermal event), mainly in the central
474 part (Curavacas-Lechada syncline) and the southern part (Valsurbio unit) of the sector. The ductile deformation
475 produced gently dipping cleavage (S_2) crosscutting earlier upright F1 folds. Some small open cascade folds were
476 also produced. The metamorphism reached epizonal and epicaizonal conditions during this event. The process
477 culminated in the Cisuralian with the intrusion of igneous rocks and the development of contact metamorphism
478 around the larger igneous bodies. In the case of Peña Prieta granodiorite, post S_1 andalusite porphyroblasts
479 developed. Hydrothermal fluids were common during this extensional episode, resulting in the development of
480 Prl and Cld, and textural alteration and high CAI dispersions in conodont samples.

481 As a whole, the metamorphic grade is independent of the stratigraphic location of the samples and the trend
482 of the main structures, indicating the late-orogenic character of the tectonothermal events. The thermal level
483 decreases progressively northwards inside the adjacent Picos de Europa unit. On the other hand, the
484 metamorphism associated with the extensional episode in the PCU is extended westwards as an elongated area
485 whose development was probably favoured by the rise of fluids along faults, especially along the León fault.

486 With the development of the adjacent Basque-Cantabrian basin, the extensional regime extended during the
487 Mesozoic, with the probable occurrence of hydrothermal processes. The temperature of the rocks was probably
488 important during Cenozoic times (> 180ºC until the late-Eocene after Fillon et al., 2012, in the eastern part of the
489 study area), so that the Alpine deformation generated locally some ductile deformation with crenulation cleavage
490 affecting the S_2 cleavage.

491 The evolution described above is summarized in Table 1.

492

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Variscan deformation	Emplacement of north-directed Palentine nappes (prior or earliest Moscovian) and the adjacent western Cantabrian nappes (late Moscovian) with associated thrusts in the PCU
Variscan deformation (cont.) (N-S shortening)	Emplacement of the south-directed Picos de Europa unit; thrusts and folds in the PCU (Kasimovian-Gzhelian)
	Folds and axial plane cleavage (S_1); first tectonothermal event, with deep diagenetic – low anchizonal and ancaizonal areas in the northern part of the PCU (late Gzhelian)
Late Variscan gravitational readjustment; extensional event	Gently dipping cleavage (S_2) associated with crosscutting folds; second tectonothermal event; very low- or low-grade metamorphism (high anchizone-epizone, and ancaizone-epicaizone) in the syncline of Curavacas-Lechada and the VU. Normal faults (latemost Gzhelian to early Cisuralian)
	Intrusion of igneous rocks, contact metamorphism and wide development of hydrothermal processes (Cisuralian)
Extension linked to the Basque-Cantabrian basin	Permian and Mesozoic hydrothermal episodes
Alpine deformation	N-S shortening, tightening of gentle folds, local crenulation cleavage and tilting of rocks northwards (Cenozoic).

826 **Table 1. Tectonothermal evolution of the southeastern sector of the Cantabrian Zone.**

