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Exhumation of oceanic blueschists and eclogites in subduction zones: Timing and mechanisms

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A B S T R A C T

High-pressure low-temperature (HP–LT) metamorphic rocks provide invaluable constraints on the evolution of convergent zones. Based on a worldwide compilation of key information pertaining to fossil subduction zones (shape of exhumation *P–T–t* paths, exhumation velocities, timing of exhumation with respect to the convergence process, convergence velocities, volume of exhumed rocks,…), this contribution reappraises the burial and exhumation of oceanic blueschists and eclogites, which have received much less attention than continental ones during the last two decades.

Whereas the buoyancy-driven exhumation of continental rocks proceeds at relatively fast rates at mantle depths (≥cm/yr), oceanic exhumation velocities for HP–LT oceanic rocks, whether sedimentary or crustal, are usually on the order of the mm/yr. For the sediments, characterized by the continuity of the *P–T* conditions and the importance of accretionary processes, the driving exhumation mechanisms are underthrusting, detachment faulting and erosion. In contrast, blueschist and eclogite mafic bodies are systematically associated with serpentinites and/or a mechanically weak matrix and crop out in an internal position in the orogen.

Oceanic crust rarely records *P* conditions > 2.0–2.3 GPa, which suggests the existence of maximum depths for the sampling of slab-derived oceanic crust. On the basis of natural observations and calculations of the net buoyancy of the oceanic crust, we conclude that beyond depths around 70 km there are either not enough serpentinites and/or they are not light enough to compensate the negative buoyancy of the crust.

Most importantly, this survey demonstrates that short-lived (<~15 My), discontinuous exhumation is the rule for the oceanic crust and associated mantle rocks: exhumation takes place either early (group 1: Franciscan, Chile), late (group 2: New Caledonia, W. Alps) or incidentally (group 3: SE Zagros, Himalayas, Andes, N. Cuba) during the subduction history. This discontinuous exhumation is likely permitted by the specific thermal regime following the onset of a young, warm subduction (group 1), by continental subduction (group 2) or by a major, geodynamic modification of convergence across the subduction zone (group 3; change of kinematics, subduction of asperities, etc).

Understanding what controls this short-lived exhumation and the detachment and migration of oceanic crustal slices along the subduction channel will provide useful insights into the interplate mechanical coupling in subduction zones.

Contents

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1. Introduction

Subduction zones are crucial areas to constrain interplate coupling (Hyndman et al., 1997; Conrad et al., 2004; Heuret and Lallemand, 2005), recycling processes to the mantle (Bebout, 1996; Schmidt and Poli, 1998; Bebout, 2007) and thermal structures of arc-magmatism (Iwamori, 1998; Peacock and Wang, 1999). They are also the distinctive locus of high-pressure low-temperature metamorphism (HP–LT; Ernst, 1970, 1972; Goffé and Chopin, 1986; Okay, 1989; Maruyama et al., 1996), as confirmed by deep drilling in active trenches (Maekawa et al., 1993; Fryer et al., 1999).

Over the past twenty years, many studies have dealt with exhumation processes of HP–LT rocks, chiefly because: (1) compared to subduction forces burying rocks at depth, processes that return rocks towards the surface are still poorly understood (for example Platt, 1993; Jolivet et al., 1998b; Ring et al., 1999; Jolivet et al., 2003), (2) the discovery that continental crust could be buried deeper than previously thought (>100 km; Chopin, 1984; Smith, 1984) challenged the research community, (3) exhumation velocities derived from *P*–*T*–*t* paths, though spanning a wide range from <mm/yr to a few cm/yr (Ernst, 1988; Rubatto and Hermann, 2001; Baldwin et al., 2004), are generally surprisingly lower than subduction plate velocities (except for Philippot et al., 2001), (4) the metamorphic evolution of HP–LT rocks brings invaluable insights into deep crustal processes (Green, 2005).

For the sake of clarity and terminology, the tectonic setting of exhumation in subduction zones is recalled in Fig. 1a. The prevalent view is that, during a period of oceanic subduction, the oceanic crust and the overlying sediments, part of which can be decoupled from the crust and accreted to form the accretionary wedge, are dragged at depth along the subduction plane into the so-called subduction channel (Fig. 1a; Shreve and Cloos, 1986; Cloos and Shreve, 1988). Exhumation of some of these rocks metamorphosed under HP–LT conditions may then take place in the wedge and/or in the channel (e.g. Jolivet et al., 2003). When a large continental piece (i.e., a passive margin or an isolated block) enters the subduction zone it may also be dragged by 'continental subduction', but generally only during a restricted period of time (c. 10 My; e.g. Ernst, 2001; Chopin, 2003), after which collision develops. The introduction of the low-density continental material is generally thought to be responsible for the choking of subduction, which then stops or jumps outboard of the continental block (e.g. Stern, 2004).

Research on the exhumation of HP–LT rocks mainly focused so far on continental high- to ultrahigh-pressure rocks (UHP; for example, Wain, 1997; Kurz and Froitzheim, 2002; Chopin, 2003; Hacker et al., 2003a), partly due to the diagnostic occurrence of the quartz polymorph, coesite (as noted by Bucher et al., 2005) and to the finding of UHP rocks from ever increasing depths (Fig. 1b). The new paradigm suggests that the buoyant UHP continental rocks return at plate velocities from mantle depths (1–5 cm/yr; Duchêne et al., 1997a;

