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Pressure-to-depth conversion models for metamorphic rocks: derivation and applications

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Key Points:

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9	•	We present and apply different pressure-to-depth conversion models to a dataset
10		of metamorphic pressure.
11	•	The lithostatic pressure assumption results in an upper estimate of depth at peak
12		pressure $(> 100 \text{ km})$.
13	•	A change in stress state < 75 km can trigger a peak to retrograde P decrease and

• A change in stress state < 75 km can trigger a peak to retrograde P decrease and is consistent with the data.

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15 Abstract

Pressure-to-depth conversion is a crucial step towards geodynamic reconstruction but 16 remains strongly debated. Here, we derive pressure-to-depth conversion models using ei-17 ther one or two pressure data points in conjunction. In the two-point method, we assume 18 that both peak and retrograde pressure are recorded at the same depth. This method 19 reduces the depth estimate uncertainty dramatically. We apply the proposed pressure-20 to-depth conversions to a large set of P data from (ultra)high-pressure metamorphic rocks. 21 We explore different cases to explain the transition from peak to retrograde pressure by 22 varying the direction and magnitude of stress components. Our results show that (1) even 23 small deviatoric stresses have a significant impact on depth estimates, (2) the second prin-24 cipal stress component σ_2 plays an essential role, (3) several models can explain the P 25 evolution of the data but lead to different depth estimates, and (4) strain data offer a 26 means to falsify two-point models. The most commonly used pressure-to-depth conver-27 sion method uses one pressure point and the assumption that pressure is lithostatic. Then, 28 the transition from peak to retrograde pressure is interpreted as the result of deep sub-29 duction (> 100 km), followed by fast exhumation to mid-crustal depth. We show that 30 alternative models where a change in the stress state at a constant depth triggers the 31 pressure transition explain the data equally well. The predicted depth is then shallower 32 than the crustal root Moho (< 75 km) for all data points. 33

³⁴ Plain Language Summary

During the formation of mountain belts, rocks are buried deep in the Earth and 35 then exhumed. In this journey, rocks undergo transformations that record the pressure. 36 We use the pressure to estimate the depth at which a rock was buried to reconstruct the 37 history of mountain belts. The pressure is the sum of the weight of the overlying column 38 of rock and tectonic forces. However, since tectonic forces cannot be measured, there has 39 been a long-standing debate on how much they influence the record of pressure in rocks. 40 Here, we use mathematics and computer code to recalculate the burial depth of a set of 41 rock from pressure data. Two extreme scenarios emerge: (1) when ignoring tectonic forces 42 (classical approach), we interpret the pressure history as the result of deep burial (up 43 to 160 km) followed by fast exhumation (1-10 cm/yr) to approximately 20 km. The 44 mechanism of such fast exhumation is itself intensely debated; (2) when considering tec-45 tonic forces, an alternative scenario is that the rock was buried to an intermediate depth 46 (< 75 km), followed by a change in tectonic forces without exhumation. If this second 47 scenario is verified, then the current history of mountain belts must be re-evaluated. 48

49 **1** Introduction

Geodynamic reconstructions presenting cross-sections, maps, or elaborate large-50 scale plate reconstructions over time are essential to conceptualize lithospheric processes 51 such as subduction or mountain building and to reconstruct Earth's history. These geo-52 dynamic reconstructions are based on quantitative data obtained with a wide range of 53 techniques from field mapping to geophysical imaging. Among these data, pressure-temperaturetime-deformation $(P - T - t - \epsilon)$ paths obtained from petrological, geochronological, 55 and mineral deformation studies constitute key constraints. These features are indeed 56 the only way to estimate the burial, temperature and deformation evolution of a piece 57 of rock and, by extension, of the geological unit to which it belongs. In particular, es-58 timated depths, in conjunction with geochronological data, are used to reconstruct the 59 formation process of orogens (e.g., Chopin, 2003; Ernst et al., 2007; Agard et al., 2009). 60

The conversion of pressure to depth is crucial in establishing a geodynamic reconstruction based on petrographic data. Depth can be retrieved from the lithostatic pressure P_{litho} , i.e., the weight of the overlying column of rock, by the formula:

$$z = \frac{P_{litho}}{\rho g},\tag{1}$$

where ρ is the average density of the rock column, g is the gravitational acceleration and 61 z is depth. However, P_{litho} cannot be directly estimated from metamorphic rocks; in-62 stead, we can estimate the mean stress, also called the pressure, P (Moulas et al., 2019). 63 Therefore, an additional step is required to relate P to P_{litho} . This additional step in-64 volves information about the three-dimensional deviatoric stress state responsible for rock 65 deformation. Unfortunately, deviatoric stresses cannot be measured, and one must, there-66 fore, make assumptions regarding the stress state to retrieve P_{litho} . Depending on the 67 assumption made, the final depth estimate can vary by more than a factor of two. Since 68 these crucial assumptions are hard or maybe impossible to falsify, there has been a long-69 standing debate over (1) what is the most adequate stress state assumption to use for 70 pressure-to-depth conversion, (2) how deeply were metamorphic rocks buried, and (3) 71 how are metamorphic rocks exhumed (e.g., Jamieson, 1963; Ernst, 1963; Brace et al., 1970; 72 Mancktelow, 1993; Godard, 2001; Green, 2005; Agard et al., 2009; Wheeler, 2014; B. Hobbs 73 & Ord, 2015; Wheeler, 2014; Tajčmanová, 2015; Moulas et al., 2013; Gerya, 2015; B. E. Hobbs 74 & Ord, 2017; Moulas et al., 2019; Schmalholz & Podladchikov, 2014; Yamato & Brun, 75 2017; Reuber et al., 2016; Schenker et al., 2015). 76

The most common assumption is to ignore deviatoric stresses because metamor-77 phic rocks are assumed to be weak at the depths considered (e.g., Guillot et al., 2009; 78 Agard et al., 2009; Beltrando et al., 2007; Rubatto et al., 2011). Thus, $P = P_{litho}$ and 79 one can readily use eq. 1. We call this assumption the "lithostatic case". In a rock, the 80 magnitude of deviatoric stresses can vary from zero to the point of rock failure. Hence, 81 the mean deviatoric stress can be of a magnitude comparable to lithostatic pressure, and 82 P can vary from 1 to 2 times the value of P_{litho} in compression for a homogeneous rock 83 (Petrini & Podladchikov, 2000). The difference between P and P_{litho} is referred to as 84 "tectonic pressure" (Mancktelow, 2008), "tectonic overpressure" (Mancktelow, 1993; Schmal-85 holz & Podladchikov, 2013) or simply "overpressure" when it is positive or "underpres-86 sure" when it is negative (Moulas et al., 2013). Note that the overpressure model is a 87 general model of which the "lithostatic case" constitutes one special case. Therefore, it 88 is essential to consider variations in the stress state when interpreting pressure-temperature 89 (P-T) paths. 90

In most cases, the P-T evolution of a (U)HP metamorphic rock can be approx-91 imated by three linear segments. A prograde segment (highlighted in blue in Fig. 1A) 92 that shows increases in both P and T and a retrograde part (in green in Fig. 1A) divided 93 in two segments: a retrograde stage 1 and a retrograde stage 2 (see Fig. 1A). The first 94 stage of the retrograde path generally shows a large decrease in pressure and only mi-95 nor variations in temperature, while the second stage presents decreases in both pres-96 sure and temperature conditions (?, ?, see)]Yamato2017. Hereafter, we use the notations 97 P_p and T_p to refer to the pressure and temperature conditions at the peak of (U)HP meta-98 morphism (time t_1 in Fig. 1A), respectively. Similarly, P_r and T_r refer to the pressure 99 and temperature conditions at the end of retrograde stage 1 (time t_2 in Fig. 1A). 100

There are arguably two events in the P-T path that cause most of the debate: 101 peak metamorphism (P_p, T_p) and retrograde stage 1 (i.e., the transition from P_p to P_r). 102 Thermobarometric studies often provide P_p, T_p and P_r, T_r , sometimes in association with 103 geochronological dating. We present the dataset of $P_p, T_r P_r, T_r$ collected from the lit-104 erature in the P - T space in Figure 1B and in the space P_p - P_r in Figure 1C. In P_p -105 P_r space, most data points are contained within a fan centered on 0, which suggests that 106 P_p and P_r are proportional, with coefficients of proportionality, P_p/P_r , between 2.4 and 107 4.8. A few data points with values $P_p < 1.5$ have a coefficient of proportionality < 2.4108 as low as 1.4. We term these points "Others (outliers)". 109