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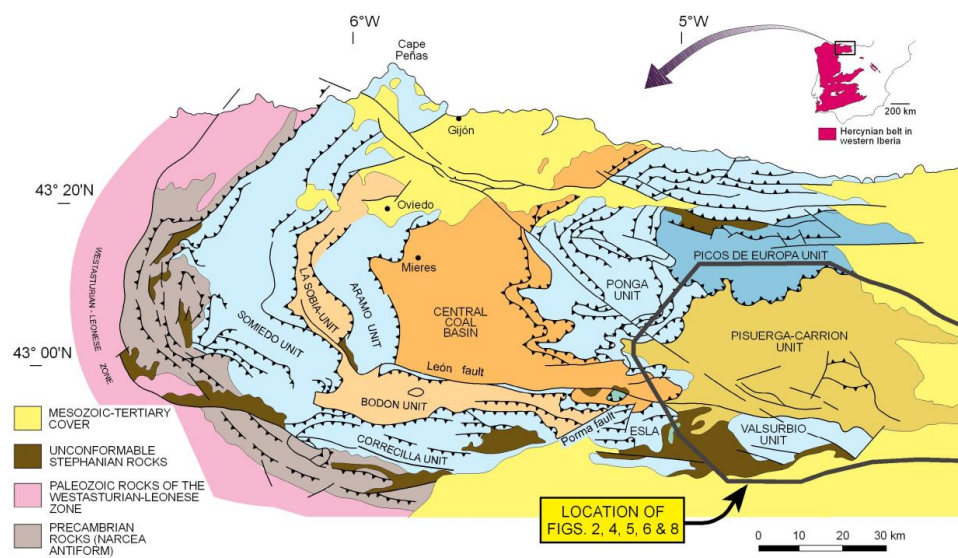
Figure 1. Generalized geological map of the Cantabrian Zone (after Julivert 1971) showing major thrust

units and the location of the study area.