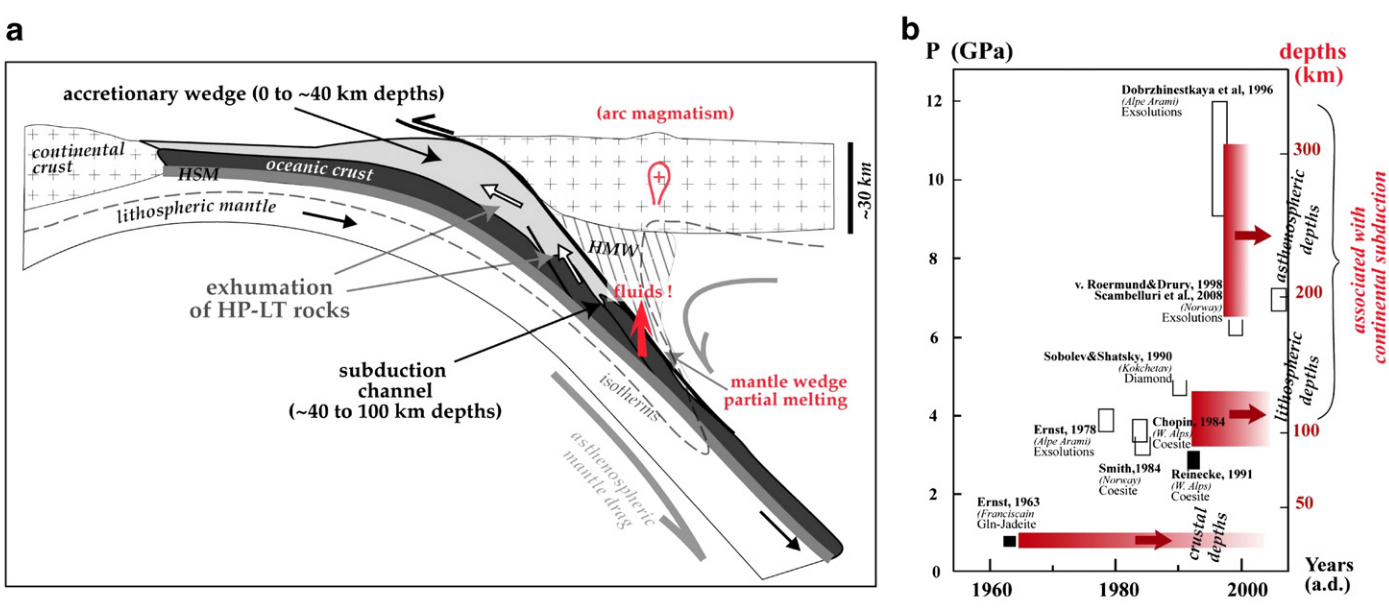

Fig. 1. (a) Sketch depicting the tectonic setting of the exhumation of high-pressure low-temperature (HP–LT) rocks in subduction zones. HMW: hydrated mantle wedge; HSM: hydrothermalized slab mantle. (b) Compilation of the main steps of discovery of HP–LT rocks illustrating the recovery of rock samples returned from increasing depths with time (Ernst, 1963, 1978; Chopin, 1984; Smith, 1984; Sobolev and Shatsky, 1990; Reinecke, 1991; Dobrzhinetskaya et al., 1996; van Roermund and Drury, 1998; Scambelluri et al., 2008). Empty boxes: continental material; black boxes: oceanic material. Note that no oceanic material has been documented from pressures >2.8 GPa so far (Reinecke, 1991, 1998).

Ernst et al., 1997; O'Brien, 2006), in some cases with the help of matrix serpentinites (Guillot et al., 2001), then decelerate in the crust, and that mechanisms driving exhumation vary with depth.

These conclusions pertain to continental subduction, however, that is to the late stages of subduction processes proper. The situation is somewhat different for the denser, oceanic HP–LT rocks, which have received much less attention, and for which exhumation velocities are still unresolved (Amato et al., 1999; Agard et al., 2002; Rubatto and Scambelluri, 2003; Federico et al., 2005). Some of them are demonstrably exhumed during oceanic convergence (for example Franciscan and Cuban rocks; Ernst, 1988, Baldwin, 1996; Schneider et al., 2004), but in other cases mainly during continental subduction, as we will later show (Western Alps, New Caledonia). Recent data for the Zagros (Agard et al., 2006) also showed that, during oceanic convergence, oceanic material was returned along the subduction plane (from depths ≥40–50 km back to depths ≤15–20 km) during short-lived time periods only, that is discontinuously with respect to the subduction duration.

In order to improve our understanding of the processes that take place at depth in subduction zones, we herein investigate the exhumation of oceanic HP–LT rocks worldwide. We more specifically address the following pending questions: (1) When are oceanic HP–LT rocks exhumed in the convergence history? Is exhumation steady or discontinuous and, in the latter case, what are the controlling processes (subduction of seamounts, obduction, etc)? (2) What are their exhumation velocities? (3) What are the driving forces? To what extent are sediments and oceanic lithosphere exhumed coevally and by the same process and/or how does the oceanic crust (+mantle) detach from the leading, sinking slab? (4) What does their exhumation tell us about the interplate mechanical coupling in subduction zones (e.g. Conrad et al., 2004) on the long-term (i.e., several My)?

For this purpose we compiled available data from various, generally young localities worldwide (<150 Ma for most of them), whose geodynamic context and oceanic origin are well-constrained (Fig. 2). For clarity, we will herein refer to two main types of oceanic material: oceanic sediments and oceanic 'crust', although in practice the latter may comprise a significant proportion of associated ultramafics derived from the lithospheric slab mantle and/or mantle wedge.

The example of the Western Alps is considered first, owing to the wealth of available data, both thermobarometric and radiometric, and for the insights it provides on key exhumation concepts and geodynamic processes. Other localities are considered next, focusing on the shape of exhumation *P–T–t* paths, on exhumation velocities, on the timing of exhumation with respect to the convergence process, and on other parameters listed in Table 1. The key constraints provided by subduction modelling are also reviewed and taken into account.

From this survey we conclude that each material (continental, sedimentary or oceanic) entering subduction zones has its own exhumation mode, and that for the latter category (the most poorly constrained, still), discontinuous exhumation appears to be the rule.

2. Insights from the Alpine case study

The Alpine case study has many advantages: it gives fruitful insights from a well-constrained geological setting, it offers the opportunity to evaluate the extent of lateral variations along a continuous subduction zone, and the lack of post-subduction collisional-related heating ensures a good record of subduction processes. It also serves as a reference frame for the comparison with other settings worldwide.

2.1. Overview of the geological context of the Western Alps

The Western Alps result from the closure of the approximately NS Valais and Liguro-Piemontese slow-spreading oceans from ~100 Ma onwards (for complete reviews of the Alpine geodynamic context: Lemoine et al., 1986; Coward and Dietrich, 1989; Polino et al., 1990; Lagabrielle and Lemoine, 1997; Stampfli et al., 1998; Dercourt et al., 2000; Dal Piaz, 2001; Agard et al., 2002; Oberhänsli et al., 2004; Rosenbaum and Lister, 2005; Ford et al., 2006), through an E-dipping subduction zone below Apulia/Africa at rates <20 mm/yr (Le Pichon et al., 1988; Lapen et al., 2003). During the Eocene (45–40 Ma), the thin, leading edge of the continental European margin reached the subduction zone and a short-lived (~10 My; Duchêne et al., 1997a; Rubatto and Hermann, 2001) period of continental subduction took place before the onset of collision between Eurasia and Apulia/Africa (at ~35–30 Ma; Sinclair and Allen, 1992). In the following, we refer to continental subduction as the period (40±5 Ma) during which continental crust material is (transiently, as recalled in the introduction) dragged to subduction depths and reequilibrates under HP–LT metamorphic conditions. During subsequent collision (~35–30 Ma