Figure 1. (A) Typical example of a P - T path. (B) Dataset in P - T space. Colors correspond to the orogenic system from which data come as presented in C. (C) Repartition of the data (see Suppl. Mat. for references) in a P_p vs. P_r diagram

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To illustrate the consequence of stress state assumptions on geodynamic interpre-110 tations, let us consider a rock presenting a mineral paragenesis equilibrated at 3.0 GPa. 111 This rock can be interpreted as having been buried up to 100 km depth under the "litho-112 static" assumption (using $\rho = 3000 \ kg/m^3$ and $g = 10 \ m/s^2$) but only approximately 113 50 km when considering a magnitude of deviatoric stresses close to the brittle yield stress 114 in compression. While the former corresponds to mid-lithospheric depth, the latter would 115 correspond to crustal-root depth. Pleuger and Podladchikov (2014), for example, pro-116 posed a geodynamic reconstruction of the central Alps based on structural arguments 117 wherein the Adula nappe, an eclogite-bearing metamorphic unit in the Alps, was buried 118 to 50-60 km depth. This depth estimate implies an overpressure of 40-80 % of the litho-119 static pressure and suggests that the burial and exhumation of this unit occurred within 120 an orogenic crustal wedge. In alternative models using the "lithostatic assumption", the 121 nappe was buried to 80 km depth during subduction and then rapidly exhumed by slab 122 breakoff (S. M. Schmid et al., 1996; Froitzheim et al., 2003) or subvertical extreme thin-123 ning (Nagel, 2008). The scenario of S. M. Schmid et al. (1996) employs one subduction 124 zone in conjunction with a normal fault, while the models of (Froitzheim et al., 2003) 125 and Nagel (2008) involve two subduction zones. Thus, different assumptions regarding 126 pressure-to-depth conversion lead to different interpretations of the process of mountain 127 building. Therefore, it is crucial to understand, compare, and evaluate the implications 128 of different assumptions about the stress state when designing geodynamic reconstruc-129 tions. 130

Retrograde stage 1, when the pressure decreases from P_p to P_r in a relatively short 131 amount of time, is also at the center of heated debate. Using the "lithostatic assump-132 tion", the transition from P_p to P_r is interpreted as an exhumation event. In conjunc-133 tion with dating data, this phase of exhumation is generally interpreted as fast, with ex-134 humation rates comparable to subduction rates $(1-10 \ cm/yr)$ (e.g., Rubatto & Hermann, 135 2001; Parrish et al., 2006). Various mechanisms have been proposed to explain these fast 136 exhumation rates, such as buoyancy-driven exhumation (Wheeler, 1991; Beaumont et 137 al., 2009; Butler et al., 2013, 2014; E. Burov et al., 2014; Schmalholz & Schenker, 2016), 138 slab breakoff (Huw Davies & von Blanckenburg, 1995), normal faulting (Platt, 1986; Ring 139 et al., 1999; S. M. Schmid et al., 1996), rollback (Brun & Faccenna, 2008), or channel 140 flow (e.g., Guillot et al., 2009). These and other mechanisms are discussed in detail in 141 several reviews (Guillot et al., 2009; B. R. Hacker & Gerya, 2013; Warren, 2013). In con-142 trast to the fast exhumation interpretation, Yamato and Brun (2017) showed that when 143 considering the large deviatoric stresses assumption, the transition from P_p to P_r can 144 be explained, for many rock samples, by a switch from a compressional to an extensional 145 stress state without exhumation. 146

In this contribution, we first review the mathematical background of pressure and 147 stress. Then, we formulate a "one-point method" of pressure-to-depth conversion to es-148 timate depth based on a single pressure data point and a "two-point method" that uses 149 both P_p and P_r with the assumption that $z_p = z_r$. We apply these methods to our dataset 150 (Fig. 1B) to determine an estimated depth range for each sample. Finally, we discuss 151 the consequences of different assumptions for geodynamic interpretation and point out 152 ways of falsifying some assumptions. Our goal is both to raise awareness about the is-153 sue of pressure-to-depth conversion and to provide tools allowing one to perform such 154 conversion easily. For this reason, we provide computer codes (Jupyter notebooks) as 155 supplementary information S2-S10. These scripts allow readers to reproduce most of the 156 figures presented in this article readily and to extend the database with their own data. 157 The codes can also be used to experiment with stress states and material properties. 158



Figure 2. Overview of the principal characteristics of the model and definitions. See text for details concerning notation. $\Delta \sigma_x$ corresponds to the stress magnitude applied in the *x*-direction. Graphics presenting σ_n vs. τ (i.e., normal stress vs. shear stress) correspond to Mohr diagrams.

¹⁵⁹ 2 One-point method of pressure-to-depth conversion

2.1 Overview of the model

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2.1.1 Sketch, coordinate system and equations of stress

Let us consider an ideal and simplified orogen submitted to horizontal tectonic stresses in a three-dimensional Cartesian orthonormal system (x, y, z) where z is vertical and points downward and x is the direction in which tectonic loading is applied (Fig. 2A). $\sigma_x, \sigma_y, \sigma_z$ are the normal components of the stress tensor in this coordinate system, and $\sigma_1, \sigma_2, \sigma_3$ are the principal stresses. We use the convention that stresses are positive in compression. We assume, in a first step, that the stress state is Andersonian, that is, one principal stress direction is vertical, and the other two are horizontal (Anderson, 1905). We fix the y-axis in the direction of σ_2 . Thus, we only consider cases where the stress state can induce normal or reverse faulting, and we ignore the stress states that would result in strike-slip faulting. Under these assumptions, the total vertical stress σ_z corresponds to the weight of the column of rock above the considered point (or P_{litho}) and is given by:

$$\sigma_z = \rho g z, \tag{2}$$

where ρ is the density of the overlying rocks, g is the gravitational acceleration and z is the depth where the computation is performed. When a tectonic stress of magnitude $\Delta \sigma_x$ is applied in the x-direction, the following equation applies:

$$\sigma_x = \sigma_z + \Delta \sigma_x \tag{3}$$

Three tectonic regimes can be considered depending on the horizontal loading condition (Fig. 2A): (1) lithostatic, when $\Delta \sigma_x = 0$; (2) compression, when $\Delta \sigma_x > 0$; (3) extension, when $\Delta \sigma_x < 0$. Equations describing the stress state for these three tectonic regimes are presented in Figure 2B.

166 2.1.2 Deformation

The magnitude of the deformation is proportional to $\Delta \sigma_x$, and the direction of maximum stretch is parallel to the direction of σ_3 . Thus, there is no deformation in the lithostatic case, and the maximum stretch is horizontal in the extensional case and vertical in the compressional case. The total stress in the y-direction is always $\sigma_y = \sigma_2$, and we use the variable α that ranges between 0 and 1 to describe σ_2 as a function of σ_1 and σ_3 such that:

$$\sigma_2 = \alpha \sigma_1 + (1 - \alpha) \sigma_3. \tag{4}$$

Figure 2C shows how α is related to the mode of deformation. When $\alpha = 0$, $\sigma_2 = \sigma_3$ (see eq. 4), and the rock deforms by flattening. When $\alpha = 1$, $\sigma_2 = \sigma_1$, and the rock deforms by constriction. When $\alpha = 0.5$, $\sigma_2 = (\sigma_1 + \sigma_3)/2 = P$, and the deformation is plane strain.