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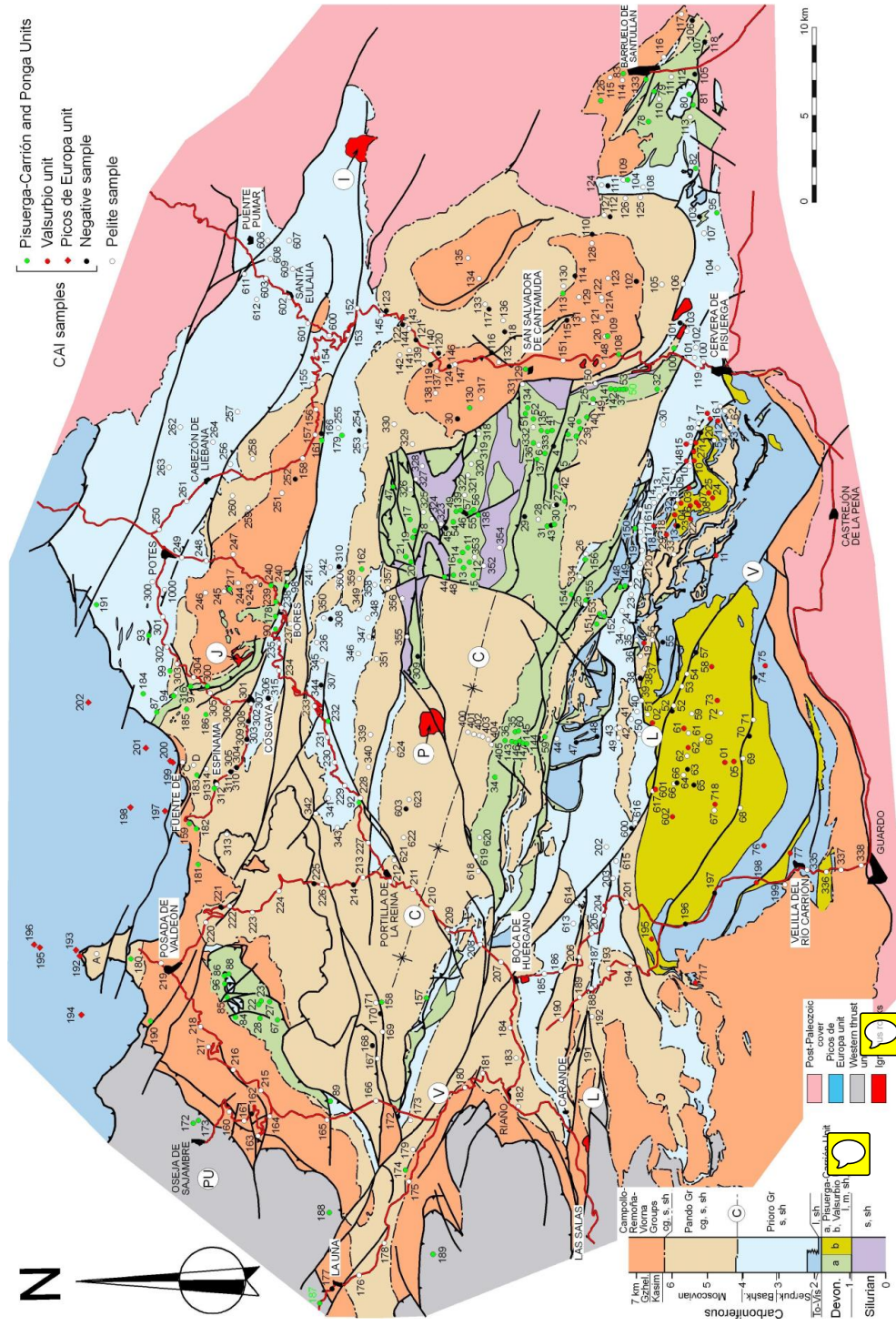
Figure 2. Geological map of the southeastern part of the Cantabrian Zone showing sampled localities (composed from Lobato, 1977; Colmenero et al., 1982; Ambrose et al., 1984; Julivert and Navarro, 1984; Martínez García et al., 1984; Lobato et al., 1985; Heredia et al, 1986, 1991, 1997; Rodríguez Fernández, 1994; Rodríguez Fernández et al., 1994). Devonian and Silurian rocks of the PCU form the Palentine nappes. cg, conglomerate; l, limestone; m, marl; s, sandstone; sh, shale. C, Curavacas-Lechada syncline; I, Pico Iján granodiorite; J. Pico Jano granodiorite; L, León fault; P, Peña Prieta granodiorite; PU, Ponga unit; V, Ventaniella fault. Picos de Europa conodont samples after Bastida et al., 2004; Valsurbio conodont samples after García-López et al., 2013.
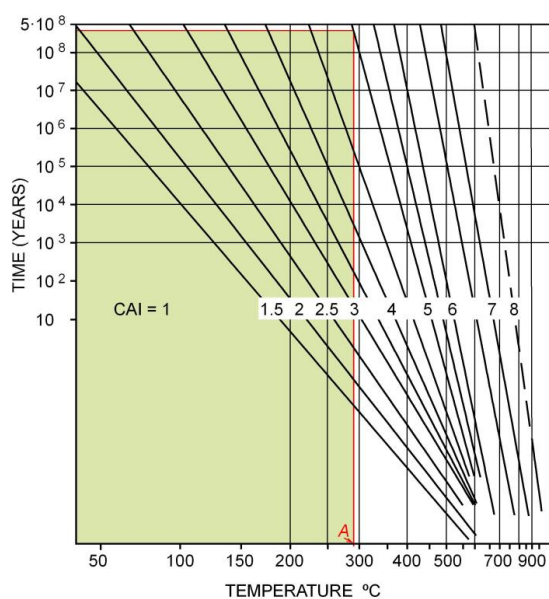
Figure 3. Arrhenius plot to determine paleotemperature from CAI values (on the lines) and heating time (from Rejebian et al., 1987). Point *A* indicates the minimum temperature necessary to obtain a CAI = 5 for a rock with an age of 400 Ma. The green area shows the temperatures and heating times unable to produce CAI ≥ 5 in this rock.

