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Metamorphism and linked deformation in understanding tectonic processes at varied scales

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Geodynamics of Continents and Oceans – A tribute to Jean Aubouin / *Géodynamique des continents et des océans – Hommage à Jean Aubouin*

Metamorphism and linked deformation in understanding tectonic processes at varied scales

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Abstract. This contribution presents a review exploring some aspects and issues surrounding the links between metamorphism and deformation at different scales. I first discuss the quantification methods of thermodynamics and the parameters able to overcome the kinetic barriers for metamorphic reactions. On the basis of some world's type iconic examples I discuss how metamorphism is likely to portray the thermo-mechanical evolution of lithosphere active zones, thus large-scale tectonic processes. Finally, I present the multiple interactions between metamorphism and deformation at rock and outcrop scales.

Keywords. Metamorphism, Ductile deformation, Tectonic processes, Subduction zones, Collision belts.

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1. Introduction

Metamorphic rocks have long held interest for geologists owing to the specificity of their mineral assemblages in relation with the diversity of mountain belts in which they outcrop [e.g. Barrow, 1893, Eskola, 1915, Franchi, 1900, Suess, 1875]. Since the second half of the twentieth century there is a broad consensus that metamorphism is the dominant transformation mechanism in the lithosphere. Thus, the link between type of metamorphism and tectonic environment is widely recognized since the seminal publication of Miyashiro [1961]. Metamorphism generally occurs when a rock is subjected to conditions under which, for thermodynamic reasons, its mineral composition is no longer stable. But the presence of fluids and the mechanisms of inter- or intracrystalline

deformation control the velocities of metamorphic reactions. It is therefore important to consider the quality and uncertainties of available methods for quantifying intensive parameters but also the age, the duration and the mechanisms of metamorphic transformations.

The goal of this review is, after a discussion of the methods of quantification of thermodynamic parameters and of the processes that help overcoming kinetic barriers for reactions, to present the state of knowledge on the links between metamorphism and large-scale geodynamic contexts. Finally the interaction between ductile deformation and metamorphism at the scale of the rock or outcrop will be presented.

2. Metamorphic pressure (P)–temperature (T) estimates and significance

2.1. Quantification of metamorphic P – T conditions

One can derive the P – T conditions experienced by metamorphic rocks by comparing their mineral associations with experimental and thermodynamic data. During the twentieth century numerous experimental data were produced, among which some have become references for the stability of metamorphic minerals (i.e. glaucophane stability field, aluminium-silicates triple point, ...). Moreover, since the late 1800's and the application of the phase rule in heterogeneous systems [Gibbs, 1875–1876], the conceptual framework of equilibrium thermodynamics has provided significant information on the formation of metamorphic rocks [Powell, 1978]. This concept led first to the topologic analysis of metamorphic reactions, i.e. the number and geometric arrangement in the P – T space of the reaction curves and most stable mineral associations [Zen, 1966]. Equilibrium thermodynamics led also, during the second half of the twentieth century, to the development of geo-thermometers and geo-barometers rooted in the thermodynamics of solid solutions and cationic exchanges and/or mutual solubility between coexisting minerals [see Krogh Ravna and Paquin, 2003, Spear, 1993, for reviews]. Applications of metallurgical concepts were at the origin of thermo-barometers based on disorder/order phase transitions [Carpenter, 1981] or discontinuous precipitations [Joanny et al., 1991]. Decisive progress in thermodynamic analysis has been made since the late 1980s with (i) the production of internally consistent thermodynamic databases [Berman, 1988, Holland and Powell, 2004, 2011], (ii) the formulations of mixing properties and activity-composition relationships for mineral solid-solutions and melts [Green et al., 2016, White et al., 2014] and (iii) the development of softwares using these multiple databases to model the mineralogical evolution of a rock from its chemical composition. This led to the development of modelled phase diagrams (often called “pseudosections”), as THERMOCALC [Powell et al., 1998] PERPLE_X [Connolly, 2005], THERIAK-DOMINO [De Capitani and Petrakakis, 2010], BINGO-ANTIDOTE [Duesterhoeft and Lanari, 2020] and GeoPS [Xiang and Connolly, 2022] are currently used. Pseudosections allow the

effects of major and minor elements, of the Fe-oxidation state and/or of the amount and nature of fluids on P – T stability fields of mineral assemblages to be explored. These models allow also the mode and the composition of minerals to be determined for a given rock composition under specific P – T conditions [Lanari and Duesterhoeft, 2019, Yakymchuk, 2017]. Furthermore, since the last two decades, thermometers based on trace elements solubility in minerals are available [Thomas et al., 2010, Tomkins et al., 2007]. Last, using trapped mineral inclusions, spectroscopic thermometers and barometers, along with laser Raman micro-spectroscopy, and called “Thermoba-Raman-try”, have been calibrated [Angel et al., 2017, Kohn, 2014].

2.2. Uncertainties of P – T quantifications

Experiments produced to calibrate the stability of a given mineral assemblage or the equilibrium curve of a given reaction should be well reversed and not rely on synthesis runs only. Many experiments were obtained at temperatures higher than 800 °C and many were not compositionally reversed under conditions similar to those under which natural rocks are metamorphosed. When undertaken on natural rock compositions a restricted number of mineral phases is generally observed in the produced runs. This discrepancy with respect to thermodynamic rules makes their direct application to natural assemblages always subject to discussion [Johnson et al., 2008].

With the exception of methods based on single element solubility within a single mineral, the basic prerequisite of all kinds of thermodynamic analyses is the achievement of chemical equilibrium in the investigated system. But metamorphic rocks commonly show textural and/or chemical evidences of disequilibrium such as corona-like textures with preservation of relicts, strongly zoned minerals, or preserved metastable phases. These various petrographic observations highlight the role of kinetics on metamorphic reactions [Lasaga, 1998]. Defining the scales from which the chemical equilibrium is reached is thus of crucial importance [Lanari and Engi, 2017]. The definition of the possibly equilibrated volume requires careful petrographic analysis of the textures of metamorphic rocks in thin sections, coupled with detailed chemical analysis of the phases observed before any attempt at

thermodynamic quantification. Mineral geothermobarometry imposes the equilibration of several coexisting phases as well as high-resolution data on activity-composition relationships for mineral solid solutions. The accuracy of the obtained results may be also affected by the difficulty of measuring the amount of Fe^{3+} in minerals and by compositional re-equilibration during retrograde evolution. The results extracted from thermodynamic modelling methods are at the first order dependent of the selected chemistry of the considered system. Their accuracy requires thus to determine the effective (or reactive) rock composition which is still challenging in retrogressed or polyphased metamorphic rocks [Lanari and Engi, 2017].

At present the most robust strategy could be to cross the results obtained by thermodynamic modelling (including variations in chemical compositions of minerals, i.e. isopleths) with P - T conditions extracted from trace elements solubility in specific minerals (quartz, zircon, titanite...) and from geothermobarometry of coexisting phases in textural equilibrium. The rise of petrochronology [Engi et al., 2017] and the recent developments of dating in-situ techniques [e.g. Tual et al., 2022] allow precise time constraints to be obtained. This has led, over the past 20 years, to improve the accuracy of P - T - t paths and thus the quantification of finite metamorphic field gradients.

The evolution of thermodynamic data and methods has led to a significant revision of the P - T fields of the classical metamorphic facies [Spear, 1993] with the definition of ultra high-pressure metamorphism (Figure 1A), thus a specific coesite and/or diamond bearing dry eclogite facies [Ernst and Liou, 2008]. This extension of metamorphic facies in the P - T space is, since the discovery of coesite-bearing assemblages in Alpine and Caledonian belts [Chopin, 1984, Smith, 1984], in conformity with the discoveries of numerous minerals or exsolution textures typical of ultra high-pressure conditions (micro-diamonds, Zr-ellenbergerite, majorite, stishovite, ... see Schertl and O'Brien, 2013 for an extensive review).