171 **2.1.3** Pressure

By definition, pressure (P) corresponds to the isotropic part of the stress tensor and, in principal stress coordinates, it can be expressed as follows:

$$P = \frac{\sigma_1 + \sigma_2 + \sigma_3}{3}.\tag{5}$$

Hence, $P = \sigma_z$ in the lithostatic case, $P < \sigma_z$ in extension (because $\Delta \sigma_x < 0$), and $P > \sigma_z$ in compression (because $\Delta \sigma_x > 0$). The Mohr diagrams in Figure 2D illustrate these relationships. In the diagrams, the horizontal and vertical axes represent the

normal stress σ_n and shear stress τ on planes within the rock mass, respectively. Pressure is represented by a cross symbol, where the central vertical bar represents the value of pressure when the rock deforms under plane-strain conditions ($\alpha = 1/2$) and the horizontal bar represents the range of pressure associated with values of α between 0 (flattening) and 1 (constriction). The equation for P as a function of α is obtained by substituting eq. (4) into eq. (5), which yields

$$P = \frac{(1+\alpha)}{3}\sigma_1 + \frac{(2-\alpha)}{3}\sigma_3.$$
 (6)

2.1.4 Limit of stress and rock failure

When tectonic loading is applied, rocks first undergo elastic or viscous deformation. Stress loading can be increased up to the point where the rock breaks. At this point, the maximum stresses on a given plane within the rock are given by the Mohr-Coulomb law as:

$$\tau = \tan \phi \sigma_n,\tag{7}$$

where ϕ is the friction angle. Rock experiments show that $\phi \approx 30^{\circ}$ for most rock types (Byerlee, 1978). To simplify the derivation, we ignore cohesion since it is small (order of 10 - 50 MPa) compared to the pressure of metamorphic rocks considered here (order of GPa). The supplementary scripts (supplementary information S2 to S10) also allow the reader to reproduce most figures in this publication while taking cohesion into account (see Yamato and Brun (2017) for the derivation). Mohr's circle is defined by

$$\sigma_n = \frac{\sigma_1 + \sigma_3}{2} - \frac{\sigma_1 - \sigma_3}{2} \sin \phi \tag{8}$$

and

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$$\tau = \frac{\sigma_1 - \sigma_3}{2} \cos \phi. \tag{9}$$

Substituting eq. (8) and eq. (9) into eq. (7) yields

$$\sigma_1 = \Phi \sigma_3$$
, with (10)

$$\Phi = \frac{1 + \sin\phi}{1 - \sin\phi}.\tag{11}$$

Figure 2E illustrates the possible stress states associated with different tectonic regimes. This figure presents the whole range of possibilities from the "lithostatic" case to the brittle case.

In extension, $\sigma_x = \sigma_3$, and $\sigma_z = \sigma_1$; therefore, the minimum total horizontal stress is $min(\sigma_x) = \sigma_z/\Phi$ (Fig. 2E, middle panel). Conversely, in compression, $\sigma_x = \sigma_1$, and $\sigma_z = \sigma_3$; therefore, the maximum total horizontal stress is $max(\sigma_x) = \Phi\sigma_z$ (Fig. 2E, right panel). The quantity $(\sigma_1 - \sigma_3)/2$, i.e., the radius of the Mohr circle, is also called the second invariant of the deviatoric stress tensor or the "magnitude of deviatoric stresses".

181 2.1.5 Summary

Finally, Figure 2F presents a Mohr-Coulomb diagram that summarizes the discussion to this point. The diagram is presented in a non-dimensional form where the overbar indicates that a quantity is normalized by σ_z (e.g., $\bar{\sigma}_x = \sigma_x/\sigma_z$). The pressure in the lithostatic case, or lithostatic pressure, is equal to σ_z (i.e., the weight of the column of rocks). The nondimensional lithostatic pressure is therefore equal to $\bar{\sigma}_z = 1$ (Fig. 2F). In compression, the normalized total horizontal stress $\bar{\sigma}_x$ can vary from 1 (no deformation) to Φ (onset of brittle deformation), and $P > \sigma_z$. In extension, $\bar{\sigma}_x$ can vary

from $1/\Phi$ (brittle deformation) to 1 (no deformation), and $P < \sigma_z$. In these three cases, following eq. (5), the nondimensional pressure \bar{P} can then be written as:

$$\bar{P}_l = 1, \tag{12}$$

$$\bar{P}_e = \frac{2-\alpha}{3}\bar{\sigma}_x + \frac{1+\alpha}{3},\tag{13}$$

$$\bar{P}_c = \frac{1+\alpha}{3}\bar{\sigma}_x + \frac{2-\alpha}{3},\tag{14}$$

where the subscripts c, e, and l relate to the compression, extension and lithostatic tectonic regimes, respectively (see also Fig. 2F). Another useful result is obtained by solving the previous equations for σ_z :

$$\begin{cases} \sigma_z = \frac{3P}{1 + \alpha + \bar{\sigma}_x(2 - \alpha)}, \text{ when } \bar{\sigma}_x \le 0, \\ \sigma_z = \frac{3P}{2 - \alpha + \bar{\sigma}_x(1 + \alpha)}, \text{ when } \bar{\sigma}_x \ge 0. \end{cases}$$
(15)

2.2 Pressure-to-depth conversion ratio z/P

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To convert metamorphic pressure estimates (P) into depth (z), one can use the simple relation $z = \frac{z}{P}P$, where z/P is the gradient of depth as a function of pressure, which we refer hereafter as the "pressure-to-depth conversion ratio", expressed in km/GPa, and is equal to

$$\frac{z}{P} = \frac{1}{\rho g \bar{P}}, \text{ where}$$
(16)

$$\bar{P} = P/\sigma_z.$$
(17)

Figure 3 shows graphs of $1/\bar{P}$ and z/P as a function of the horizontal stresses expressed 183 by $\bar{\sigma}_x$ (horizontal axis) and α (different lines). The graphs were calculated by substitut-184 ing \bar{P} in eq. (16) with eq. (13) when $\bar{\sigma}_x \leq 1$ (i.e., in extension) or eq. (14) when $\bar{\sigma}_x \geq 1$ 185 1 (i.e., Panels A to E show Mohr diagrams illustrating the stress state for given values 186 of $\bar{\sigma}_x$). Throughout this article, we use $\rho g = 28000 \ kg/m^2/s^2$, representing crustal rocks. 187 A value of $tan(\phi) = 0.6$ is often used in the literature. This value is the result of fit-188 ting data from rock friction experiments by Byerlee (1978). In the main article, we use 189 the value $\tan(\phi) = 0.65$ that offers a better fit to the data in the absence of cohesion. 190 The difference has only a negligible influence on pressure estimates. Readers can easily 191 recompute our results using any value of cohesion, ϕ , or ρ by using the scripts provided 192 in the supplementary material (supplementary information S2 to S10). 193

When the pressure is considered lithostatic ($\bar{\sigma}_x = 1$, Fig. 3C), the pressure-to-194 depth conversion ratio is $z/P = 35 \ km/GPa$. However, this ratio varies significantly 195 when $\bar{\sigma}_x$ increases (compression) or decreases (extension). For example, in the case where 196 $\sigma_2 = \sigma_1$ and $\bar{\sigma}_x$ is minimum, $z/P = 64 \ km/GPa$ (Fig. 3A). In contrast, when $\bar{\sigma}_x$ is 197 maximum, $z/P = 16 \ km/GPa$ (Fig. 3E). Small deviations of $\bar{\sigma}_x$ from 1 have signifi-198 cant impacts on the pressure-to-depth conversion ratio. For example, when the applied 199 tectonic stress $\Delta \bar{\sigma}_x = min(\Delta \bar{\sigma}_x)/4$, $z/P = 39 \ km/GPa$ (Fig. 3B), and when $\Delta \bar{\sigma}_x =$ 200 $max(\Delta \bar{\sigma}_x)/4$, $z/P = 25 \ km/GPa$ (Fig. 3D). The value of α also exerts a strong con-201 trol on the pressure-to-depth conversion ratio, particularly in extension; e.g., when $\bar{\sigma}_x =$ 202 Φ^{-1} , the conversion ratio varies from 45 to 64 km/GPa depending on the value of α . 203

2.3 Application of the one-point method

We now apply the pressure-to-depth conversion ratio derived in the previous section to our dataset of peak (P_p) and retrograde (P_r) metamorphic pressures. Figure 4 shows the depths estimated from this conversion. Depth estimates at peak pressure (z_p)



Figure 3. Pressure-to-depth conversion ratio (z/P) as a function of normalized horizontal stress $\bar{\sigma}_x = \sigma_x/\sigma_z$. The vertical axis indicates the ratio of vertical stress to pressure (Sz/P), blue axis) or the pressure-to-depth conversion ratio (z/P), red). We use $\rho g = 28000 \ kg/m^2/s^2$, and $\tan \phi = 0.65$. The three lines correspond to different values of α (i.e. σ_2). The mode of deformation associated with α is illustrated by the cartoons on the left, where the white and blue boxes represent the undeformed and deformed states, respectively. The Mohr diagrams represent the outer envelope of stress states at points A to E. In these panels, pressure P is represented by a cross, as in figure **??**F.