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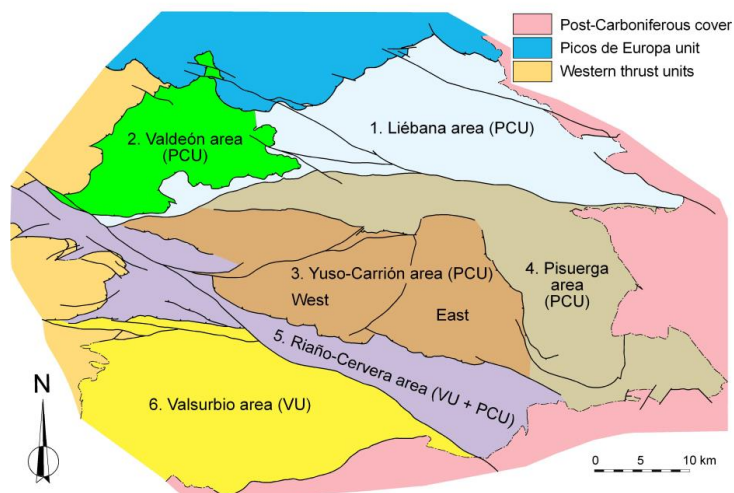
Figure 4. Diagramatic map indicating the areas in which the study units have been subdivided, following

Martín-Merino et al. (2014)

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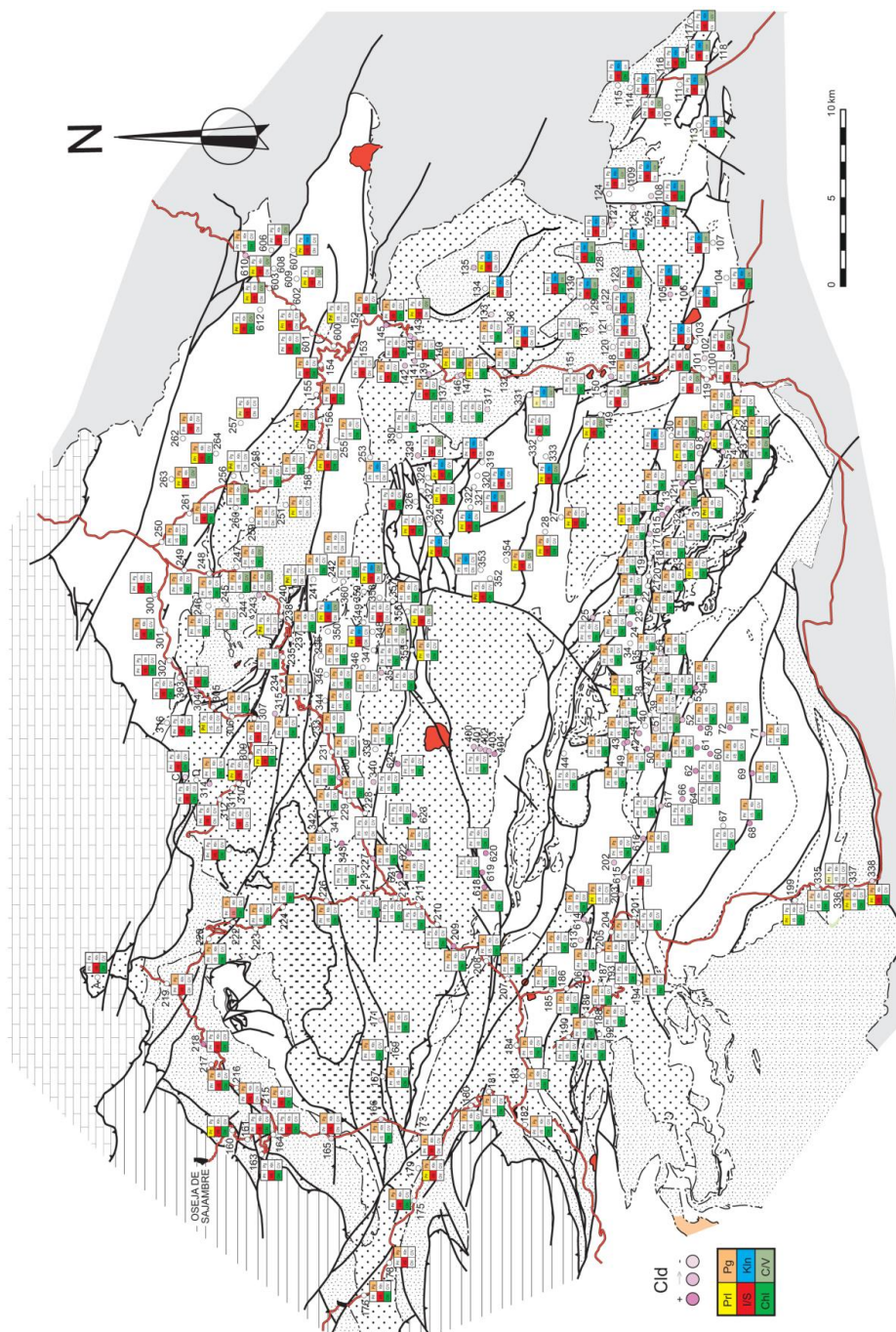
Figure 5. Distribution of clay minerals in the study area. All the samples contain illite and therefore this phase has not been considered in the plot. Presence of a certain phase in the sample is indicated by the colour in the corresponding square. For legend see Figure 8.