2.3. Geological significance of calculated metamorphic T and P values

The temperature distribution in the lithosphere depends on possible heat sources and efficiency

of heat transfer by conduction or advection. The main heat source is the radioactive elements decay, even if localized, generally minors, mechanical ("shear or frictional heating") or chemical (latent heat of reaction or crystallization) should also be taken into account. This global heat production is dissipated by conduction within the lithosphere. However, in some geological contexts, an active transport of heat by advection is added to the system by magmatic intrusions, hydrothermal fluids or deep mantle diapirs, while subduction zones are largely cooled by downgoing oceanic lithosphere. Thus, understanding the temperature distribution with depth requires a deep geological knowledge of the area of interest. In most of the studies, a conductive geothermal gradient serves as a reference for interpreting metamorphic temperatures.

Pressure increases with depth due to the force exerted by the mass of rocks overlying a given point of interest (i.e. lithostatic pressure). The relationship between depth and pressure is thus given by $P = \rho \cdot g \cdot z$ (with ρ the density, g the acceleration due to gravity, i.e. 9.81 ms^{-2} , and z the depth). In the Earth, pressure is thus considered to be hydrostatic. This classical relationship must be in some cases questioned due to the accumulation of deviatoric, thus nonhydrostatic, stress during deformation ["tectonic overpressure" Mancktelow, 2008, Moulas et al., 2019]. In such a case, the depth can be over-estimated. However both numerical thermo-mechanical modelling on convergent zones and non-equilibrium thermodynamics applied to deformed metamorphic rocks highlight that magnitude of pressure deviations are on the order of 10–20% from the lithostatic values [Hobbs et al., 2010, Li et al., 2010, Marques et al., 2018], thus of the same order of error margins bars of calculated metamorphic pressures. At shallow crustal levels (the first few kilometres) where fluids are abundant, both hydrostatic and lithostatic pressure must be considered as physical components of the "effective pressure" [Jaeger and Cook, 1979, Moulas et al., 2019]. In the first 2~3 kilometres of continental crust, if fluids are H_2O rich, the hydrostatic gradient must be 10 MPa/km and thus, with a density of upper crustal rocks of 2500 kg/m^3 , $P_{\text{fluid}} \approx 1/2.5 P_{\text{lithostatic}}$. In this case, the calculated depth must be corrected by this factor.

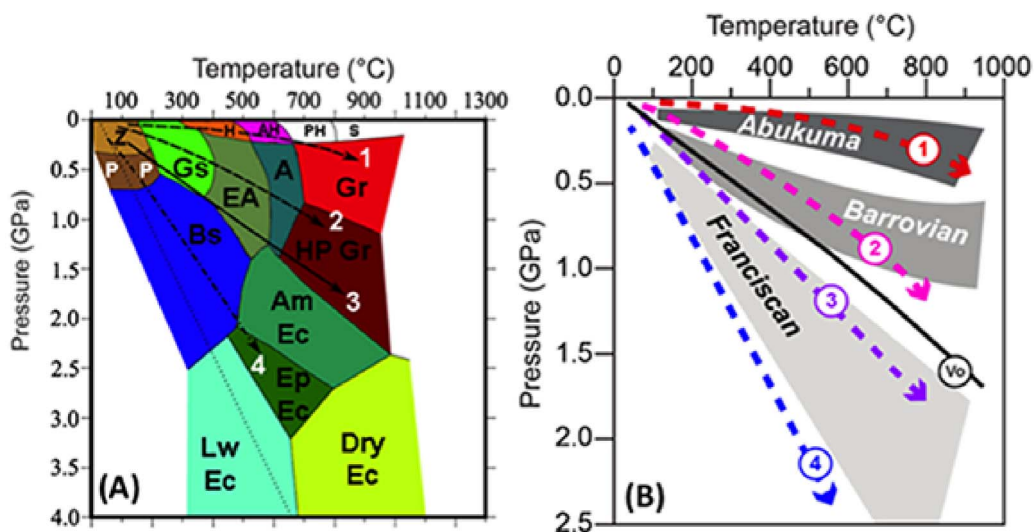


Figure 1. (A) Metamorphic facies and geothermal gradients (modified after Regorda et al., 2021; Z: zeolite; PP: prehnite-pumpellyite; Gs: greenschist; EA: epidote amphibolite; A: amphibolite; Gr: granulite; HPGr: high-pressure granulite; Bs: blueschist; AmEc: amphibole-eclogite; ApEc: epidote-eclogite; LwEc: lawsonite-eclogite; DryEc: dry-eclogite; S: sanidinite; PH: pyroxene-hornfels, AH: amphibole-hornfels; H: epidote-hornfels; 1: arc regions; 2: collision zones; 3: warm subductions; 4: cold subductions). (B) Typical metamorphic field gradients. 1: magmatic arcs and ridges; 2: collision zones; 3: warm subductions; 4: cold subductions [after Cloos, 1993]. Black line: standard geothermal gradient.

3. Metamorphism and large-scale geodynamic contexts

Since Miyashiro [1961] and Ernst [1971a,b], it is widely recognized that specific metamorphic facies series (i.e. observed metamorphic field gradients) portray the thermo-mechanical evolutions of lithosphere active zones, as a function of tectonic environments (Figure 1).

Franciscan, Barrovian and Abukuma type series reflect the observed metamorphic field gradients, and therefore the finite metamorphic architecture of an orogen. But in each area of an observed finite metamorphic zonation, rocks were more or less buried and in all cases exhumed through time, thus their tectonic history can only be understood by the construction of quantitative P - T - t paths [England and Richardson, 1977, Spear and Peacock, 1989, Thompson and England, 1984, Figure 2], or by the production of time-dependant metamorphic maps [Lardeaux, 2014a,b]. These latter approaches allow identification of the progressive metamorphic evolu-

tion through time and space of a given metamorphic unit.

3.1. Subduction zones

3.1.1. Metamorphism of the oceanic lower plate

It is primarily in subduction zones that the link between metamorphism and large-scale geodynamics was established. Along the Californian active margin the occurrence of eclogite pods and garnet-glaucophane-bearing schists led to the recognition of the “iconic” California Mesozoic-Cenozoic HP-LT accretionary complex [Coleman and Lanphere, 1971, Ernst, 1971a]. Eclogites and blueschist-facies rocks are embodied within serpentinites and mud-rich meta-sediments within the so-called Franciscan mélangé [Cloos, 1982, Ernst, 2015, Wakabayashi, 1990]. Along the Pacific Ocean, this Franciscan complex is located west of the Great Valley Group, which is a typical forearc basin [Ernst, 2011, 2015, Orme and Surpless, 2019]. Eclogites and blueschists represent fragments of oceanic crust subducted and

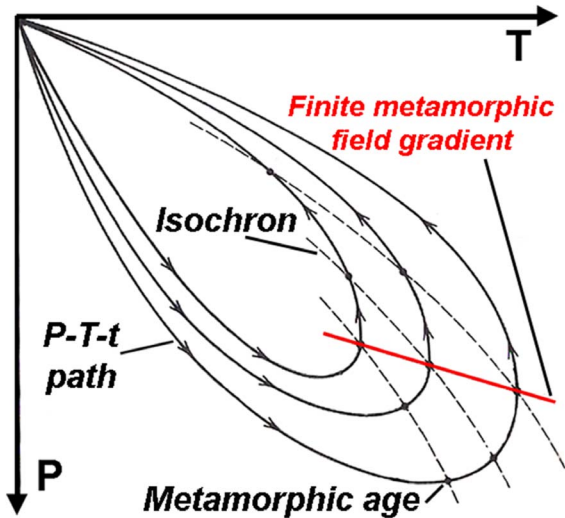


Figure 2. Theoretical relation between finite field gradient and P - T - t paths.

exhumed by return flow within subduction channels, composed of low-density meta-sediments or serpentinites, during long-lasting oceanic subduction [Cloos, 1986, Cloos and Shreve, 1988, Ernst and Liou, 2008, Grigull et al., 2012, Wakabayashi, 1992, Wassmann and Stöckhert, 2012]. A similar HP-LT metamorphic accretionary complex is recognized since several decades in Japan [Ernst, 1971b, Ishiwatari and Tsujimori, 2003, Maruyama et al., 1996, Miyashiro, 1961]. These circum-pacific orogens are at the origin of the world's type model of HP-LT metamorphism of the subducted oceanic lower plate (Figure 3).