Figure 4. Depth estimates at peak pressure (A) and at retrograde pressure (B) for all samples in our dataset. Colors are coded for areas. The "normal Moho" depth corresponds to the average depth of the continental Moho in regions where the crust is neither thickened nor thinned and is 30 km. The deepest Moho (75 km) corresponds to the current depth of the Moho below the Tibetan Plateau. This figure can be reproduced using the computer script from supplementary information S4.

are shown only for compressive stress states (Fig. 4A), while depth estimates at retro-208 grade pressure (z_r) are shown for both compressive and extensional stress states (Fig. 209 4B). We indicate two reference depths: (a) 30 km (red dashed line), which is the depth 210 of a "normal continental Moho" defined as the thickness of an isostatically balanced con-211 tinental crust with topography at sea level, and (b) 75 km (thick red line), which is the 212 depth of the Moho below the Tibetan Plateau and is the present-day "deepest Moho" 213 on Earth. For each sample, the black horizontal bar indicates the lithostatic pressure case. 214 The two columns for each sample indicate the two extreme deformation regimes: flat-215 tening ($\alpha = 0$, or $\sigma_2 = \sigma_3$) and constriction ($\alpha = 1$, or $\sigma_2 = \sigma_1$). 216

At peak pressure conditions, the upper estimate of depth z_p (Fig. 4A, black bars) 217 corresponds to lithostatic conditions (i.e., with no deformation), with a conversion ra-218 tio z/P = 35 km/GPa (Fig. 4A and Fig. 3C). Under this condition, z_p values are ap-219 proximately 165 km for samples from the Kokchetav and Sulu-Dabie regions, 140 km for 220 the Tso Morari and Caledonides, and 120 km for the Alps and Kaghan valley. The min-221 imum estimate of z_p results from assuming constricting deformation at brittle failure un-222 der compression (i.e., $\alpha = 1$ and $\bar{\sigma}_x = \Phi$). The conversion ratio is then z/P = 16km/GPa223 (Fig.3E) and $z_p < 75 km$ for all samples, i.e., shallower than the present-day deepest 224 Moho on Earth. The uncertainty range for z_p for a single data point varies from ≈ 15 km 225 for sample #40 to ≈ 100 km for samples #16 and #23. 226

Under retrograde conditions, the lithostatic case represents an intermediate esti-227 mate because we consider both compressive and extensive tectonic regimes (Fig. 4B). 228 The upper estimate for z_r results from assuming flattening deformation at brittle fail-229 ure in extension (i.e., $\alpha = 0$ and $\bar{\sigma}_x = 1/\Phi$). The conversion ratio is then z/P = 64km/GPa230 (Fig. 3A). A few samples from the Alps have a maximum depth estimate of $z_r > 85$ 231 km. For samples from the Kokchetav and Sulu-Dabie orogens, $z_r = 75$ km, and $z_r =$ 232 50 km for samples from the Kaghan valley, Tso Morari and Caledonides. The minimum 233 estimate of z_r results from assuming constricting deformation at brittle failure in com-234 pression (i.e., $\alpha = 1$ and $\bar{\sigma}_x = \Phi$). z_r can be as shallow as 10 to 20 km for all sam-235 ples. The uncertainty range on the estimate of z_r for a single data point is up to 70 km 236 for sample #11 whose maximum depth is ≈ 90 km. All samples have at least part of 237 their range shallower than the deepest present-day Moho at both peak and retrograde 238 pressures. 239

Figure 5A shows the estimated exhumation calculated as the difference between z_{p} and z_{r} . We present six special cases involving different values of $\bar{\sigma}_{x}^{p}, \bar{\sigma}_{x}^{r}, \alpha_{r}$, and α_{r} to illustrate the dependence of the estimated exhumation on the stress state. In Figure 5(C-H), we present Mohr diagrams for these six cases calculated using P_{p} and P_{r} from a reference sample.

The maximum exhumation is predicted when P_p corresponds to lithostatic pressure and P_r is recorded at brittle failure in compression (Fig. 5A, top of color bars, and Fig. 5C). The maximum predicted exhumation in our dataset varies between 20 and 150 km.

We use the term "always lithostatic" for the case where both P_p and P_r are lithostatic pressures. This case is shown with black horizontal bars in Figure 5A-B and illustrated in Figure 5D. In this case, exhumation varies between 25 km and 125 km for our dataset. Since the "always lithostatic" case is the most commonly used solution in the literature, we use it as a reference to normalize the results. The normalized graph (Fig. 5B) allows us to express the exhumation amount as a percentage of a reference case and outline similarities between samples.

The red rectangle symbol (Fig. 5A-B) corresponds to a case where deformation is compressive for P_p and extensive for P_r , the magnitude of deviatoric stress is a quarter of the maximum value, and deformation is plane strain (Fig. 5E). This stress state rep-



Figure 5. (A) Estimated amount of exhumation calculated as the difference between the estimated depth at peak and retrograde pressures $(z_p - z_r)$ for all samples from our dataset. Bar colors indicate the provenance. (B) Same as (A) but normalized by the amount of exhumation obtained by considering P_p and P_r as lithostatic pressures (i.e., "always lithostatic" case). We calculated six special cases by combining different values of $\bar{\sigma}_x^p, \bar{\sigma}_x^r, \alpha_p, \alpha_r$, i.e., maximum and minimum exhumation, and four intermediate cases (colored rectangles). The values used are shown in the table inset in (B). (C-H) Mohr diagrams for the six special cases using P_p, P_r from a reference sample indicated by red arrows in (A) and (B). The characteristics of the special cases are (C) maximum exhumation case, (D) "always lithostatic" case, (E-F) cases with moderate deviatoric stress, (G) exhumation amount close to zero, and (H) minimum exhumation (negative exhumation, i.e., burial). This figure can be reproduced using the computer script from supplementary information S5.

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resents a conservative estimate for rocks that deform by viscous deformation at depth. 259 This low deviatoric stress has a significant impact on the quantity of exhumation: on av-260 erage, this case results in an estimate of exhumation that is only 60% of that for the "al-261 ways lithostatic" case (see red line in Fig. 5B). The blue rectangle symbol (Fig. 5A-B) represents a case of intermediate stress where the magnitude of deviatoric stress is half 263 of the maximum value (Fig. 5F). On average, this case's results are 35% of the estimate 264 for the "always lithostatic" case (see red line in Fig. 5B). The dark yellow rectangles in-265 dicate the scenario where deformation is brittle in compression at peak pressure, and P_r 266 corresponds to lithostatic pressure under plane strain deformation. This scenario pre-267 dicts at most 30 km of exhumation and a minimum of -10 km (i.e., 10 km of additional 268 burial). In this case, the predicted exhumation is $\approx 10\%$ of that for the "always litho-269 static" case, on average. The minimum exhumation estimate is obtained when deforma-270 tion is brittle and constrictive in compression for P_p , and deformation is brittle in ex-271 tension and occurs by flattening for P_r . The minimum exhumation estimate is between 272 0 and -50 km. 273

For most samples, the normalized amount of exhumation for a specific case, e.g., low stress (red rectangles), is contained within a small range around an average value. However, the samples from the category "Others (outliers)" have significantly different values. Although their values of P_p and P_r are not anomalous (e.g., Fig. 1), their combination clearly differs from other samples (see Fig. 1C). The relatively low dispersion of exhumation is related to the apparent proportionality between P_p and P_r (see Fig. 1C).

In this section, we show that one can interpret the transition from P_p to P_r as the result of exhumation from great depth (Fig. ??C-D). The data are also compatible with an opposite interpretation: that this transition is the result of a change in stress state while depth is constant (Fig. ??G) or even while burial continues (Fig. ??H).

3 Two-point method of pressure-to-depth conversion

In this section, we re-examine our dataset with the additional constraint that $z_p = z_r$. In this way, we can use P_p and P_r together to reduce the uncertainty range for the depth estimate. We call this method "two-point pressure-to-depth conversion". In the case of a homogeneous rock and ignoring the possible role of fluids, the stress state can be modified in only two ways: (1) by modifying the magnitude of the horizontal stress or (2) by rotating the stress field. We explore these mechanisms independently, as well as a special case, in the following sections.