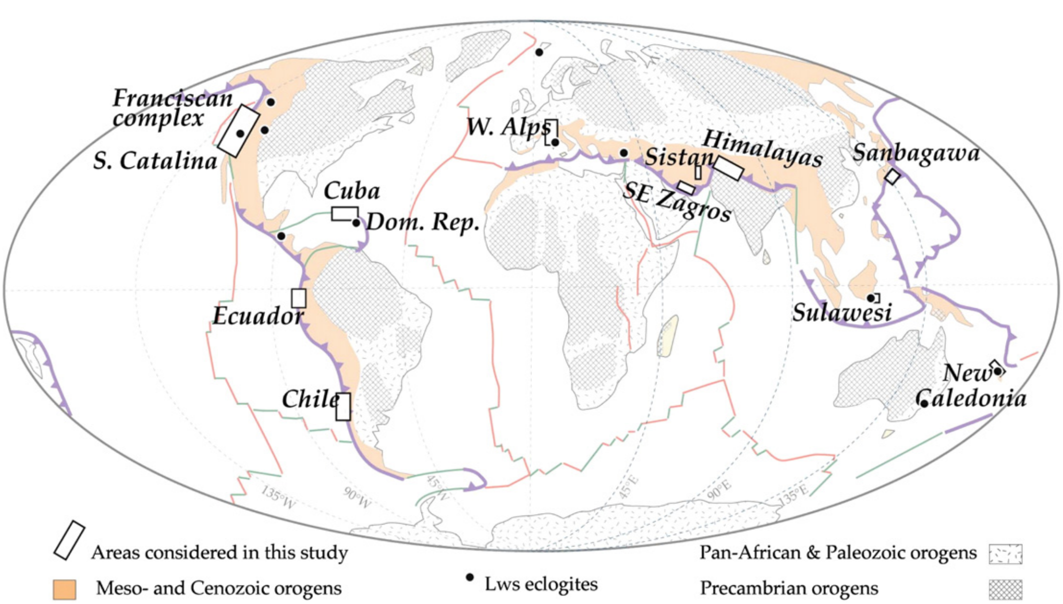

Fig. 2. Map showing the HP–LT areas considered here. Precambrian cratons and more recent orogens are also indicated. Lawsonite eclogite localities taken from Tsujimori et al. (in press).

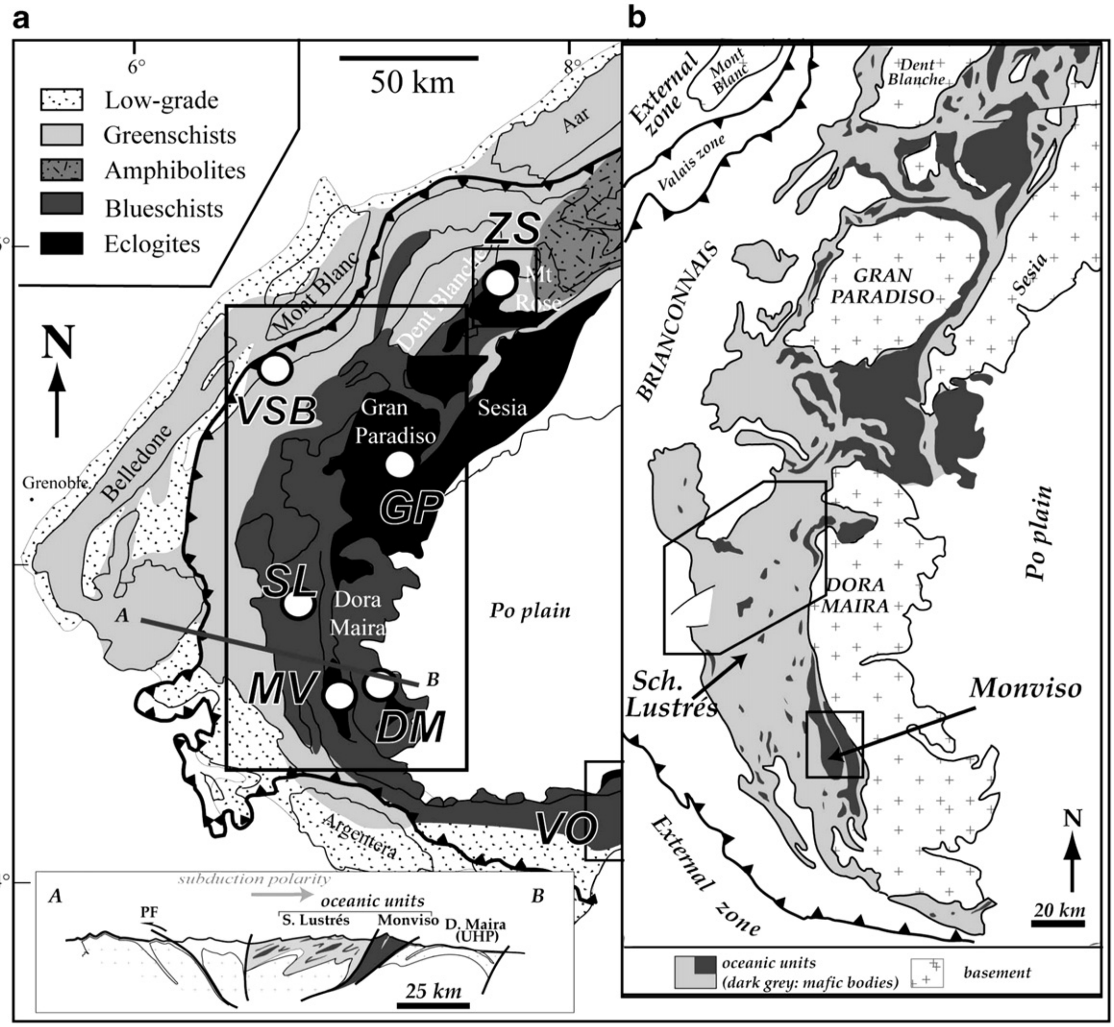

Fig. 3. (a) Metamorphic map of the Western Alps (simplified from Pognante, 1991; Oberhänsli et al., 2004; Agard and Lemoine, 2005). Inset: schematic cross-section along profile AB, showing the respective position of the ocean-derived units and the internal position of the mafic Monviso units with respect to the subuction polarity (same color scale as for panel b). Abbreviations: DM: Dora Maira; GP: Gran Paradiso; MV: Monviso; SL: Schistes Lustrés; VO: Voltri; VSB: Versoyen Grand St Bernard; ZS: Zermat-Saas. (b) Spatial distribution of oceanic sedimentary and crustal units, respectively, in the internal units. Note, again, the internal position of the mafic bodies and their close association with the internal crystalline massifs (Dora Maira, Gran Paradiso). Note the coincidence of eclogite facies conditions (panel a) with the distribution of mafic bodies (panel b).

onwards) the continental crust undergoes medium-pressure medium-temperature metamorphism (e.g., Central Alps: Todd and Engi, 1997).