A remarkable example of the link between formation and exhumation of HP-LT rocks and long-term intra-oceanic subduction is offered by the Caribbean example, with the distribution of eclogites and blueschists along both northern and southern Greater Antilles Arcs (Figure 4). The Caribbean present-day configuration is the result of the eastward progressive Pacific plate moving between North and South America from Late Jurassic until its current position indicated by the active Lesser Antilles arc [Kerr et al., 2003, Pindell and Kennan, 2009]. This displacement was accompanied by a reversal of the subduction polarity, relative to the initial and abandoned trench, at ca. 120 Ma and by

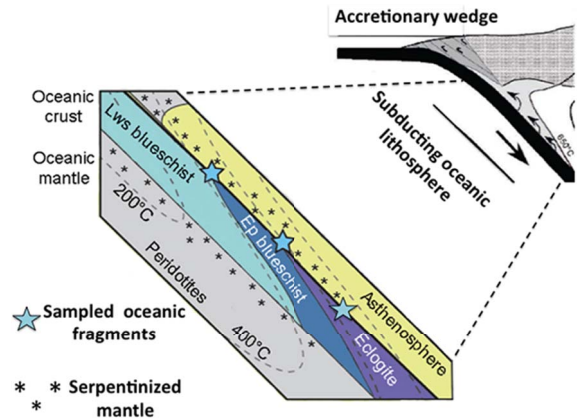


Figure 3. HP-LT metamorphic facies in subducted oceanic plate.

regional-scale transpression. In this context, HP/LT metamorphism developed during Cretaceous and eclogites and blueschists were exhumed under low-temperature conditions [García-Casco et al., 2006, Hu et al., 2022, Stöckhert et al., 1995, Tsujimori et al., 2006, West et al., 2014].

Exhumation took place during on-going subduction by mantle corner flow in serpentinized subduction channels [Blanco-Quintero et al., 2011, Guillot et al., 2009, Górczyk et al., 2007, Maresch and Gerya, 2005]. If quartz-bearing eclogites and blueschists are dominant, UHP conditions are also reported in Guatemala [Martens et al., 2017, Tsujimori et al., 2006, and Figure 5]. The “classical” P - T domain of the “Franciscan” metamorphic gradient must be widened to both high pressures and low temperatures. Thus, the lawsonite-bearing eclogites from Guatemala are close to the famous “forbidden zone” of Liou et al. [2000], i.e. a geotherm of ~ 5 °C·km⁻¹. The division proposed by Cloos [1993] deserves to be questioned, as in a given subduction context, both hot and cold metamorphic paleo-gradients can develop simultaneously. Thereby, the data obtained in eastern Cuba and Guatemala show that within the same tectonic unit, eclogites outcropping close to each other recorded contrasting P - T conditions.

This dataset is in line with the thermo-mechanical models of subduction channels. Indeed, since Shreve and Cloos [1986] and Cloos and Shreve [1988], numerous models of a specific layer of low viscosity rocks dragged by the subducted plate beneath the upper plate have been developed [Gerya et al.,

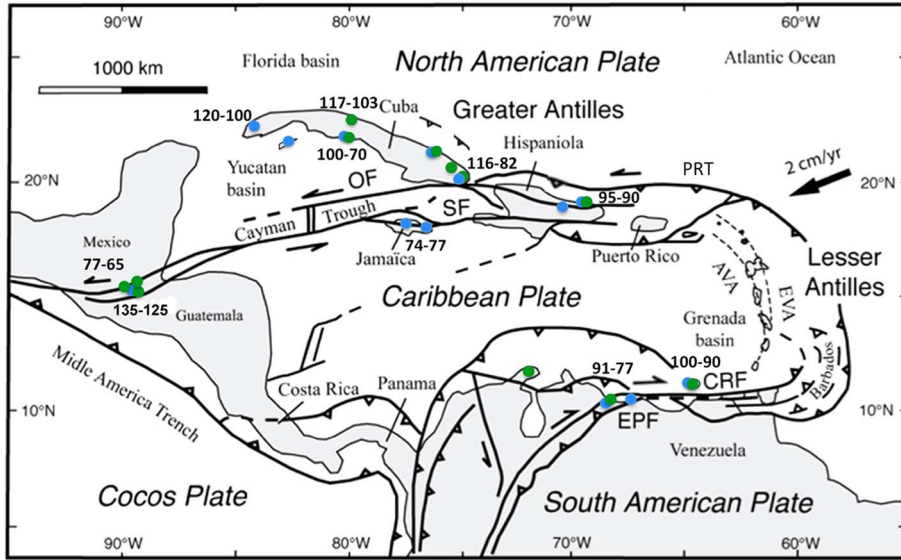


Figure 4. Location and ages of eclogites (green dots) and blueschists (blue dots) around the Caribbean plate. EVA: extinct volcanic arc; AVA: active volcanic arc; PRT: Puerto-Rico trench. EPF, El Pilar fault; CRF, central range fault; OF, oriental fault; SF, septentrional fault.

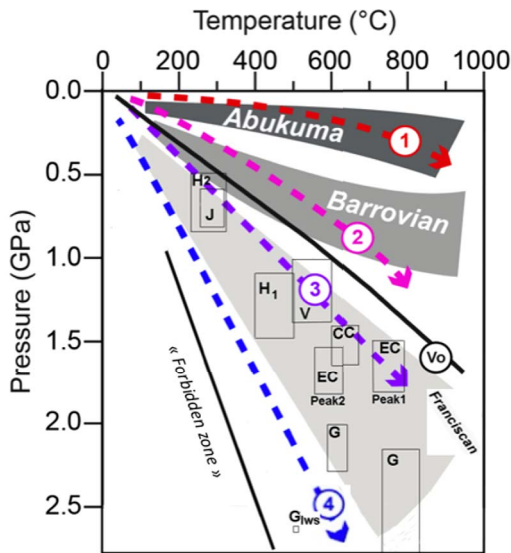


Figure 5. *P-T* conditions of Caribbean HP-LT metamorphic rocks. Legend same as Figure 1B. G: Guatemala eclogites; Glws: Guatemala lawsonite-eclogites; EC: Eastern Cuba eclogites; CC: Central Cuba eclogites; V: Venezuela eclogites; H1: Hispaniola eclogites; H2: Hispaniola blueschists; J: Jamaica blueschists.

2002, Marotta and Spalla, 2007, Roda et al., 2012, Van Hunen and Allen, 2011]. These models, combining a linear viscous rheology for the sub-lithospheric mantle and/or accretion prism with a linear viscoplastic rheology for the lithosphere, provide for short wavelength convective flow in the subduction channel (downward and upward displacements of subducted rocks) and thus simultaneously developed contrasted *P-T* conditions [Angiboust et al., 2013, Hebert et al., 2009, Maresch and Gerya, 2005, Meda et al., 2010, Regorda et al., 2021]. This probably reflects the versatility of subduction zones gradients with respect to a dogmatic vision of “steady-state” systems.

3.1.2. *Metamorphism in the upper plate (magmatic/volcanic arcs)*

Still in subduction zones the concept of paired metamorphic belts was developed [Miyashiro, 1961]. This model, revisited and extended by Brown [2010], predicts the development of regional-scale high heat flow in the overriding plate, coupled with HP-LT conditions in the subducted plate, leading to the development of intermediate-pressure granulite facies in the lower crust up to zeolite facies at uppermost crustal levels (Abukuma type metamorphism). Medium to

low P /high T series typically develop within the upper plate depending on subduction velocity, age of subducted plate and amount of magmas produced in the magmatic arc.