293

3.1 Mechanism 1: change in the magnitude of horizontal stress (S-model1)

First, we consider the change in pressure triggered by a change in the magnitude 294 of the horizontal stress ($\bar{\sigma}_x$). Figure 6A shows five Mohr circles constructed with var-295 ious values of $\bar{\sigma}_x$. Note that the Mohr circle with $\bar{\sigma}_x = 1$ is a point. In Figure 6B-J, 296 we represent our dataset as colored circles in the P_p vs. P_r space. These data points are 297 placed on top of a colored contour map of $\bar{\sigma}_x^r$ computed for given values of $P_p, P_r, \bar{\sigma}_x^p, \alpha_p, \alpha_r$, 298 where subscripts or overscripts p and r refer to the peak and retrograde stages, respec-299 tively. The values used are indicated at the top of columns and the beginning of rows 300 of panels. A contour map of $z = z_p = z_r$ is also shown (black horizontal lines). The 301 range of values calculated for $\bar{\sigma}_x^r$ covers stress states that do not exceed the Coulomb fail-302 ure criterion. Gray areas correspond to areas where $\bar{\sigma}_x^r$ has no meaningful solution (i.e., 303 because the stress magnitude would exceed the brittle yield stress). This means that the model cannot explain data plotting in the gray area. In contrast, when a data point is 305 on top of the colored contour map, the combination of P_p, P_r for this data point can be 306 obtained using eqs. 13 to 16, the combination of $\bar{\sigma}_{x}^{p}, \alpha_{p}, \alpha_{r}$ given and the value of $\bar{\sigma}_{x}^{r}$ and 307 z shown by the contour map. 308



Figure 6. Results for the horizontal stress change-driven model. (J) A Mohr diagram illustrating the stress states associated with different values of $\bar{\sigma}_x$. The normal stress (horizontal axis) and shear stress (vertical axis) are normalized by σ_z . (B to J) Peak pressure as a function of retrograde pressure for data (colored circles) and model (colored contour plot). The estimated depths, in km, for each model are indicated by black contour lines. Gray areas indicate zones where the model does not have a solution (i.e., the deviatoric stress would exceed the yield stress). The model peak pressure is calculated from eq. (16) with parameters $\rho g = 28000 kg/m^2/s$, $\tan(\phi) = 0.65$, $\bar{\sigma}_x = \bar{\sigma}_x^p$, $\alpha = \alpha^p$. The model retrograde pressure uses $\bar{\sigma}_x^r$, α^r . Each panel in a row uses the value of $\bar{\sigma}_x^p$ indicated in the leftmost panel of the row. Each panel in a column uses the values of α^p and α^r indicated at the top of the column. This figure can be reproduced using the computer script from supplementary information S6.

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When the stress state is lithostatic at peak conditions, i.e., $\bar{\sigma}_{r}^{p} = 1$, only outliers 309 plot in the solution domain (Fig. 6B to D), which means that the transition from P_p to 310 P_r observed in the data cannot be explained only by increasing or decreasing the hor-311 izontal stress at constant depth if the stress state is lithostatic under peak conditions. 312 When the initial horizontal stress is $\bar{\sigma}_r^p = (1 + \Phi)/2$, a few data points lie in the solu-313 tion domain for the combinations $\alpha_p = 0, \alpha_r = 1$ (Fig. 6E) and $\alpha_p = \alpha_r = 1/2$ (Fig. 314 6F). However, approximately half of the points lie in the solution domain when $\alpha_p =$ 315 1, $\alpha_r = 0$ (Fig. 6G). Outliers can be explained by $\bar{\sigma}_r^r > 1$ (i.e., compressive stress state), 316 while other points are explained by $\bar{\sigma}_x^r < 1$ (i.e., extensional stress state, Fig. 6G). When 317 the initial horizontal stress is $\bar{\sigma}_x^p = \Phi$ (i.e., brittle deformation), few data points lie in 318 the solution domain for the combinations $\alpha_p = 0, \alpha_r = 1$ (Fig. 6H). When $\alpha_p = \alpha_r =$ 319 1/2, half the points lie in the solution and these points correspond to values of $\bar{\sigma}_x^r < 1$ 320 (except for outliers, Fig. 6I). When $\alpha_p = 1, \alpha_r = 0$, all the points have a solution (Fig. 321 6J). Most points have $\bar{\sigma}_x^r < 1$, but a few points are also associated with small values 322 $\bar{\sigma}_x^r > 1$. Outliers are characterized by high values of $\bar{\sigma}_x^r$. 323

In all models except the one in Fig. 6J, some data points have a higher P_p than 324 acceptable within the model bounds. On the other hand, there is no data point with P_i 325 lower (or P_r higher) than that predicted by the model. The outlier points also plot within 326 the bounds of the model. Overall, each data point is within the model boundaries or close 327 to its boundary on at least one graph (e.g., Fig. 6J). Therefore, the model where the tran-328 sition from P_p to P_r is triggered by a change in the stress state at constant depth ($z_p =$ 329 z_r) explains the data. While some points lie within the model boundaries only for a de-330 viatoric stress with a large magnitude, other points can be explained by a change in stress 331 with only moderate deviatoric stresses (Fig. 6G). Values of α_p, α_r are also important to 332 explain the data; e.g., some data points can be explained only when $\alpha_n = 1, \alpha_r = 0$. 333 For these data points, the model predicts a change in the mode of deformation from con-334 striction to flattening during the transition from P_p to P_r . Therefore, analyses of the mode 335 of deformation in metamorphic rock samples provide a way to validate or falsify our model. 336

We compute the depth depending on the given value of $\bar{\sigma}_r^p$ and α_p from eqs. 14 and 337 16. For $\bar{\sigma}_x \geq 1$, the pressure-to-depth conversion ratio increases with decreases in both 338 $\bar{\sigma}_x$ and α (see Fig. 3). Graphically, this is expressed as the spacing between depth con-339 tours widening towards the right (e.g., from 6B to D) and bottom panels (e.g., from 6B 340 to H). The cases where $\bar{\sigma}_x^p = 1$ provide the highest pressure-to-depth conversion, but 341 only outliers lie within the solution domain. Their maximum depth is approximately 55 342 km (Fig. 6D). The deepest depth estimates, approximately 75 km, are obtained when 343 $\bar{\sigma}_x^p = (1+\Phi)/2$ (Fig. 6F-G). In the case where $\bar{\sigma}_x^p = \Phi$, many points lie in the solu-344 tion range, but a low pressure-to-depth conversion ratio limits the depth. Thus, the max-345 imum depth is approximately 65 km (Fig. 6I-J). We discuss depth estimates in detail 346 in section 3.4. 347

348

3.2 Mechanism 2: stress rotation (S-model2)

We now consider the change in pressure triggered by a rotation of the stress field. We assume that when the rock records P_p , the vertical and horizontal directions are principal stress directions, as in the previous sections. Then, the stress field rotates by an angle θ around axis y, and the rock records P_r . Figure 7A shows Mohr circles with five different values of θ .

Graphically, when we apply a rotation to a stress state where σ_1 is initially horizontal (i.e., compressional tectonic regime), the Mohr circle is shifted to the left (Fig. 7A). The maximum shift corresponds to $\theta = 90^{\circ}$, and σ_1 is vertical (i.e., extensional tectonic regime). Eventually, the Mohr circle may become tangent to the Coulomb yield envelope. Since the model does not admit stress states beyond this envelope, the radius



Figure 7. Summary of results for the stress rotation-driven model. θ is the counterclockwise rotation angle. (A) Mohr diagram illustrating the stress states associated with different amounts of rotation. The normal stress (horizontal axis) and shear stress (vertical axis) are normalized by σ_z . (B to H) Peak pressure as a function of retrograde pressure for different parameters (see text for details). Colored dots correspond to the data from our dataset. The estimated depths, in km, for each model are indicated by black horizontal lines. Gray areas indicate zones where the model does not have a solution. We use parameters $\rho g = 28000 kg/m^2/s$ and $\tan(\phi) = 0.65$. Each panel in a row uses the value of $\bar{\sigma}_x^p$ indicated in the leftmost panel of the row. Each panel in a column uses the values or α^p and α^r indicated at the top of the column. The colors of the contour maps are coded for values of θ . This figure can be reproduced using the computer script from supplementary information S7.

of the Mohr circle has to decrease upon further rotation to remain tangent to it (see Fig. 7A, $\theta \ge 45^{\circ}$).