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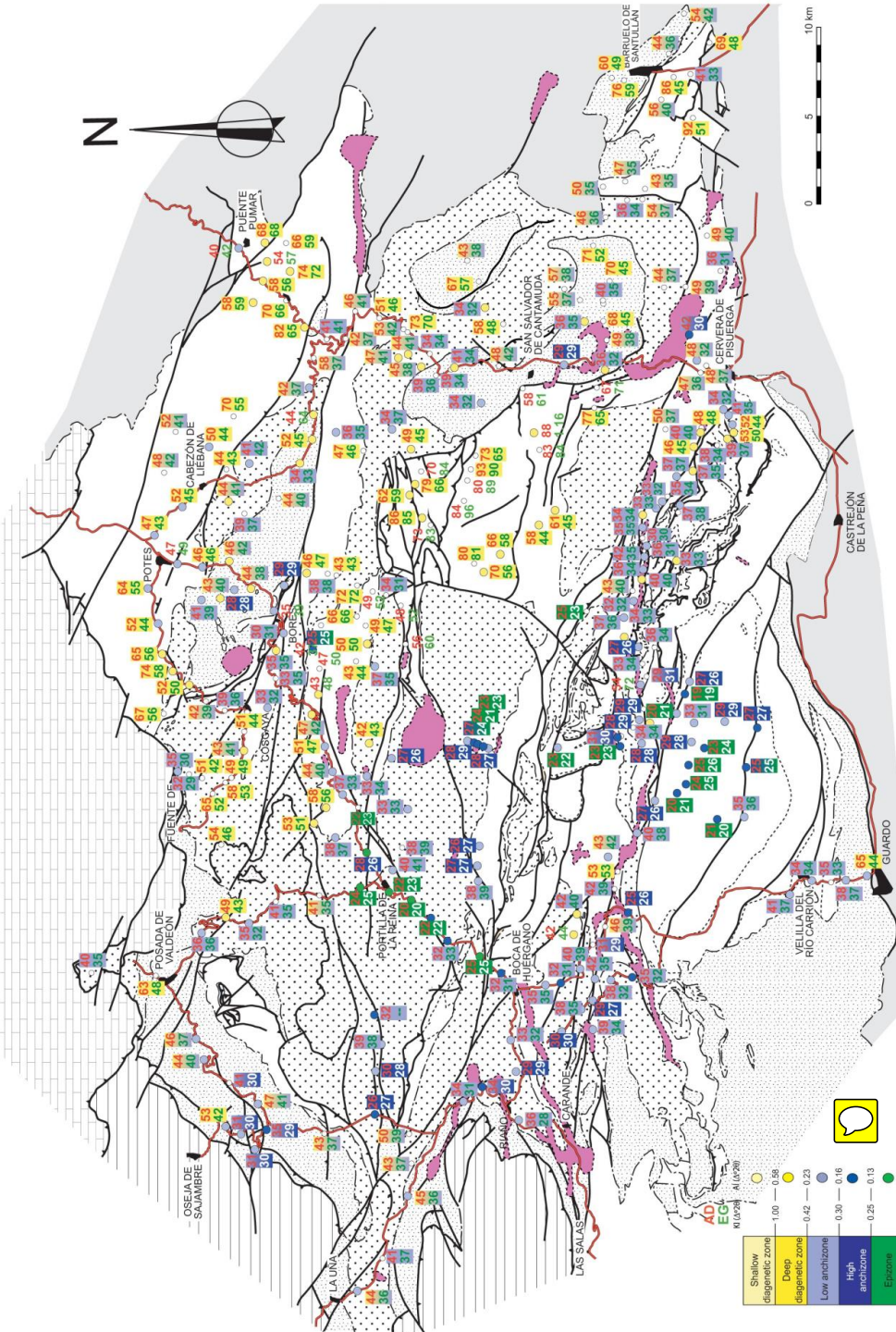
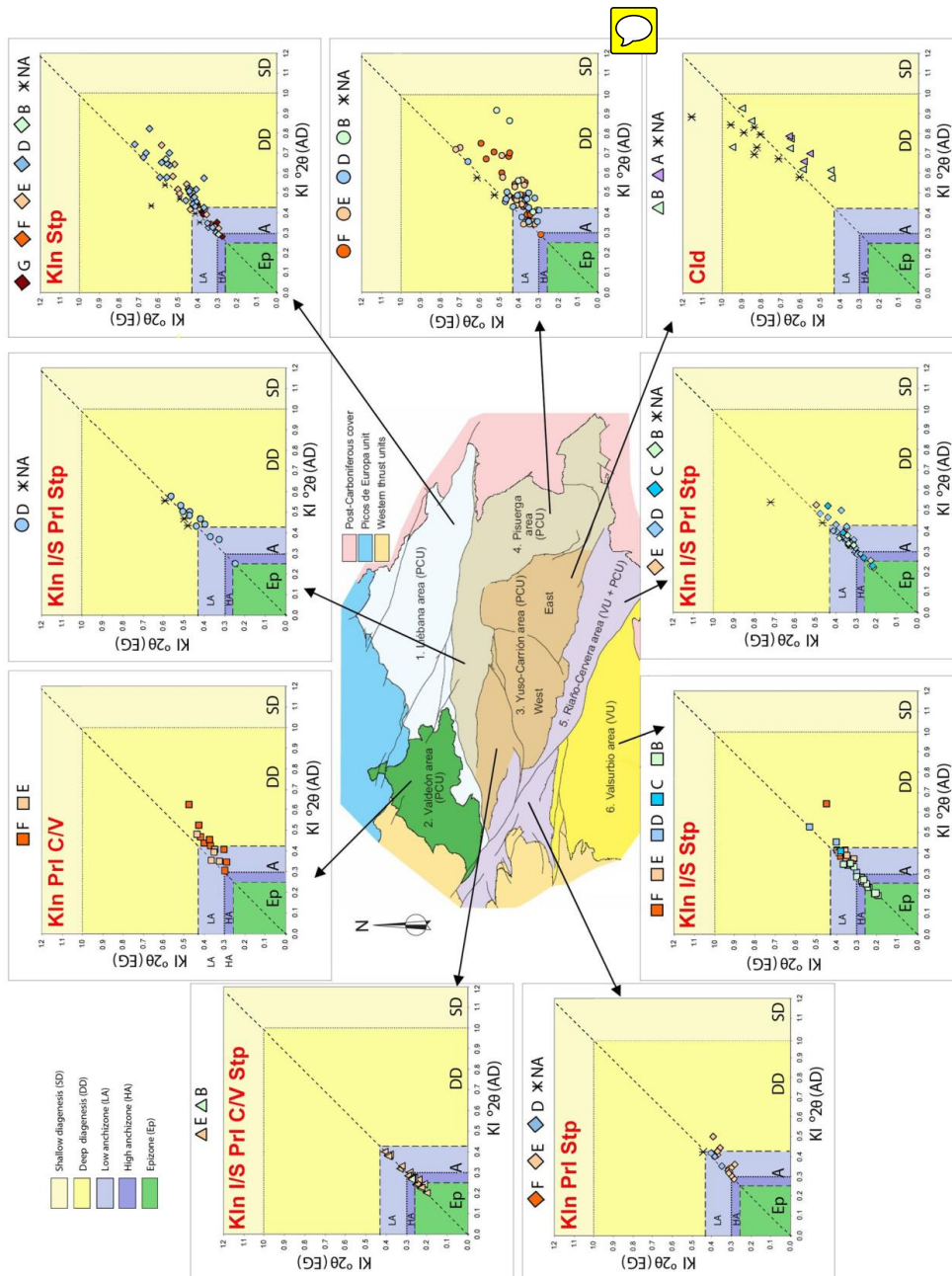