Extinct and deeply exhumed continental magmatic arcs exhibit regional scale amphibolite to granulite facies conditions and extensive partial melting [Ducea et al., 2015]. The Coast Mountains in British Columbia or the southern Sierra Nevada in California document arc crustal sections exhumed from paleodepths of 5 to 35 km [Chapman et al., 2014, Ducea et al., 2015]. In both cases mafic magmatic underplating is described and arc-related metamorphic conditions evolve from prehnite-pumpellyite or greenschist facies up to medium to high-pressure granulite facies from the shallowest to the deepest exposed levels [Day and Springer, 2005, Girardi et al., 2012]. High- P granulites are also reported from the extinct Arthur River continental arc complex in New Zealand [De Paoli et al., 2009, Flowers et al., 2005]. At shallow levels of active continental magmatic arcs, as exemplified by the Andean Cordillera, low-grade metamorphism (i.e. zeolite and/or prehnite-pumpellyite facies up to greenschist facies) is well documented [Aguirre et al., 2000, Muñoz et al., 2010]. Because this metamorphism affects sediments and interlayered volcanic rocks, it is frequently mentioned as “burial” metamorphism, which requires a temporal evolution from extensional versus compressional tectonic regimes. In any cases, high thermal gradients (i.e. >40 °C/km) and extensive hydrothermal fluids circulation are reported to produce the observed metamorphic mineralogy. All these characteristics must be combined to propose a schematic vertical crustal section showing the typical metamorphic pattern of continental magmatic arcs (Figure 6).

Exhumed fossil oceanic magmatic arcs display similar metamorphic patterns, as for example the fossil (i.e. obducted) Kohistan oceanic magmatic arc in Pakistan where abundant mafic magmatic underplating together with amphibolite and high- P granulite facies are reported [Burg, 2011].

Arc-related metamorphism has been recently discovered at shallow crustal levels of the active volcanic island arc from the Guadeloupe archipelago [Lesser Antilles arc, Favier et al., 2021, Vérati et al., 2018, Figure 4]. The 2 cm/yr, southwest oriented, subduction of the American plate beneath the Caribbean one led to development of this oceanic

volcanic arc [López et al., 2006]. The arc is divided in two ridges with an eastern Mid-Eocene ancient arc and a still active, since 5 Ma, western arc [Samper et al., 2007]. The conductive geothermal gradient was measured in between 69.3 ± 1.5 and 98.2 ± 8.8 °C/km around the Guadeloupe archipelago [Manga et al., 2012]. In Guadeloupe, the active arc includes from north to south different volcanic complexes [Ricci et al., 2015, Samper et al., 2007, Figure 7] and arc-related metamorphism, associated with ductile deformation, was identified in the oldest, thus most eroded and exhumed, 4.28 to 2.67 Ma aged Basal Complex [Favier et al., 2019].

The volcanic rocks are affected by young (<3 Ma) greenschist to sub-greenschist (i.e. prehnite-bearing) facies metamorphism (0.6–2 kbar for 250–300 °C), and a re-equilibration under zeolite facies conditions, consistent with the elevated geothermal gradients measured regionally for the arc axis [Favier et al., 2019]. In the Basal Complex, the ductile finite strain pattern is defined by localized, pressure-solution induced, schistose domains surrounding pods of undeformed rocks. A conceptual model for this arc-related hydrothermal tectono-metamorphic pattern is presented in Figure 8.

3.2. Continental collision zones

3.2.1. Continental collision zones with preserved subduction metamorphism pattern

The Alpine belt is the result of the Tertiary continental collision between the Adriatic promontory of the African plate and the thinned continental margin of the European-Iberian plate [Handy et al., 2010, Schmid et al., 2004]. This collision followed the closure, by subduction, of the Ligure–Piemontese Ocean started during Upper Cretaceous times [Agard, 2021, Agard and Handy, 2021, Dal Piaz et al., 2003, Rosenbaum and Lister, 2005, Rubatto et al., 1998, Spalla et al., 2010].

In this tectonic framework, the western Alps are a remarkable example of collision chain having preserved a subduction-related metamorphic signature. Significant occurrences of low-temperature eclogites and blueschists are reported since the end of the nineteenth century [see Lardeaux, 2014a, for review]. Ernst [1971b] was the first to identify a typical,

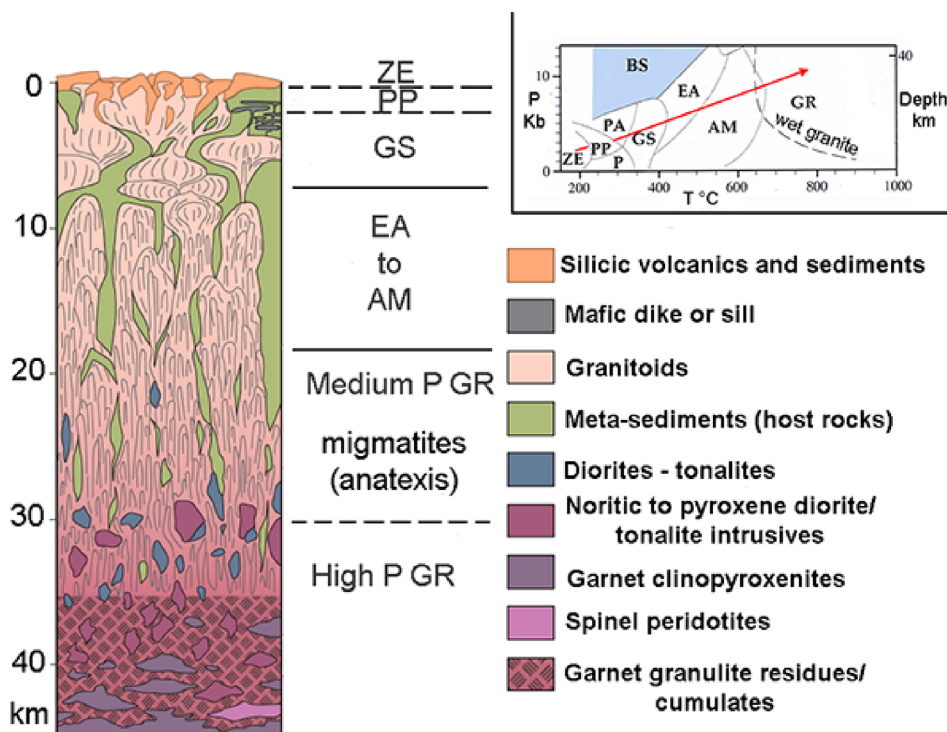


Figure 6. Vertical metamorphic pattern in continental arcs [modified after Ducea et al., 2015]. Inset shows the related metamorphic field gradient (red arrow). Metamorphic facies after Ernst and Liou [2008]. ZE: Zeolite facies; PP: Prehnite-Pumpellyite facies; GS: Greenschist facies; EA: Epidote-Amphibolite facies; AM: Amphibolite facies; GR: Granulite facies; BS: Blueschist facies; P: Prehnite facies; PA: Prehnite-Actinolite facies;

east-oriented, Franciscan metamorphic field gradient. Four decades of detailed petrologic investigations led to the production of synthetic metamorphic maps highlighting this subduction related metamorphic gradient in the internal western Alps [i.e. east of the Penninic front, Bousquet et al., 2008, Goffé et al., 2004, Lardeaux, 2014a, Oberhänsli et al., 2004, Figure 9].

From a geodynamic and/or tectonic point of view, the metamorphic data available today for internal western Alps highlight the following fundamental points:

- The finite metamorphic field gradient is in the range of 5–10 °C/km, thus similar to the previously presented circumpacific ones, compatible with long-lasting oceanic subduction (ca. 55 Ma) between Upper Cretaceous [Handy et al., 2010, Rebay et al., 2018] and subsequent Late-Eocene/

Oligocene continental collision [Belhassen et al., 2014].