Peak HP–LT metamorphic conditions grade eastward from high-pressure greenschist to blueschist facies in the Schistes Lustrés complex and in the Briançonnais (Fig. 3a; Goffé and Velde, 1984; Agard et al., 2001a,b; Oberhänsli et al., 2004) to eclogitic facies conditions in the oceanic Zermatt-Saas and Monviso units, and in the continental Dora Maira, Gran Paradiso, and Sesia massifs (Chopin, 1984; Philippot and Kienast, 1989; Pognante, 1991; Lardeaux and Spalla, 1991; Spalla et al., 1996; Van der Klauw et al., 1997; Reinecke, 1998; Fig. 3a). We focus in the following on the vanished, ~1000 km wide Liguro-Piemontese ocean. Metamorphic remnants of the Valais Ocean are very restricted and subordinate in the Western Alps (VSB, Fig. 3a; Bousquet et al., 2002) and not further considered here.

2.2. Oceanic sediments and oceanic crust (+mantle): two contrasting settings

Two contrasting settings are apparent in the geological map among ocean-derived units (Fig. 3b): dominantly blueschist facies sedimentary units with minor crustal lenses such as the Schistes Lustrés complex (SL complex; Lemoine and Tricart, 1986) are located to the west, whereas large, mainly eclogitic oceanic crust bodies (such as those found in Monviso and Zermat-Saas units) mainly crop out to the east (Pognante, 1991). Moreover, on the southern Western Alps transect, the metamorphic *P*–*T* conditions evolve progressively in the Schistes Lustrés (Fig. 4a; Agard et al., 2001a), whereas slices of units with contrasting *P*–*T* peaks characterize the Monviso area (Fig. 4b; Schwartz et al., 2000a). Besides, Monviso and Zermat-Saas units are closely associated in space with the continental Internal Crystalline Massifs (ICM, hereafter; Fig. 4a,b) and show similar exhumation-related tectonic patterns (for example eclogitic NS lineations and later, penetrative W-vergent shear senses; Philippot, 1990; Henry et al., 1993; Van der Klauw et al., 1997). These contrasts suggest that the fates of the subducted sediments and oceanic crust were different.

Another line of evidence comes from estimates of their respective volumes and from the ratio of the exhumed rock volumes over initial subducted volumes ($R_{e/s}$). Considering (1) that the ocean consisted of a ~400 m thick sedimentary cover (Michard et al., 1996; Lemoine et al., 2000) and a 2–6 km thick oceanic crust (including underlying serpentinites of this low-spreading ophiolite; Lagabrielle and Cannat, 1990) and (2) that a ~1000 km wide ocean (Lemoine et al., 1986; Stampfli et al., 1998) disappeared across the entire length of the Western Alps (~500 km along strike), it can be estimated that $\sim 2\cdot 10^5$ km^3 of sediments and ~ 1–$3\cdot 10^6$ km^3 of oceanic crust (+mantle) were subducted. Assuming that oceanic units can be extended at depth

Table 1a

Compilation of available data on exhumed ocecanic rocks worldwide (see also Fig. 7 and the Appendix) formed mainly from the Cretaceous onwards, whose geodynamic context and oceanic origin are reasonably well known. Some prominent blueschist and eclogite localities were discarded, such as those from Oman (not oceanic), Greece (partly continental) or Eastern Alps (geodynamic context insufficiently well-constrained)

Area		Subduction history					
		Locality	Ages (Ma) [Subd. period] (Onset collis.)	P–T grad (°C/km)… and time evolution	Conv. Velocities (mm/a)	Age of subd. plate	Slab dip
Chile	1	East. Series (Valdivia)	[310–210?] (no collis.)	15 to 10	~70	0–100	
Western USA Franciscan complex	2	California belts	[170–100] (no collis.)	15 to <8–9	50–100	10–60	M
and others	3	Santa Catalina	[~120–90] (no collis.)	16 to 9	~50–100?	10–40	M
	4	Cascades	[35–0] (no collis.)	12–15	30	<20	L
Iran	5	SE Zagros (Hajiabad)	[150–35] (30–25)	15 8	50–60	~100	L
	6	Sistan (Ratuk)	[~100–50] –50	7–10	–	30	?
New Caledonia	7	Pam Peninsula	[60–45] (no collis.) obd 40–35	~12	50	20–30	M?
Antilles	8a	S. Cuba (Escam bray)	[120–70?]	~10	–	30–40	–
	8b	N. Cuba (N. Serp. Melange)	[130–100?] subd flip	9–10	–	<50	–
	8c	E. Cuba	[80–60?]		–	?	–
	8d	Dominican Rep. (Samana Penins.)	[130–70?]	5 to 13	–	<50	–
Western Alps	9	Sch. Lustrés	[100–45] (35)	8	5–20	50–100	M–S
	10	Voltri		8	"	100	"
	11	Monviso		8	"	"	"
	12	Zermatt-Saas		8	"	"	"
	13	Alpine Corsica	[100–45] (no coil)	7–8	"	"	"
Himalaya	14a	Shancila	[110–55]	~12–13	100–180	~50?	M–S
	14b	Sapi-Shergol	(45)			"	M?
Japan	15	Sanbagawa (Beshi+Kotsu)	[140–60] (new wedge)	9–12	50–100	60–80	L–M
Others	16	Sulawesi (Bantimala)	[140–110?] (no collis.?)	8	?		?
	17	Ecuador (Raspas)	[>150–0]	8–10	~100	50–100	M

according to the available cross-sections (Roure et al., 1990; Schwartz et al., 2000a; Schmid and Kissling, 2000), approximately $5 \cdot 10^4$ km^3 of sediments and crust are preserved in the Western Alps today. Taking an average value for the erosion rate (0.5 mm/yr; Duchêne et al., 1997b) over the last 25 Ma, the volume of exhumed sediments and oceanic crust initially was ~10^5 km^3 each. These mass-balance calculations thus yield rough $R_{e/s}$ estimates of 0.5 and 0.03–0.1 for the sediments and the oceanic crust, respectively. Even if these values are somewhat crude, they nevertheless underline the fact that only a very small percentage of the subducted oceanic crust (+mantle) was preserved (<5 vol.%) compared to metasediments (~30–50 vol.%).

2.3. Oceanic sediments: the Schistes Lustrés paleoaccretionary complex

The Schistes Lustrés complex is mainly made of Upper Mesozoic pelagic metasediments (De Wever and Caby, 1981; Lemoine et al., 1984; Polino, 1984). Agard et al. (2001a) showed that the *P–T* conditions evolve rather continuously in the SL complex from 1.2–1.3 GPa–350 °C to 1.8–2.0 GPa–500 °C, as exemplified by the progressive eastward increase of the phengitic substitution in carpholite- or chloritoid-bearing assemblages (Fig. 4a). Subordinate, dm-sized mafic bodies found in the SL complex, particularly in the east of the complex, yielded comparable *P–T* values (Schwartz, 2002). The SL complex, as for the classical Franciscan complex (see below), therefore represents a fossil accretionary wedge, with metasediments scrapped off the underlying oceanic crust during subduction, whose tectonic and metamorphic architecture is roughly preserved (Agard et al., 2001a; Schwartz, 2002; Seno et al., 2005).