To formalize this behavior mathematically, we first define the yield function for Mohr-Coulomb plasticity:

$$F = \frac{\sigma_1 - \sigma_3}{2} - \frac{\sigma_1 + \sigma_3}{2} \sin \phi.$$
 (18)

Then, the principal stresses σ_1 and σ_3 as a function of θ are expressed as

when
$$F < 0$$
,
$$\begin{cases} \sigma_3 = \sigma_z \left(1 + (\bar{\sigma}_x^p - 1) \frac{\cos 2\theta - 1}{2} \right), \\ \sigma_1 = \sigma_z + \tau_{II} (\cos 2\theta + 1), \end{cases}$$
(19)

when
$$F = 0$$
,
$$\begin{cases} \sigma_3 = \frac{2\sigma_z}{1 + \Phi - \frac{\Phi - 1}{\cos 2\theta}}, \\ \sigma_1 = \Phi\sigma_3. \end{cases}$$
 (20)

Note that the first equation is only valid for $\bar{\sigma}_x^p \geq 1$. σ_2 is calculated using eq. 4, P_p is the mean stress for $\theta = 0$, and P_r is the mean stress for a given value of θ .

Figures 7B-H are constructed in the same way as Figure 6, but the brownish colored contour map now represents θ .

A lithostatic stress state is isotropic. Thus, pressure remains constant upon rota-365 tion $\bar{\sigma}_x^p = \bar{\sigma}_x^r = P_p = P_r = 1$ (Fig. 7B). The resulting line in the P_p, P_r space does 366 not cross the data cloud, i.e., does not explain the data. For stress states tangent to the 367 Coulomb envelope, the upper limit of the contour map for θ ($\theta = 90^{\circ}$) is the same as 368 the upper limit of $\bar{\sigma}_x^r$ ($\bar{\sigma}_x^r = 1/\Phi$) (see Fig. 6), while the lower limit ($\theta = 0^\circ$) corre-369 sponds to the case $\bar{\sigma}_x^r = \bar{\sigma}_x^p$ in the previous model. Therefore, the boundaries of the model 370 are similar for this model (involving θ) and for the previous model (involving $\bar{\sigma}_r^r$). Al-371 though extreme stress states are identical, intermediate cases are different (e.g., compare 372 Figs. 6A and 7A). When $\bar{\sigma}_x^p = (1 + \Phi)/2$, a minimum of $\theta = 45^\circ$ is required to ex-373 plain the data (Fig. 7C-E). As with the previous model, all data points are consistent 374 with a model where $\bar{\sigma}_x^p = \Phi, \alpha_p = 1, \alpha_r = 0$ (Fig. 7H). In this case, outliers are ex-375 plained by $\theta = 0-30^{\circ}$ and other points by $\theta = 30-90^{\circ}$. Since the depth contour map 376 is computed based on $\bar{\sigma}_{p}^{p}$, α_{p} only and the model range is similar to the previous model, 377 the remarks concerning depth made in section 3.1 also apply here. 378

Since both this model (Fig. 6) and the previous one (Fig. 7) can explain the data, there is an ambiguity about which mechanism is responsible for the stress change. Once again, the predictions of this model can be validated or falsified using strain data. Indeed, the rotation of the principal stress directions implies a rotation of the principal strain direction.

384 385

3.3 A special case: compression to extension in the brittle limit (YB-model)

When $\bar{\sigma}_x^p = \Phi$, depending on the values of α_p, α_r , the solution for $\theta = 90^\circ$ (which 386 corresponds to the upper limit of the solution domain) can outline the lower extent of 387 the data point cloud (Fig. 7F), pass through it (Fig. 7G), or outline its upper extent (Fig. 388 7H). In other terms, the data distribution can also be explained by a more restrictive 389 model where depth is constant, $\bar{\sigma}_x^p = \Phi, \theta = 90^\circ$ (or $\bar{\sigma}_x^r = 1/\Phi$, cf. Fig. 6) and α_p 390 and α_r are free parameters. This model has previously been employed by Yamato and 391 Brun (2017). Here, we extend their analysis by providing the associated pressure-to-depth 392 conversion. 393

To obtain a mathematical expression for P_p , we substitute eq. (14) with $\bar{\sigma}_x = \Phi$ for \bar{P} in eq. (16) and solve for P. For P_r , we use eq. (13) with $\bar{\sigma}_x = 1/\Phi$ instead of eq.



Figure 8. Data points in the P_p vs. P_r space. Contours of depth according to (a) the lower estimate and (b) upper estimate of our model. The model has solutions within the white fan and no solution in the gray domain. The color of the data points indicates the geographic region. This figure can be reproduced using the computer script from supplementary information S8.

(14). This process yields:

$$P_{p} = \frac{\rho g z}{3} (2 - \alpha^{p} + \Phi(1 + \alpha^{p})), \qquad (21)$$

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$$P_r = \frac{\rho g z}{3} (1 + \alpha^r + \Phi^{-1} (2 - \alpha^r)).$$
(22)

Similar to previous figures, Figure 8 shows the domain of the solution of eqs. (21)394 and (22) for values of α_p and α_r between 0 and 1. Data points are also plotted in this 395 P_p vs. P_r space. We also show contours of depth obtained by solving eq. (21) or (22) 396 for z. The value of ρq influences the distance between depth contours but not the shape 397 of the solution domain. The parameter Φ (or ϕ , cf eq. 11) controls the orientation and 398 opening angle of the fan-shaped solution domain. The outlier points (gray) lie outside 399 the solution domain, while the other data points (colored) lie within it or close to its bound-400 ary. The location of a point within the solution domain reflects the depth and mode of deformation under peak and retrograde conditions (α_p, α_r) . Points along the central line 402 $\alpha_p = \alpha_r$ have the same mode of deformation in the peak and retrograde stages. Points 403 below this line deform by flattening under peak conditions and by constriction under ret-404 rograde conditions, and points lying above the central line deform by constriction un-405 der peak conditions and by flattening under retrograde conditions. Samples from one oro-406 gen tend to span a large range of α_p, α_r that could reflect local differences in the mode 407 of deformation. 408

For a given depth, a range of P_p , P_r is possible depending on the value of α_r , α_p (see eqs. 21 and 22). The opposite is also true: for a given P_p , P_r , there is a range of possible depths. We represent the lower and upper estimates of this range in Figures 8A and 8B, respectively. In this model, all points lie below the "deepest Moho" reference depth for the lower depth estimate, and only one point is deeper than the "deepest Moho" when using the upper depth estimate.

415

3.4 Depth estimates using the two-point method

Figure 9 shows depth estimates for our data according to the horizontal stress changedriven model and the stress rotation model ("S-model", thin bars) and the compression to extension model of the previous section ("YB-model", thick bars). Depth estimates for peak pressure assuming a lithostatic stress state (see section 2.3) are also shown as the "L model" for reference (short bars).

In the following passage, we use the terms L-depth, S-depth and YB-depth to refer to the depth estimates according to the L-, S- and YB-models, respectively. The methods for computing the depth ranges for the S-model and YB-model are given in Appendix
A.

The minimum and maximum YB-depths are equal for points on the border of the 425 solution domain fan, while the range is largest for points along the central line (Fig. 8). 426 The range of S-depth tends to be larger for points with a low P_p/P_r ratio and decreases 427 with increasing P_p/P_r (because fewer solutions exist; see Figs. 6 and 7). For all samples 428 except the outliers, the S- and YB-depths are significantly lower than the L-depth. For 429 example, one point in Kokchetav and one point in Sulu-Dabie have L-depths > 160 km, 430 whereas their S- and YB-depths are 65-70 km and 60-85 km, respectively. For out-431 liers, the upper estimate of the S-depth is close to the L-depth. In the L-model, the depth 432 is proportional to the peak pressure. Thus, large differences in peak pressure between 433 two samples result in large differences in depth. However, the S- and YB-models take 434 both peak and retrograde pressures into account, which can smooth out this difference. 435 For example, the two data points with the highest pressures in the Alps have L-depths 436 of 95 and 125 km, whereas the maximum S-depth is 70 km for both. Conversely, points 437 with the same L-depth (i.e., same peak pressure) can have different YB- and S-depths. 438 This contrast is best exemplified by comparing points with the same P_p in Figure 8: points 439



Figure 9. Estimated depth of each sample using the two-point method of pressure-to-depth conversion. The graph also shows depth estimates using the one-point lithostatic case for reference. L-model: peak pressure in the one-point lithostatic case. S-model refers to the models described in sections 3.1 and 3.2. The YB-model refers to the model described in section 3.3. For the YB-model, filled rectangles indicate depth estimates for points that lie within the model boundaries, while open rectangles apply to points outside the model boundaries. The depth estimates indicated by open rectangles are relevant for points close to the model boundary (e.g., samples #1 and #14) but less relevant for points far from the boundary (i.e., category Others (outliers)). This figure can be reproduced using the computer script from supplementary information S9.