- The finite metamorphic field gradient portrays the Alpine subduction polarity. However, ages of HP and/or UHP metamorphism vary from top to bottom of the Alpine tectonic pile. Indeed, eclogite facies metamorphism is Late Cretaceous—Paleocene (85–65 Ma) in the Austro-Alpine Sesia-Lanzo Zone, Early to Middle Eocene (60–45 Ma) in the oceanic units and late Eocene (40–35 Ma) in the European plate-derived Dora Maira Massif [see Bousquet et al., 2008, Lardeaux, 2014a, Manzotti et al., 2022, with references therein].
- Significant metamorphic heterogeneities (i.e. P and T gaps) can be seen even within a single tectono-metamorphic unit. Examples are reported in the Dora Maira unit

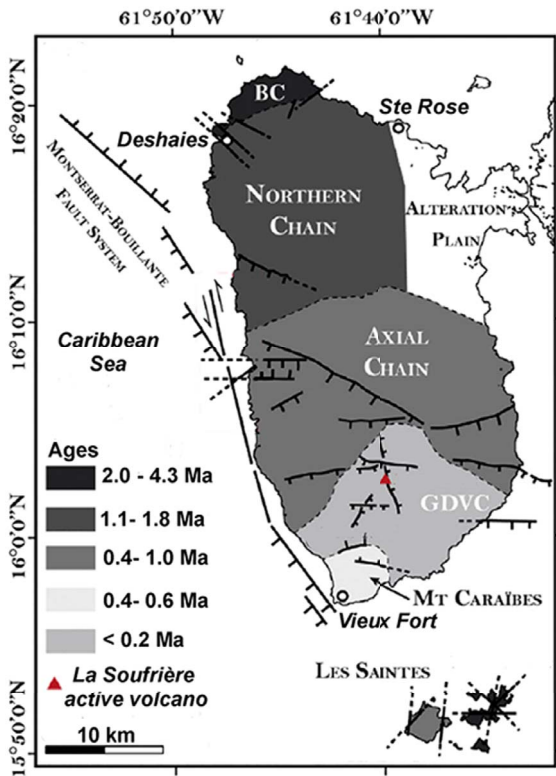


Figure 7. Basse-Terre (Guadeloupe) volcanic complexes and main faults [modified after Samper et al., 2007, Favier et al., 2019]. BC: Basal Complex, GDVC: Grande Découverte Volcanic Complex.

[Groppo et al., 2019, Manzotti et al., 2022], in the “Schistes Lustrés” unit located west of the Gran-Paradiso Massif [Bousquet, 2008], or in the Sesia-Lanzo Zone [Giuntoli et al., 2018, Roda et al., 2020, 2021, Zucali and Spalla, 2011]. In Zermatt zone or Monviso meta-ophiolites these heterogeneities are still a matter of debate. Some authors recognized contrasted P - T conditions [Luoni et al., 2019, Rebay et al., 2012, 2018], while Angiboust et al. [2009, 2012] proposed quite homogeneous peak conditions for these units. The size of the units, which have undergone a homogeneous evolution, differ widely according to the authors, resulting in contrasted exhumation models. Some authors consider syn-subduction exhumation of Alpine HP-LT rocks in accretionary

wedge and subduction channel [Allemand and Lardeaux, 1997, Bousquet, 2008, Gilio et al., 2019, Guillot et al., 2009, Polino et al., 1990, Roda et al., 2012, Spalla et al., 1996, Stöckhert and Gerya, 2005]. On the other hand, Agard et al. [2009] and Angiboust et al. [2012] consider a synchronous exhumation of coherent slivers along the entire belt.

- Since the discovery of coesite in the Dora-Maira unit [Chopin, 1984], the identification of UHP conditions in the Western Alps is progressively increasing. UHP or nearly UHP conditions are now documented in continental meta-sediments and meta-granitoids [Chopin et al., 1991, Kienast et al., 1991, Groppo et al., 2019, Manzotti et al., 2022, Chen et al., 2023], in oceanic meta-sediments [Reinecke, 1991] in oceanic meta-basalts and meta-gabbros [Angiboust et al., 2012, Bucher et al., 2005, Ghignone et al., 2022, Luoni et al., 2019] as well as in oceanic mantle serpentinites [Rebay et al., 2012, 2018].
- Slices of both oceanic and continental crusts were subducted during Alpine orogeny. It was in the western Alps that the subduction of the continental crust was for the first time depicted since the early 1970s. It is in the Austro-Alpine units that subduction of the upper continental crust [Compagnoni and Maffeo, 1973, Compagnoni et al., 1977, Dal Piaz et al., 1972] and the lower continental crust [Dal Piaz et al., 1983, Lardeaux et al., 1982, Lardeaux and Spalla, 1991] was identified. HP to UHP metamorphic conditions were depicted later in all the Internal Crystalline Massifs. These Internal Crystalline Massifs are attributed to the distal part of the extended Briançonnais palaeo-margin [i.e. European lower plate, Handy et al., 2010, Schmid et al., 2004] and are presently located beneath the oceanic units. On the other hand, the Austro-Alpine units derived from the Adria margin (i.e. promontory of the African upper plate) and are thrust onto the oceanic Piemonte Zone [Polino et al., 1990]. Ablative subduction at the continent-ocean active margin, as predicted by numerous thermo-mechanical models of subduction

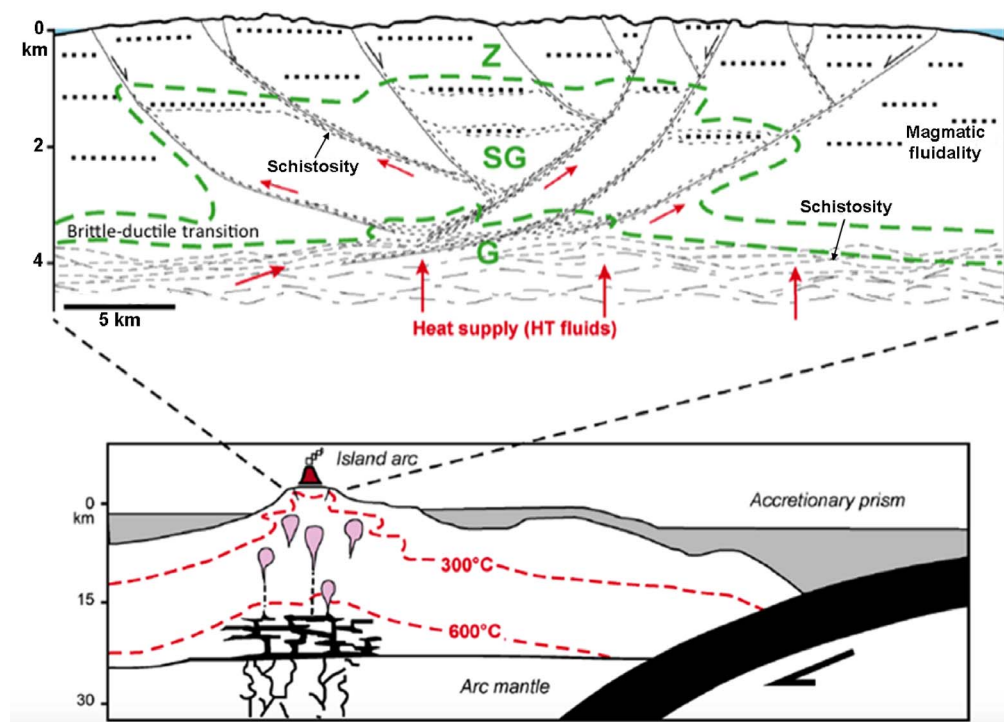


Figure 8. Metamorphic pattern and deformation in the upper crust of the Guadeloupe active volcanic arc [modified after Favier et al., 2019]. Z: zeolite; SG: sub-greenschist; G: greenschist.

zones [Gerya et al., 2002, Gerya and Stöckert, 2006, Meda et al., 2010, Regorda et al., 2021, Roda et al., 2012], is required to generate eclogite facies metamorphism in the upper plate-derived units. In such a case, the metamorphic pattern depends on oceanic subduction velocity and to a lesser extent on value of shear heating at plates interface [Regorda et al., 2021].

In contrast, continental subduction driven by the down going oceanic lithosphere is responsible for HP to UHP metamorphism in the lower plate-derived units. In the latter case, continental subduction may be considered as the premises of continental collision.

Taking into account the contrasted ages of HP and/or UHP metamorphism (see above), the western Alps offer thus an iconic example of a belt where both upper and lower plate-derived continental slices are subducted (Figure 10).

3.2.2. Continental collision zones with hidden subduction pattern and diffuse cryptic suture zones

The Variscan chain is an 8000 km-long belt formed as the consequence of Palaeozoic subduction and collision events [Martínez Catalán et al., 2020, Matte, 2001, Schulmann et al., 2014]. Successive rifting, subduction and collision of different continental micro-blocs and/or island arcs were identified before the final collision of Laurussia and Gondwana [Franke et al., 2017]. Finally, late orogenic strike-slip faulting driving block rotations, and/or late orogenic extension led to the present-day configuration [Edel et al., 2018, Faure, 1995, Faure et al., 2005].