2.4. Oceanic crust (+mantle): slices and slabs

The main structural features and the *P–T* conditions of the three main oceanic crustal massifs of the Western Alps, together with northern Corsica, are recalled below.

Monviso: This elongate massif comprises several coherent slices of mafic and ultramafic material and insignificant amounts of metasediments. Two main slices (~20 km long), one of mafic material to the west, one dominated by serpentinites to the east, represent 90% of the massif (Fig. 4b). Contrasting peak *P–T* conditions were found in the different mafic slices (Fig. 4b; Messiga et al., 1999; Schwartz et al., 2000a), leading Schwartz et al. (2000a) and Guillot et al. (2004) to propose that this massif corresponds to a nappe of units sampled from various depths along the subduction channel as they returned to the surface. Maximum *P–T* conditions at 2.4 GPa–630 °C are found in Mg-rich metagabbros near the Lago Superiore (Messiga et al., 1999; star no. 2 in Fig. 4b). Serpentinites in Monviso may also display the prograde formation of olivine+Ti-clinohumite after previous antigorite assemblages (Lombardo et al., 1978).

Zermatt-Saas: This ophiolite fragment from the Liguro-Piemontese Ocean (Li et al., 2004; Fig. 4c) comprises a mixture of mafic bodies (glaucophane schists, eclogites, metagabbros) and ultramafic rocks containing antigorite–forsterite–diopside–chlorite–Ti-clinohumite, which are isofacial with the eclogites, and minor metasediments (Bearth, 1959; Ernst and Dal Piaz, 1978). Most *P–T* estimates of the litterature range from 1.8 GPa (Barnicoat and Fry, 1986; Cartwright and Barnicoat, 2002) to 2.8 GPa (in coesite-bearing metasediments; Reinecke, 1991, 1998) for temperatures around 600±30 °C. In the northern part of the Zermatt-Saas unit, however, Bucher et al. (2005) demonstrated that the same HP/UHP conditions, around 2.7 GPa/550–600 °C, prevailed at least 30 km along strike, as typified by the widespread garnet–omphacite–glaucophane–epidote–chloritoid± talc±chlorite asssemblage. These authors thus proposed that the oceanic crust detached from a 'return-point' depth of 100 km, which more or less coincides with antigorite breakdown depths (and water liberation; see discussion below). In the area south of the Aosta valley, recent *P–T* estimates point to somewhat lower, yet homogeneous pressure conditions over several tens of km (e.g., 2.1 GPa–550 °C, Martin et al., 2008; 2.3 GPa–540 °C, Agard et al., unpubl. data).

Table 1b

Area		Exhumation history								
		Sedim or crust	Exhum period (Ma)	Time lag (My) between exhum and onset of subduction (or area collision:*)	Nature of the material	Relation/ serpentinites	Rock volumes (metabas)	Re/s (vol.%)	Tect setting C/S	Contin. *P–T* in Accr wedge?
Chile	1	S	300–210	Continuous	Metagraywackes	–	hm	<<1		y
		C	305–295	<10–15	Mafic bodies	Wrapped/serp.			I	–
Franciscan	2	S	150–100	Continuous	Clastics	–				y
		C	165–150	<15	Mafic blocks	Wrapped/serp.	m–hm	<<1	NI	n
S. Catalina	3	S	110–95		Graywackes		hm			
		C		<10–15	Mafic+melange	Wrapped/serp.			I	
Cascades	4	S	14–0	Continuous	Graywackes	–	m	~80–90	–	y
		C								
SE Zagros	5	S	95–85		Clastics+rayw	–	m			n
		C	95–85	Short-lived	Mafic+volc	Wrapped/serp.	hm–km	<<1	I	
		C	110–85?		Blocks in melange	Serp dominated	hm–km		I	
Sistan	6	S								
		C	>70	Before coll.	Dism ophiolite	Wrapped/serp.			I	n
N. Caled.	7	S	45–35		Clastic+melange		hm–km	~10–20		y
		C	"	5 before*	Mafic	Wrapped/serp.		<1	I	
S. Cuba	8a	S	80–70	~10 before*	Carb micaschists	Various blocks	hm–km	–		
		C	"		Mafic	Wrapped/serp.			I	
N. Cuba	8b	S	105 (+80–70)	Continuous?	–			–		
		C	"		Melange	Serp dominated	>10 km			y
E. Cuba	8c	C	80–70?		Melange		hm–km	–		
D. Republ.	8d	S	80		Melange		hm	–		
		C	"	10 before*					–	
S. Lustrés	9	S	65–35	Continuous	Pelitic	Very few	m	~50	–	y
Voltri	10	C	49–40	Continuous?	Mafic lenses+calcsch	Some serp	hm–km		I	y?
Monviso	11	C	(50) 45–35	5–10 before*	hm to km mafic slices	Wrapped/serp.	>10 km	<3	I	n
Z. Saas	12	C	(50) 45–35	5–10 before*	hm to km mafic slices	Wrapped/serp.	>10 km	"?	I	y
Corsica	13	S	65–35?	Continuous?	Pelitic	Very few	hm–km	–	–	y
		C	?	?	Mafic	Some serp				
Shangla	14a	S	95–80	Short-lived	Volcanoclastics	–	hm	<<1		
S. Shergol	14b	C	100–90	"	Arc fragments?	Wrapped/serp.?	hm	<<1		
Sanbagawa	15	S	100–60?	Continuous	Metagraywackes	Few serp	hm–km	~10–50?	–	y
		C	95–75?	?	Mafic+volcanoclast	Some serp		<1	I	
Sulawesi	16	S	130–115		Volcanoclastics		hm–km			
		C			Mafic blocks	wrapped/serp.			I	
Andes	17	C	~130	Short-lived	Mafic+ultramafic		km	<<1		

Voltri: The Voltri massif comprises three main units (Fig. 4d; Hermann et al., 2000), from bottom to top: (1) the Voltri–Rossiglione unit, made of metasediments and metavolcanics, comprises blueschists and eclogites equilibrated at 1.3–1.8 GPa–500 °C (Federico et al., 2004); (2) the Beigua unit is made of serpentinites enclosing hm- to km-sized eclogite blocks (metabasalts and Fe–Ti metagabbros) equilibrated at 1.5 to 2±0.2 GPa/550 °C (Messiga and Scambelluri, 1991; Vignaroli et al., 2005) and comprises, in places, melange zones with blocks at 20/580 °C

Table 1b (*continued*)