in the upper half of the fan $(\alpha_p > \alpha_r)$ align on a lower depth estimate contour, while 440 points in the lower half of the fan $(\alpha_p < \alpha_r)$ align on the upper depth estimate con-441 tour. Thus, at constant P_p , the mean depth estimate increases with increasing P_r , and 442 the uncertainty increases towards the center of the fan ($\alpha_p = \alpha_r$ line). For example, 443 for points at $P_p = 3GPa$, the mean depth estimate increases from 40 km at $P_r = 0.6GPa$ 444 to 60 km at $P_r = 1.2 GPa$. At these two extreme P_r values, the depth estimate has a 445 unique value, while at the center of the fan $(P_r = 0.85GPa)$, the depth estimate ranges 446 between 40 and 60 km. Overall, the most striking features of the S- and YB-models are 447 that all data points have at least part of their range shallower than the "deepest Moho" 448 line and that the deepest S-depth is approximately 90 km compared to 165 km for the 449 L-depth. 450

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$_{451}$ 4 Discussion

Pressure is a function of both depth and deviatoric stresses. However, since devi-452 atoric stress cannot be measured, pressure-to-depth conversions require assumptions. In 453 the previous sections, we propose several pressure-to-depth conversion methods involv-454 ing one or two pressure data points. In particular, we show that the proportionality be-455 tween P_p and P_r can be explained by a model where P_p and P_r are recorded by the rock 456 at the same depth but under different stress states (Figs. 6 to 8). For simplicity, we only 457 present two-point models with either stress rotation or horizontal stress magnitude change 458 and no exhumation. Combining rotation and magnitude change may further decrease 459 the magnitude of deviatoric stresses required to explain the data. Relaxing the assump-460 tion that $z_p = z_r$ and accounting for some exhumation would also decrease the mag-461 nitude of deviatoric stresses required. 462

463

4.1 Perspectives on using strain data

In our formulation of the pressure-to-depth conversion, we use α instead of a stress value (see eq. 4). α characterizes the shape of the stress ellipsoid and is thus similar to 465 commonly used parameters for characterizing the shape of ellipsoid such as Lode's ra-466 tio or Flinn's k-value (Mookerjee & Peek, 2014). Because strain results from applied stress, 467 obtaining a value for α using markers of deformation could provide key data to better 468 constrain depth. The two-point models relying on a change in the magnitude of hori-469 zontal stress (Fig. 6) or stress orientation (Fig. 7) give ambiguous results since both mod-470 els can explain the data. However, one could falsify the predictions of the model based 471 on stress rotation (Fig. 7) by using the directions of the strain ellipsoid or paleostress 472 inversion of fault orientations to estimate stress directions. 473

474

4.2 Data distribution and model

The data suggest that P_p and P_r are proportional (see Fig. 1). However, by us-475 ing the one-point method, because P_p and P_r are considered independently, it is diffi-476 cult to explain this proportionality. In the lithostatic case, for instance, the decompres-477 sion from P_p to P_r is controlled only by the exhumation of rocks. However, the currently 478 proposed exhumation mechanisms (e.g., subduction channels and corner flows) do not 479 suggest that exhumation would be proportional to maximum depth. On the other hand, 480 the two-point model treats both P_p and P_r together. Since we assume that $z_p = z_r$, 481 the maximum change from P_p to P_r is limited by Byerlee's law, and the yield stress func-482 tion is linearly dependent on P. Considering reasonable values for the friction coefficient 483 (e.g., 0.65), the limits of the model outline the distribution of the data. For example, for 484 the models shown in Figures 6J and 7H, the extent of the model domain outlines the up-485 per extent of the distribution, and the lower limit of the model corresponds to the lower 486 extent for outliers. 487

The YB-model simulates the case where rocks are brittle in both compression and extension and thus constitutes a particular case of the two-point model. It is interesting to note that although the YB-model allows us to largely explain the data, it excludes the outliers (Fig. 8). However, all data (including outliers) can be explained considering the more general S-models (Figs. 6 and 7).

The upper extent of the data distribution $(P_p/P_r \ 4.8)$ can only be explained when $\bar{\sigma}_x^p = \Phi, \alpha_p = 1$ (brittle constrictive deformation in compression) and $\bar{\sigma}_x^r = 1/\Phi, \alpha_r =$ 0 (brittle flattening deformation in extension). The lower extent of the data distribution excluding outliers $(P_p/P_r \ 2.4)$, however, can have several explanations. In the YB-model, it corresponds to $\alpha_p = 0, \alpha_r = 1$. For S-models (e.g., Figs. 6G, I, J; 7E, G, H), the lower bound of the data can be within the solution domain and coincides with different values of $\bar{\sigma}_x^r$ or θ . Interestingly, the lower limit coincides with Sxr=1 (i.e., lithostatic case) ⁵⁰⁰ in Fig. 6J. Two-point models can fit all data points from the dataset (or lie very close ⁵⁰¹ to the model boundary), which suggests that in all orogens, a change in stress state may ⁵⁰² be responsible for the decompression from P_p to P_r . The different predictions in terms ⁵⁰³ of the change mode of deformation (α_p to α_r) bring additional constraints concerning ⁵⁰⁴ the mechanism responsible for the change in the stress state. Monitoring the evolution ⁵⁰⁵ of α in 3D numerical geodynamic models may provide more answers.

4.3 Inclusion-host system

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The models presented here explain pressure variations in a homogeneous material 507 subjected to a change in depth or deviatoric stresses. In a heterogeneous system, the pres-508 sure in one material may be affected by deviatoric stresses in another material. A well-509 studied example is the case of an elliptical inclusion embedded in an elastic or linear vis-510 cous matrix. In this system, the magnitude and sense of deviatoric stresses are functions 511 of the relative strength between the matrix and the inclusion, as well as the orientation 512 of the inclusion in the stress field (D. W. Schmid & Podladchikov, 2003, 2005; Moulas 513 et al., 2014). The pressure in the inclusion is controlled both by the stress state in the 514 inclusion and by the stress state in the host rock. An important point is that deviatoric 515 stresses in a weak inclusion may be negligible, while the pressure can still be as high as 516 $\sigma_1 = \Phi$ in a strong host rock. This is very different from a homogeneous material where 517 the pressure is lithostatic in the absence of deviatoric stresses. Thus, pressure can vary 518 between the values of σ_3 and σ_1 for the strongest material in the inclusion/host system, 519 whereas in a homogeneous material, pressure can vary only between $2/3\sigma_3 + 1/3\sigma_1$ and 520 $1/3\sigma_3+2/3\sigma_1$ (see eq. (6)) (Moulas et al., 2014; Schmalholz & Podladchikov, 2013). Field 521 examples of this phenomenon have been documented by Luisier et al. (2019) in the Monte 522 Rosa nappe (Alps) and by Jamtveit et al. (2018) in the Bergen Arc (Caledonides). 523

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4.4 Local vs. regional stress state

In this paper, we present several methods that can be used to determine possible 525 stress states associated with peak and metamorphic pressures. Stress states are by essence 526 local. Some researchers even propose that metamorphic pressure may reflect the stress 527 state in only a single grain (see the discussion about inclusions in the previous paragraph) 528 and that large pressure gradients responsible for pressure differences on the order of GPa529 can be recorded within a single grain (Tajčmanová et al., 2014, 2015). In our dataset, 530 samples from the same region have a wide variety of P_p/P_r ratios (see Fig. 1C) and are 531 often distributed from one side of the fan to another (between $P_p/P_r = 1.4$ and $P_p/P_r =$ 532 4.8), which indicates differences in the change in stress magnitude, stress orientation or 533 relative magnitude of σ_2 (α). This could be an indication that pressure data reflect the 534 local (grain- to 10 km-scale) rather than regional (100 km) stress state. Thus, using the 535 stress states determined in this study to interpret regional-scale processes requires tak-536 ing some caution. For example, the YB-model assumes that the peak to retrograde pres-537 sure can be explained by a transition from a compressional to an extensional stress state, 538 with both stress states close to the brittle limit. These stress states reflect km-scale con-539 ditions or are smaller in the sense that the whole system is submitted to convergence (i.e., 540 in Fig. 1). The distance between two points far from the subduction zone, one located 541 on the subducting plate and the other located on the overriding plate, is constantly de-542 creasing, but the part undergoing exhumation is locally subjected to extension. This cor-543 responds well with the fact that the exhumation of a coherent metamorphic unit is im-544 possible without a normal fault on top. As we have shown, stress orientation has a strong 545 control on pressure changes, and in a complex orogen, stress orientations can vary sig-546 nificantly in space and time, e.g., due to changes in the subduction angle or the friction 547 along the plate boundary (Wang & Hu, 2006), the proximity to magma chambers (Gerbault 548 et al., 2018) or faults (e.g., Shao & Hou, 2019; Martínez-Díaz, 2002; Maerten et al., 2002), 549 or the position within the orogen (e.g., Kastrup et al., 2004). 550