In the tectonic framework of the Europe Variscan belt, the French Massif Central (FMC) essentially belongs to the Moldanubian Zone, the internal zone of this orogen [Lardeaux et al., 2014, Martínez Catalán et al., 2020, Figure 11]. The architecture of the FMC results from the tectonic superposition of allochthonous tectono-metamorphic units upon para-autochthonous units [see Faure et al., 2009, with references therein].

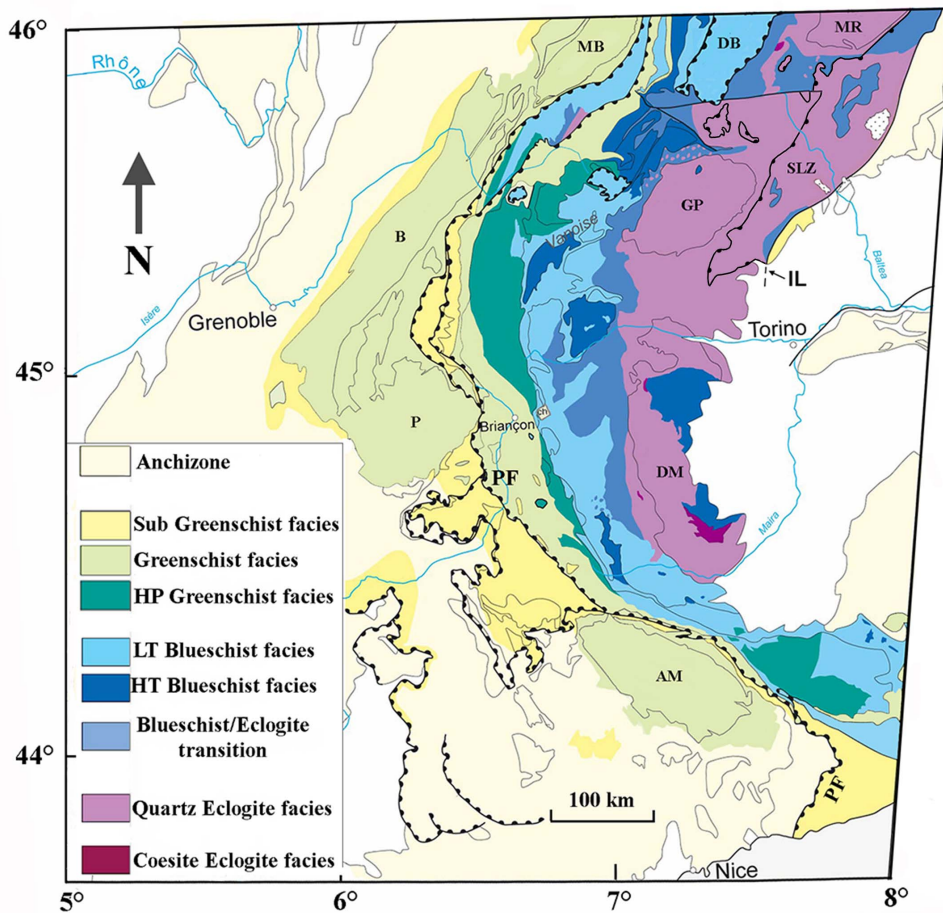


Figure 9. Metamorphic map of the western Alps [modified after Oberhänsli et al., 2004, Lardeaux, 2014a]. PF: Penninic front; IL: Insubric Line; AM: Argentera-Mercantour; P: Pelvoux; B: Belledonne; MB: Mont-Blanc; DM: Dora-Maira; GP: Gran-Paradiso; MR: Monte Rosa; DB and SLZ: Austro-alpine Dent-Blanche and Sesia-Lanzo Zone respectively.

The allochthonous units are characterized by the occurrence of eclogites associated with garnet-bearing peridotites [Lardeaux, 2014b] which are the only visible witnesses of the early subduction prior to the continental collision. Unlike in the Western Alps, blueschists and Low- T eclogites are rare in the Variscan belt [Ile de Groix in the Armorican Massif or Saxothuringian zone of the Bohemian Massif, Ballèvre et al., 2009, Schulmann et al., 2009, with references therein]. In the FMC, the main metamorphic field gradient is related to thickening processes during continental collision. The protoliths of these eclogites range from E and N-MORB tholeiites to ocean-floor tholeiites or supra-subduction zone basalts [Berger et al., 2005, 2006, Briand et al.,

1988]. The FMC eclogites are generally severely retrogressed under lower-pressure and HT conditions, resulting in the disappearance of HP peak equilibrium assemblages. However, since the discovery of small-sized relicts of coesite within garnets from the Mont du Lyonnais eclogites [Lardeaux et al., 2001], the application of modern methods of thermodynamic modelling to these rocks demonstrated their formation under UHP to HP conditions, thus under paleo-gradients typical of subduction zones [Benmamar et al., 2020, Berger et al., 2010, De Hoym de Marien et al., 2020, Lotout et al., 2018, 2020, Figure 12].

Following these last studies, eclogite facies metamorphism is Devonian (400–365 Ma) and thus coeval

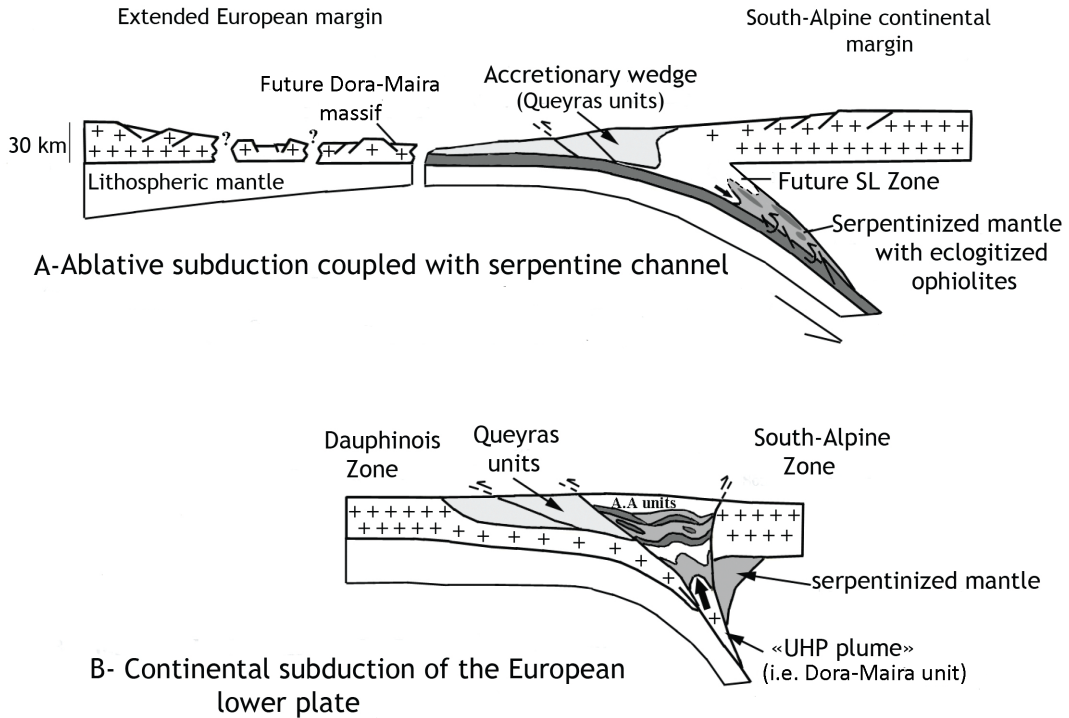


Figure 10. One of the possible tectonic sketches showing two episodes of continental subductions during the evolution of the western Alps [modified after Lardeaux, 2014a].

with the development, at the scale of the entire European Variscan belt, of arc and back-arc magmatic systems drawing a nowadays hidden subduction system [Schulmann et al., 2022]. In this framework, the Montagne Noire eclogites (southern FMC) offer a different scenario, with conditions of higher temperature and an age of 310 Ma [Whitney et al., 2020]. Although this age is widely debated [Faure et al., 2014, Faure, 2020, 2021], they could be witnesses of lower crust fragments exhumed from the root of orogenic thickened crust rather than relics of suture zones.