Area		Exhumation history				
		P–T max (GPa/°C)	Exhum. Velocities (mm/y)	Type**	Others (oblique conv., arc magm, evolution of subd zone gradients)	References
Chile	1	1.2–1.4/400° 1.5/680°	0.6	B	Arc, cooling SZ Counterclock/blocks	Willner et al. (2004), Glodny et al. (2005)
Franciscan	2	1.0/400° 1.5/520° 2.2/550°	~1–2 ~1–2? 4–5	B	Cooling SZ Arc	Anczkiewicz et al. (2004), Baldwin (1996), Cloos (1982, 1984, 1985), Ernst (1965, 1971, 1993), Ernst and Liou (1995), Jayko et al. (1986), Kimura et al. (1996), Krogh et al. (1994), Moore and Blake (1989), Oh and Liou (1990), Sedlock (1996), Tagami and Dumitru (1996), Tsujimori et al. (2006), Wakabayashi (1990)
S. Catalina	3	1.2/660°	~1–2	B	Cooling SZ	Bebout (1991), Bebout and Barton (2002), Grove and Bebout (1995), Sorensen and Barton (1987), Sorensen (1988)
Cascades	4	–	~1	B	Arc	Batt et al. (2001), Brandon et al. (1998), Feehan and Brandon (1999), Orange et al. (1992), Ring and Brandon (1999)
SE Zagros	5	1.1/530° 1.1/530° 1.9/500°	~2–3 ~2–3 –	B	Cooling SZ? Arc, oblique Exhum in syntaxis	Agard et al. (2005b, 2006), Molinaro et al. (2005), Paul et al. (2006), Sabzehei (1974), Sabzehei et al. (1994)
Sistar	6	1.92.2/600°		?		Fotoohi Rad et al. (2005), Tirrul et al. (1983)
N. Caled.	7	1.2–1.6/550° 1.9/590°	~5? "	A?	Heating SZ? Arc?	Carson et al. (1999), Clarke et al. (1997), Cluzel et al. (2001), Fitzherbert et al. (2003, 2004, 2005), Marmo et al. (2002), Rawling and Lister (2002), Schellart et al. (2006), Spandler et al. (2005)
S. Cuba	8a	1.6–2.1/600°	2–4?	B	Arc	Kerr et al. (1999), Schneider et al. (2004), Stanek et al. (2006)
N. Cuba	8b	2.0/600°	?	A?	Arc Instab (oscill gt)	Kerr et al. (1999), Garcia-Casco et al. (2002, 2006)
E. Cuba	8c	1.5/660–750°	?	?		Cobiella-Reguera (2005), Garcia-Casco et al. (2006), Maresch et al. (2000)
D. Republ.	8d	0.7/320° 2.2/580	? 5–6	A?	Oblique, heating SZ?	Goncalves et al. (2000), Zack et al. (2004), Gorczyk et al. (2007), Krebs et al. (2008)
S. Lustrés	9	1.6–1.8/500°	~1–2	A	IsoT to cooling paths	S.Lustrés: Agard et al. (2001a,b, 2002), Goffé and Chopin (1986),
Voltri	10	1.8–2.5–600°	~3–4	A	Counterclock/blocks	Schwartz (2002); Voltri: Federico et al. (2004, 2005),
Monviso	11	2.4/630°	>5–10	A	No arc Near continental UHP	Hermann et al. (2000), Messiga et al. (1995), Rubatto and Hermann (2003), Vignaroli et al. (2005); Monviso: Guillot et al. (2004), Messiga et al. (1999), Schwartz et al. (2000a);
Z. Saas	12	2.8/600°	>5–10	A	UHP!	Z. Saas; Amato et al. (1999), Bucher et al. (2005),
Corsica	13	2.1/420° 2.4/480°	<5(?) <5(?)	A		Lapen et al. (2003), Li et al. (2004), Reinecke (1991, 1998), Rubatto et al. (1998); Corsica: Brunet et al. (2000), Caron and Pequignot (1986), Daniel et al. (1996), Fournier et al. (1991), Lahondére and Guerrot (1997), Ravna et al. (submitted for publication)
Shangla	14a	0.7/400°		B	Arc, partly oblique	Anczkiewicz et al. (2000), Guillot et al. (2007-in review), Honegger et al. (1989), Maluski and Matte (1984)
S. Shergol	14b	1.0/350–420°	?	B	Arc, syntaxis exhum.	Guillot et al. (1997, 1999), Mahéo et al. (2006), Searle et al. (1997), Tonarini et al. (1993)
Sanbagawa	15	1.8–2.1/610° 1.8–1.9/590°	– ~1	B	Arc	Banno (1986), Enami (1998), Inui and Toriumi (2002), Ko et al. (2005), Matsumoto et al. (2003), Takasu et al. (1994), Takeshita et al. (2005), Wallis (1998), Wallis and Aoya (2000), Wallis et al. (2004), Yagi and Takeshita (2002)
Sulawesi	16	1.8–2.4/620°	?	B	Arc Cooling SZ?	Miyazaki et al. (1996), Parkinson (1998)
Andes	17	~2.0/550°	?	B	Arc entrance oc. Plateau	Arculus et al. (1999), Feininger (1980), Gabriele et al. (2003), Kerr et al. (2002)

Abbreviations (by order of apparition in columns): Slab dips classified as low (L; <30°), moderate (M; 30°<.<50°), or steep (S; >50°). Sediments (S) or Crust (C). Rock volumes: typical size of rock-volumes of oceanic crust. Re/s: ratio of the exhumed rock volumes over initial subducted volumes (see text for details). Tectonic setting of crust with respect to sediments (C/S): I stands for internal, NI for non internal. Type ** refers to the classification of Maruyama et al. (1996) into type A or B protoliths. Others: SZ (subduction zone).

(Federico et al., 2007); (3) the Erro-Tobio unit, interpreted as a piece of sub-continental lithospheric mantle exhumed during the Alpine rifting and later subducted (Hoogerduijn Strating et al., 1993; Hermann et al., 2000), is made of variably serpentinized lherzolites hosting a number of mafic blocks (including Mg-rich metagabbros equilibrated at 1.8–2.5 GPa/600±30 °C; Messiga et al., 1995).

The Voltri massif therefore appears as intermediate between the Monviso/Zermatt-Saas massifs and the Schistes Lustrés complex:

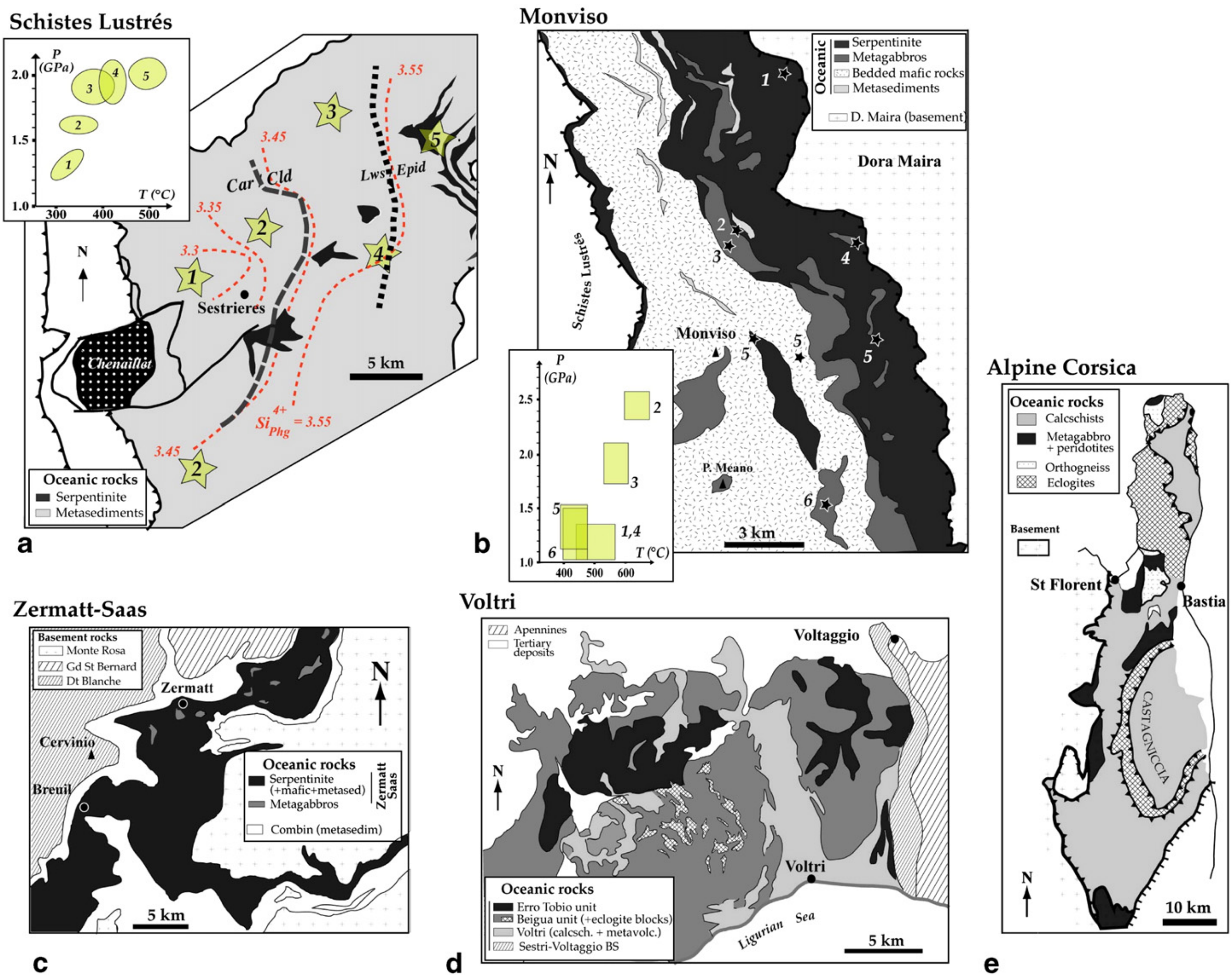

Fig. 4. Simplified geological maps of the four main metamorphosed oceanic units of the Western Alps (MV, SL, VO, ZS boxes in Fig. 3a,b) plus Corsica. Same abbreviations as for Fig. 3. (a) Map of the Schistes Lustrés area. Note the scarcity of serpentinites and the absence of large mafic units (m-sized blocks of metabasites are rare but can be found in places). Note the progressive eastward increase of both pressure (as shown by the increase of the Si content in phengite; doted lines) and temperature (as shown by the crossing of the carpholite/chloritoid and lawsonite/epidote isograds). *P–T* conditions after Agard et al. (2001a,b). (b) Map of the Monviso area (after Philippot, 1990; Schwartz, 2002). *P–T* conditions after Messiga et al. (1999) and Schwartz et al. (2000a). Note the contrasting *P–T* values a few hectometers or kilometers apart, and along strike. (c) Map of Zermatt-Saas area (after Bucher et al., 2005). (d) Map of Voltri massif (adapted from Hermann et al., 2000; Federico et al., 2005; Vignaroli et al., 2005). (e) Map of Alpine Corsica (after Jolivet et al., 1998a).

(1) calcschists and serpentinites wrapping gabbroic lenses are extensive and closely associated in the Voltri group (Fig. 4d; Federico et al., 2005 and references therein), (2) mafic bodies are smaller on average than for the Zermatt-Saas and Monviso units. Large mafic slabs, tens of km long and >km thick are not found, (3) *P–T* conditions either resemble those of the Schistes Lustrés (e.g., Voltri group) or of the Zermatt-Saas and Monviso units (Beigua and Erro-Tobio units). Some eclogitic blocks of the Beigua unit, however, provided evidence for a counter-clockwise *P–T* evolution (Vignaroli et al., 2005), (4) there is no internal crystalline massif nearby.

Corsica: HP–LT rocks from Corsica formed in the Alpine subduction zone. Corsica, however, evolved separately from the W. Alps after ~35 Ma, which marks the onset of widespread extension in the Mediterranean realm (Jolivet and Faccenna, 2000; Jolivet et al., 2003; Lacombe and Jolivet, 2005). Contrary to the W. Alps, Corsica thus escaped the later collisional deformation and provides useful insights into the fossil Alpine wedge (Fournier et al., 1991). Mafic bodies mainly crop out as nappes interleaved within dominantly blueschist facies metapelitic units analogous to those of the Schistes Lustrés (Fig. 4e). Lawsonite eclogites (Caron and Pequignot, 1986; Ravna et al., submitted for publication) and pseudotachylites in peridotites deformed under HP–LT conditions (Austrheim and Andersen, 2004) were reported. Blueschists and eclogites were finally exhumed in Oligo-Miocene extensional shear zones (Daniel et al., 1996; Jolivet et al., 1998b; Brunet et al., 2000).

2.5. Radiometric constraints: P–T–t paths and exhumation velocities

The shapes of the *P–T* paths of the five oceanic units are compared in Fig. 5a, as well as to those deduced for the neighbouring continental units. Radiometric constraints for the HP/UHP oceanic units of the Western Alps are somewhat scattered but mainly span the range 60–35 Ma (Fig. 5b).