4.5 Implications for geodynamic models

Rock strength strongly depends on temperature. Hence, considering classic rhe-552 ological yield stress envelopes (e.g., E. B. Burov, 2011), it seems inadequate to consider 553 large deviatoric stresses deep in the lithosphere (> 120 km) due to the temperature in-554 crease with depth. This statement could favor using lithostatic pressure-to-depth con-555 version but remains debatable. Indeed, the depth estimates using the S- and YB-models 556 are consistent with the depth of the crustal roots of orogens, and in these places (i.e., 557 at the base of the crust or in the lithospheric mantle), significant deviatoric stresses are 558 possible. Significant deviatoric stresses are even more likely at this depth in a subduc-559 tion zone with a cold geotherm. 560

Several elements suggest significant deviatoric stresses near the Moho depth: (1) earthquakes are not uncommon at such depths in a subduction context and provide evidence that brittle deformation can occur (e.g., B. Hacker et al., 2003; Hetényi et al., 2007), and (2) several field and petrological studies have already evidenced brittle deformation associated with HP metamorphism (e.g., Austrheim & Boundy, 1994; John & Schenk, 2006; Angiboust et al., 2012; Hertgen et al., 2017; Yang et al., 2014).

Samples with high P_p/P_r require a stress field close to the brittle limit using the S-model (e.g., Fig. 7). However, samples with $P_p/P_r < \Phi$ are consistent with a stress state where the magnitude of the deviatoric stress (second invariant) is only half that required for brittle deformation (i.e., $\bar{\sigma}_x^p = (1+\Phi)/2$) when peak pressure is recorded. This means that even in the ductile realm, the effect of the deviatoric stresses should not be neglected.

The release of fluids from dewatering metamorphic reactions can decrease the effective pressure. Thus, one might argue that the transition from P_p to P_r is caused by fluid pressure. However, this mechanism seems unlikely because fluid pressure would need to remain high during exhumation (otherwise, a new peak pressure would be recorded). Townend and Zoback (2000) argue that high fluid pressure leads to rock fracturing, which creates space and thus causes fluid pressure to decrease.

579 5 Conclusion

In this contribution, we reviewed the basic mathematical formulations of pressureto-depth conversion for a homogeneous rock. First, we derived the standard "one-point method of pressure-to-depth conversion" and applied it to a large dataset of metamorphic pressures to independently estimate a range of depths at which rocks may have recoded their peak (P_p) and retrograde pressures (P_r) . Since the most common assumption in the literature is to consider that metamorphic pressure corresponds to the lithostatic pressure, we used this "lithostatic case" as a reference.

By introducing deviatoric stress components and considering only the compressional 587 stress regime (σ_1 horizontal) at P_p and both compressional and extensional (σ_1 verti-588 cal) stress regimes for P_r , we showed that the deviations from the reference case can be 589 significant. For P_p , the estimated depths vary between 40 and 100 % of the reference case. 590 For P_r , the estimated depth range is 40–185% of the reference case. Thus, under our 591 assumption, the lithostatic case represents an upper bound estimate of depth for P_p and 592 an intermediate value for P_r . Moreover, the uncertainty ranges of both peak (z_p) and 593 retrograde (z_r) depths are large enough to lead to overlap for these two depth estimates. 594 This means that the transition from P_p to P_r can be triggered by exhumation, a change 595 in the stress state at constant depth, or a combination of both processes. 596

Second, we presented "two-point methods of pressure-to-depth conversion" that use both P_p and P_r to estimate depth under the hypothesis that $z_p = z_r$. For the twopoint method, we considered two mechanisms of stress change between P_p and P_r : (1) change in the magnitude of horizontal stresses and (2) rotation of the stress state. We also treated a particular case where the magnitude of deviatoric stresses is maximum, and the stress regime varies from compression at P_p to extension at P_r . The two-point method greatly decreases the uncertainty range of depth estimates and yields stricter constraints on the possible stress state. Remarkably, all P_p , P_r points in our dataset are consistent with a change in the stress state at a constant depth.

In our dataset, the maximum depth estimates under the "lithostatic assumption" 606 are approximately 160 km for P_p and 50 km for P_r . Thus, the lithostatic assumption re-607 quires deep burial and exhumation from great depth. On the other hand, the two-point 608 models reveal that points in our dataset are consistent with depths shallower than 75 609 km (i.e., the current deepest Moho). This suggests instead that all metamorphic rocks 610 in our dataset have been buried at crustal depths with no (or only minor) exhumation 611 between P_p and P_r . The validity of either of these models cannot be assessed based only 612 on pressure and temperature data. However, the principal stress directions and the rel-613 ative magnitude of σ_2 (i.e., α) may be estimated from the strain ellipsoid or paleostress 614 analysis. Thus, a precise analysis of the deformation in association with the P estimates 615 in metamorphic rocks could validate or falsify depth estimates from the two-point model 616 and further decrease the depth estimate uncertainty. 617

⁶¹⁸ Appendix A Depth estimates for the two-point model

A1 S-model

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The depth estimate range for S-models is calculated numerically by testing a large array of combinations of $\bar{\sigma}_x^p$, α_p and α_r for each sample. The ranges considered are $1 \leq \bar{\sigma}_x^p \leq \Phi$ and $0 \leq \alpha_p, \alpha_r \leq 1$, and we use 50 values to discretize the range of each parameter for a total of $50^3 = 125,000$ parameter combinations. We proceed in two steps. First, we compute σ_z using eq. (15) with $P = P_p$, $\bar{\sigma}_x = \bar{\sigma}_x^p$, and $\alpha = \alpha_p$, and we compute $z = \sigma_z/\rho g$. Second, we need to test whether the previous solution is within the acceptable bounds of the model (i.e., not in the gray area of Figs. 6 to 8). For this purpose, we compute $\bar{\sigma}_x^r$ using the following equation:

$$\begin{cases} \bar{\sigma}_x = \frac{3P/\sigma_z - 2 + \alpha}{1 + \alpha}, \text{ when } \bar{\sigma}_x \le 0, \\ \bar{\sigma}_x = \frac{3P/\sigma_z - 1 - \alpha}{2 - \alpha}, \text{ when } \bar{\sigma}_x \ge 0, \end{cases}$$
(A1)

with $P = P_r$, $\bar{\sigma}_x = \bar{\sigma}_x^r$, and $\alpha = \alpha_r$. Then, we test whether $1/\Phi \leq \bar{\sigma}_x^r \leq \Phi$ and update the range of depth if the test is successful.

A2 YB-model

To compute the range of depth for the YB-model, we use the minimum and up-623 per estimates of depth whose contours are plotted in Figures 8A and 8B, respectively. 624 In practice, we compute σ_z using eq. (15) with parameters $[P, \alpha, \bar{\sigma}_x]$. For data points 625 where $P_p/P_r > \Phi$ (i.e., above the line marked $\alpha_p = \alpha_r$ in Fig. 8), we use parameters 626 $[P_p, 1, \Phi]$ to compute $min(\sigma_z)$, and $[P_r, 0, 1/\Phi]$ for $max(\sigma_z)$. For data points where $P_p/P_r \leq$ 627 Φ , we use $[P_r, 1, 1/\Phi]$ for $min(\sigma_z)$ and $[P_p, 0, \Phi]$ for $max(\sigma_z)$. Then, we compute z =628 $\sigma_z/\rho g$. In this algorithm, depth is calculated using either $[P_p, \alpha_p, \bar{\sigma}_x^p]$ or $[P_r, \alpha_r, \bar{\sigma}_x^r]$. If 629 α_p is used as input, α_r can be computed back from σ_z , and we can perform the test $0 \leq \alpha_p$ 630 $\alpha_r \leq 1$ to verify that the solution is within the bounds of the model. If α_r is used as 631 input, α_p is computed instead. If the test is successful, we plot the range as a colored 632 box in Figure 9 or as an open box otherwise. 633

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