3.2.3. Barrovian inverted metamorphism: archetype of collision-related metamorphism

First defined by Barrow in the meta-pelites in the southeast Highlands of Scotland (Caledonian orogeny, 1893, 1912), Barrovian metamorphism is archetypical for orogenic metamorphism in continental collision belts. It is characterized by a se-

quence of zones and the typical “index” mineral in each is: Chlorite zone, Biotite zone, Garnet zone, Staurolite zone, Kyanite zone and Sillimanite zone [see Spear, 1993, for a review]. For example, in the FMC, this sequence is temporally associated with progressive under-thrusting of the lowermost units (i.e. Lower Gneissic and Para-autochthonous Units) during continental collision [Burg et al., 1989, Faure et al., 2009]. In these lowermost units, this sequence defines a syn-tectonic prograde evolution from greenschist facies to medium-pressure granulite facies conditions and to crustal anatexis generally during the first stages of decompression along the P - T - t path (Figure 13).

In collision belts, this metamorphism is frequently described as “inverted”. First reported in the European Variscan belt [Bohemian Massif and FMC, Jung and Roques, 1936, Suess, 1926], it has become an emblematic example in the Himalaya with the report of an increase in metamorphic grade towards the

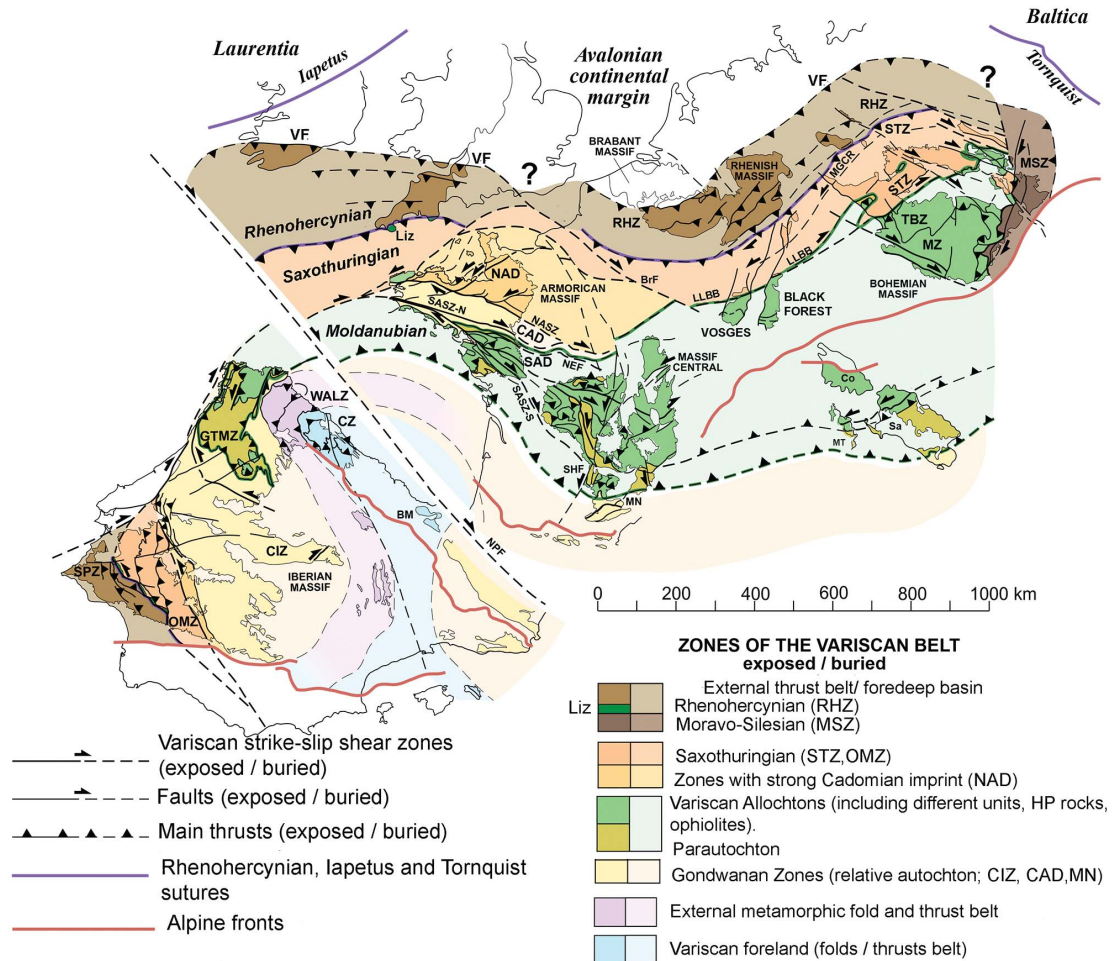


Figure 11. The French Massif Central in the framework of the late Carboniferous pattern of the Variscan belt [modified from Martínez Catalán et al., 2020]. VF: Variscan Front; OMZ: Ossa-Morena Zone; CIZ: Central Iberian Zone; SPZ: South Portuguese Zone; GTMZ: Galicia-Trás-os-Montes Zone; WALZ: West Asturian-Leonese Zone; CZ: Cantabrian Zone; BM: Basques Massifs; NPF: North Pyrenean Fault; Liz: Lizard Complex; NAD: North Armorican Domain; CAD: Central Armorican Domain; SAD: South Armorican Domain; NASZ: North Armorican Shear Zone; SASZ-N: South Armorican Shear Zone (northern part); SASZ-S: South Armorican Shear Zone (southern part); NEF: North-sur-Erdre Fault; SHF: Sillon Houiller Fault; MN: Montagne Noire; BrF: Bray Fault; RHZ: Rhenohercynian Zone; LLBB: Lalye-Lubine-Baden-Baden Fault Zone; MGCR: Mid-German Crystalline Rise; STZ: Saxothuringian Zone; TBZ: Teplá-Barrandian Zone; MZ: Moldanubian Zone; MSZ: Moravo-Silesian Zone; Co: Corsica; Sa: Sardinia.

Main Central Thrust [Johnson, 2005, Le Fort, 1975]. If the link between Barrovian inverted metamorphism and thrust tectonics is undeniable in numerous belts, different models are in competition to explain it: (1) thrust-related isotherms inversion inducing “hot-iron” effect with deep and hot metamorphic units upon colder ones [Molnar and England, 1990],

(2) post-metamorphic large-scale folding of isograds [Searle and Rex, 1989], (3) post-metamorphic sequential stacking of contrasted metamorphic zones [Brunel and Kienast, 1986, Jessup et al., 2008], (4) heterogeneously distributed shearing across the tectonic sequences [Jamieson et al., 1996], (5) shear heating [England et al., 1992] and (6) variations in lo-

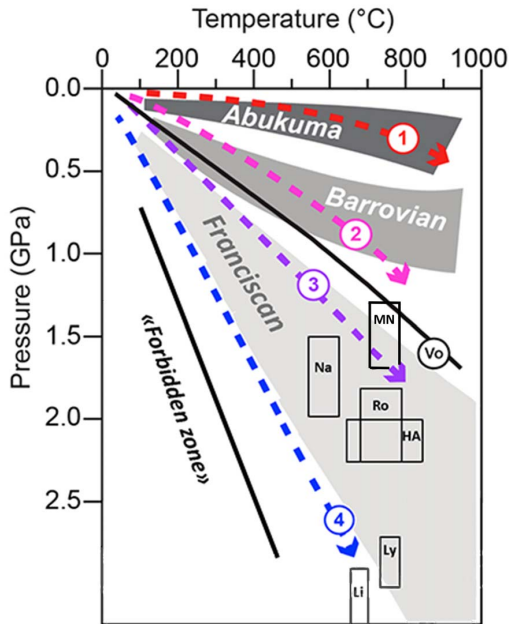


Figure 12. P - T conditions of Massif Central eclogites. Legend as Figure 1B. Na: Najac; Ro: Rouergue; HA: Haut-Allier; Ly: Monts du Lyonnais; Li: Limousin; MN: Montagne Noire (see text for references).

calization of deformation along the main thrust zone in association with contrasted thermal budget of the overriding unit [Štípská et al., 2019].