852 **Figure 6. Map ~~with~~ location of Kübler Index (KI) values (Kübler scale). Upper value, in red, air dried**

853 **sample; lower value, in green, sample treated with ethylene glycol. Árkai Index (AI) is indicated by the**

854 **colour of the sampling point. Values corresponding to samples with significant amounts of I/S, Prl and/or**

855 **Pg-Pg/Ms have not been highlighted with grade colours. Areas with small outcrops of intrusive rocks are**

856 **shadowed with purple colour. For legend see Figure 8.**

857

858 **Figure 7. Plot of Kübler Index (KI) measured on air dried (AD)** *versus* **KI measured on ethylene glycol**

859 **solvated samples (EG), standarized at Kübler scale. DD, Shallow Diagenesis; LA, Low Anchizone; HA,**

860 **High Anchizone; Ep, Epizone. Samples are plotted according to the areas outlined in Fig.4 and have been**

861 **grouped following the divisions of Fig. 2: A, Silurian; B, Devonian; C, Tournasian-Visean; D, Prioro**

862 **Group; E, Pando Group; F, Viorna Group; G, Campollo-Remoña Group. Samples with significant**

863 **amounts of I/S,Prl and/or Pg-Pg/Ms are indicated as NA (not applicable). Minerals absent in each of the**