3.2.4. Continental collision zones with late-orogenic collapse

In the southern and eastern parts of the Variscan FMC, Barrovian-type metamorphic field gradients are replaced by a new high-temperature/low-pressure, up to significant partial melting, metamorphic pattern during Late Carboniferous [i.e. Pennsylvanian, Gardien et al., 1997, Lardeaux, 2014b]. This late metamorphic evolution is the record of high paleo-gradients, ca. 70 °C/km, associated with post-thickening intracontinental extension. The latter results from collapse of the orogenic crust that also led to the formation of intramontane basins [Faure, 1995, Faure et al., 2009, Gardien et al., 2022]. These high-temperature conditions developed during pressure decrease (i.e. crustal thinning) and are in most cases coeval with intrusion of mantle-derived mafic magmas at the base of the widely molten orogenic

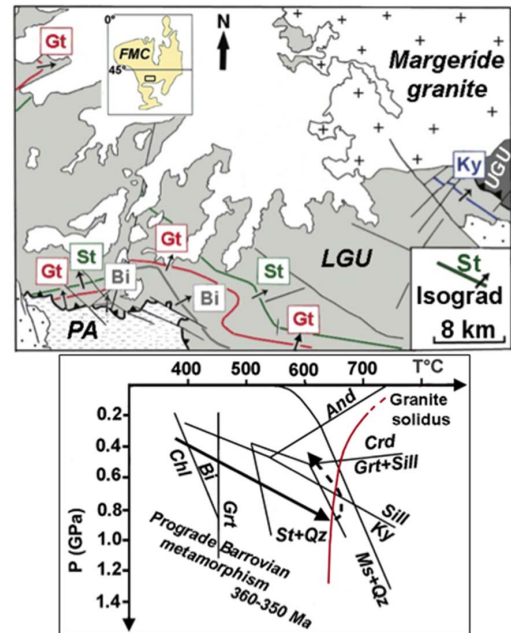


Figure 13. Carboniferous Barrovian metamorphism in the Lot serie [Massif Central, modified after Burg et al., 1989, Lardeaux, 2014b]. UGU, LGU, PA: Upper Gneiss Unit, Lower Gneiss Unit and Para-Aurochton respectively [after Faure et al., 2009]. Chl: Chlorite; Bi: Biotite; Grt: Garnet; St: Staurolite; Qz: Quartz; Ms: Muscovite; Sill: Sillimanite; Crd: Cordierite; And: Andalusite.

root [Ledru et al., 2001]. In domains of continental collision zones affected by these types of late-orogenic events, the early regional-scale metamorphic imprints can be completely erased and replaced by Abukuma-type metamorphic field gradients similar to those identified in magmatic arcs.

4. Ductile deformation and metamorphic rate controls

4.1. Deformation features and micro- to meso-structures of metamorphic rocks

Since Zwart [1962], the first approach to the relationship between metamorphism and tectonics was temporal. The chronologic relationships are based on the identification of ante-, syn-, or post-tectonic

minerals. This type of analysis is always fundamental, making it possible to identify under which P - T conditions, and therefore at what structural level a given deformation episode has developed. In return, the recognition of superposed generations of tectonic structures is a relative chronological marker that constitutes a robust basis to construct P - T - t paths, as successfully exemplified in the Western Alps [Ballèvre, 1988, Bousquet et al., 2008, Dal Piaz et al., 1983, Gosso et al., 2015, Lanari et al., 2012, Lardeaux et al., 1982, Spalla et al., 2005, Zucali and Spalla, 2011]. However, in naturally deformed terrains, the amount of finite strain is always heterogeneously distributed [Ramsay, 1980] resulting in the common occurrence of un-deformed rock volumes embodied within multiply folded domains and/or ductile shear zones. Thus under constant P - T conditions, metamorphic rocks may be subject to different strain rates giving rise to contrasted metamorphic textures (i.e. coronitic, foliated and mylonitic textures).

4.2. Ductile deformation and achievement of thermodynamic equilibrium

Teall [1885] reported the first demonstration of ductile deformation control on kinetics of metamorphic reactions, describing the progressive transformation of undeformed meta-dolerites into hornblende schists within shear zones from NW Scotland. From the early 1980s until quite recently, numerous studies have highlighted the critical role of intra- or inter-crystalline deformation mechanisms in the various stages involved in a metamorphic reaction [i.e. dissolution of reactants, chemical transfer with or without fluids, nucleation and growth, Gratier et al., 2013, Hobbs et al., 2010, Jamtveit and Austrheim, 2010]. High strain rates, grain size reduction, generation and migration of crystal defects and tectonically induced paths for fluids migration decrease the necessary activation energy for mineral reactions and thus increase reaction kinetics. In metamorphic terrains, mylonitic rocks within shear zones are the best candidates for sampling rocks in which thermodynamic equilibrium is achieved for a given metamorphic stage, while undeformed rocks with coronate textures offer the best chances of finding preserved relic minerals. These concepts lead to a sampling strategy that, if applied, improves the construction of P - T - t paths in metamorphic rocks.

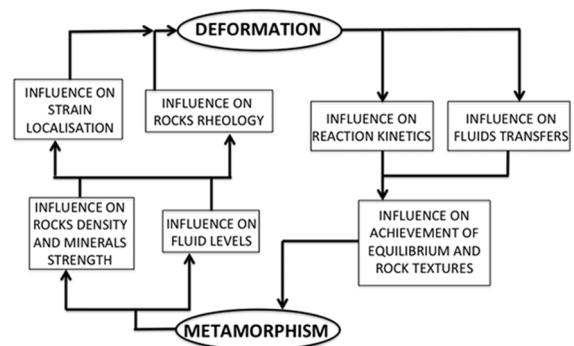


Figure 14. Graphical chart showing the links between metamorphic and tectonic processes [modified after Hobbs et al., 2010].

4.3. Metamorphic evolutions and associated crustal rheology

Deformation and metamorphism interrelate in a reciprocal way. Deformation increases the kinetics of reactions but, in return, phase transitions can induce significant variations in rock density and crustal rheology. A world-renowned example is the transformation of oceanic basalts and gabbros (density ca. 2.9) into eclogites (density ca. 3.5), and then became an important driving force of oceanic subduction. Furthermore, in the upper continental crust, mainly constituted by quartzo-feldspathic lithology, quartz is regarded as the weaker phase whose deformation mechanisms control crustal rheology. Still in the continental crust, lower-crust derived mafic rocks are generally considered as one of the strongest lithology [Ranalli, 1997]. However, rocks metamorphosed under granulite facies conditions reveal rheological inversions for instance for feldspar and quartz, for garnet and feldspar or more generally for mafic and felsic rocks as clearly reported from southern Madagascar Pan-African high-strain zones [Martelat et al., 2012].

Summing up, the multiple interactions between metamorphism and deformation may be illustrated in Figure 14.

5. Conclusions: future perspectives

After this review that explored some links between metamorphism and tectonics, for the future the following tasks seem exciting and relevant:

(1) improving the methods of thermodynamic modelling, particularly the quantification of the real reactive rock volumes to consider for models definition, (2) acquiring high-resolution temporal constraints on P - T - t paths with systematic petrochronologic studies (combined in-situ dating and chemical analyses, comprising trace elements, of metamorphic minerals), (3) understanding better in situ deformation processes using laboratory experiments to address overcoming strain rate and temperatures issues (4) progress in the application of non-equilibrium thermodynamics to metamorphic and tectonic processes to shed light on the finite rock structures and their potential record in active tectonic processes (5) as most active subduction zones are ablative, better quantify in orogens the volumes of HP/UHP tectonic units derived from the overriding upper plates, (6) identifying the volumes of metamorphic rocks which have been subject to strictly identical P - T conditions and better explain the intriguing metamorphic heterogeneities reported from many metamorphic domains, (7) producing thermo-mechanical models that better reflect the fluxes of material exchanges between subducted oceanic and/or continental crust and mantle as well as the crustal relamination and trans-lithospheric diapirism reported from some emblematic orogens (i.e. Himalayas, Variscan belt, ...).

Conflicts of interest

The author has no conflict of interest to declare.

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