864 **areas are indicated in red.**

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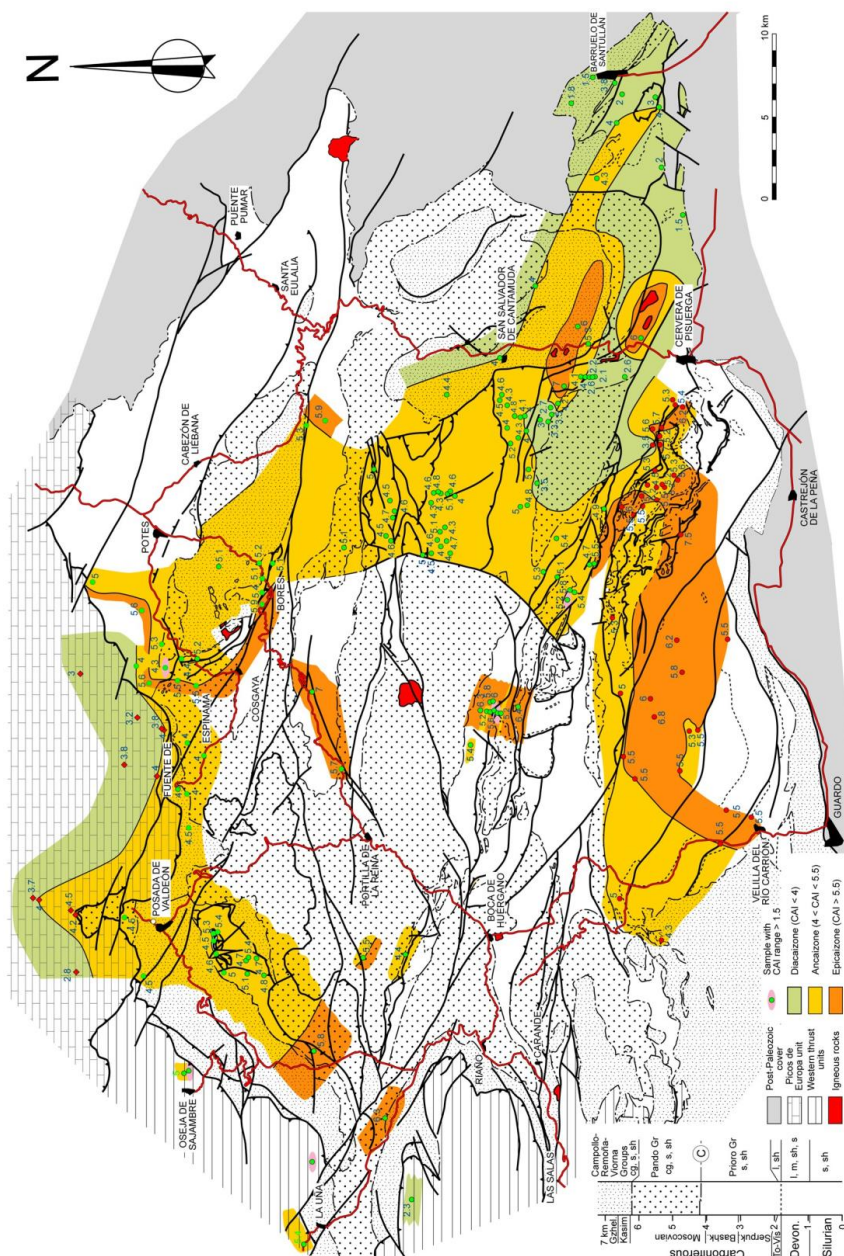
Figure 8. Map with location of CAI values and delimitation of CAI isogrades. Picos de Europa CAI data after Bastida et al., 2004; Valsurbio CAI data after García-López et al., 2013. Symbols of the CAI samples as in Fig. 2.

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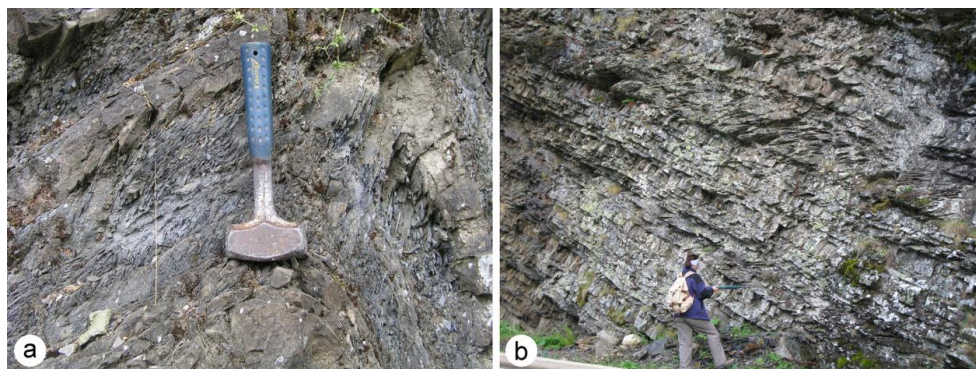
Figure 9. (a) S$_1$ cleavage associated with a nearly upright fold (eastern part of the Liébana area; north to the right). (b) S$_2$ cleavage ~~less~~ dipping than ~~the~~ bedding in a normal stratigraphic succession (Pando Group; Curavacas-Lechada syncline; north to the left).

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