Age constraints for the SL complex (Agard et al., 2002) support the idea of a steady state accretion between 60 and 45 Ma at least (Jolivet et al., 2003; Yamato et al., 2007). Exhumation velocities, calculated from age averages of the depth equilibrium steps (e.g. Fig. 5c), point to exhumation rates on the order of a few mm/yr (1–5 mm/yr at the most, assuming some minor excess argon; Yamato et al., 2007). A similar age range (49–40 Ma; Federico et al., 2005, 2007) and the same conclusions were drawn for the Voltri group, for which exhumation velocities are 3–4 mm/yr (Federico et al., 2005). Higher

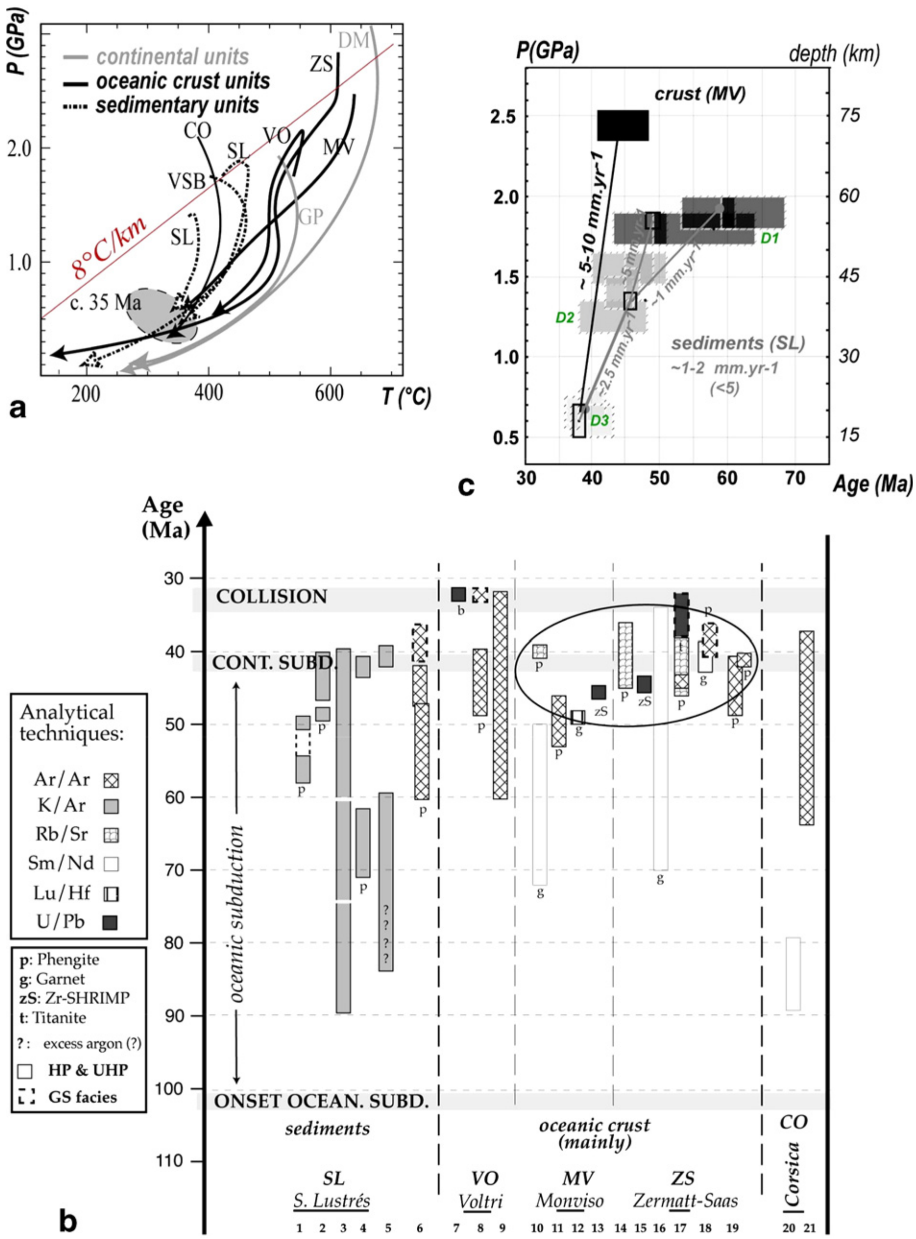

Fig. 5. (a) Compilation of P–T paths for the HP–LT areas outlined in Fig. 3a. Abbreviations: CO: Corsica; DM: Dora Maira; GP: Gran Paradiso; MV: Monviso; SL: Schistes Lustrés; VO: Voltri; VSB: Versoyen Grand St Bernard; ZS: Zermat-Saas. Data from Agard et al. (2001a,b), Bousquet et al. (2002), Bucher et al. (2005), Chopin et al. (1991), Federico et al. (2005), Jolivet et al. (1998a,b), Le Bayon et al. (2006), Messiga et al. (1995, 1999), Reinecke (1991, 1998), Skchwartz et al. (2000a), Vignaroli et al. (2005). (b) Compilation of recently published radiochronological data on the differents domains of the Western Alps and Alpine Corsica. Note, as outlined by the ellipse, the clustering of ages (apart from two Sm–Nd ages) during the period of continental subduction for Zermatt-Saas and Monviso. For each study, we give the method and minerals used (see legend box on the left hand side) and the overall domain of P–T conditions investigated (HP and UHP: high-pressure and ultrahigh-pressure conditions; GS: greenschist facies conditions). Studies are listed as follows: (1) Liewig et al. (1981), (2) Bonhomme et al. (1980), (3) Takeshita et al. (1994), (4) Delaloye and Desmons (1976), (5) Caby and Bonhomme (1982), (6) Agard et al. (2002), (7) Rubatto and Scambelluri (2003), (8) Federico et al. (2005), (9) Federico et al. (2004), (10) Cliff et al. (1998), (11) Monié and Philippot (1989), (12) Duchêne et al. (1997a,b), (13) Rubatto and Hermann (2003), (14) Reddy et al. (1999), (15) Rubatto et al. (1998), (16) Bowtell et al. (1994), (17) Barnicoat et al. (1995), (18) Amato et al. (1999), (19) Reddy et al. (2003), (20) Lahondère and Guerrot (1997), (21) Brunet et al. (2000). (c) Radiometric constraints for the Schistes Lustrés from Agard et al. (2002) and for Monviso from Monié and Philippot (1989), Duchêne et al. (1997a), Rubatto and Hermann (2003).

velocities were obtained for the Beigua unit (25 mm/yr; Rubatto and Scambelluri, 2003), based on U/Pb dating of baddeleyite from eclogite blocks, yet seem to conflict with sedimentary constraints from the nearby Early Oligocene basins (Federico et al., 2004).

Age constraints for the Monviso and Zermatt-Saas units cluster mostly between 45 and 35 Ma, apart from two poorly constrained Sm/Nd ages (Fig. 5b). Exhumation velocities deduced from these values are significantly higher than for the SL complex and range between 5–10 and 30 mm/yr (Fig. 5c; Amato et al., 1999; Rubatto and Hermann, 2003). We note that their exhumation is broadly coeval with the period of continental subduction (40±5 Ma; Fig. 5b).

Continental subduction necessarily developed after 45 Ma, given the presence of Mid-Eocene flysch deposits, later metamorphosed under HP–LT conditions, on the thin continental margin (Barfety et al., 1995), but may have been slightly diachronous along strike. Most recently published radiometric age constraints for this HP to UHP continental metamorphism suggest that peak burial conditions were reached between 40 and 35 Ma (see Agard et al., 2002 for a review).

Radiometric constraints for Alpine Corsica do not support fast exhumation (Lahondère and Guerrot, 1997), but data pertaining to the subduction stages are still too few to conclude.

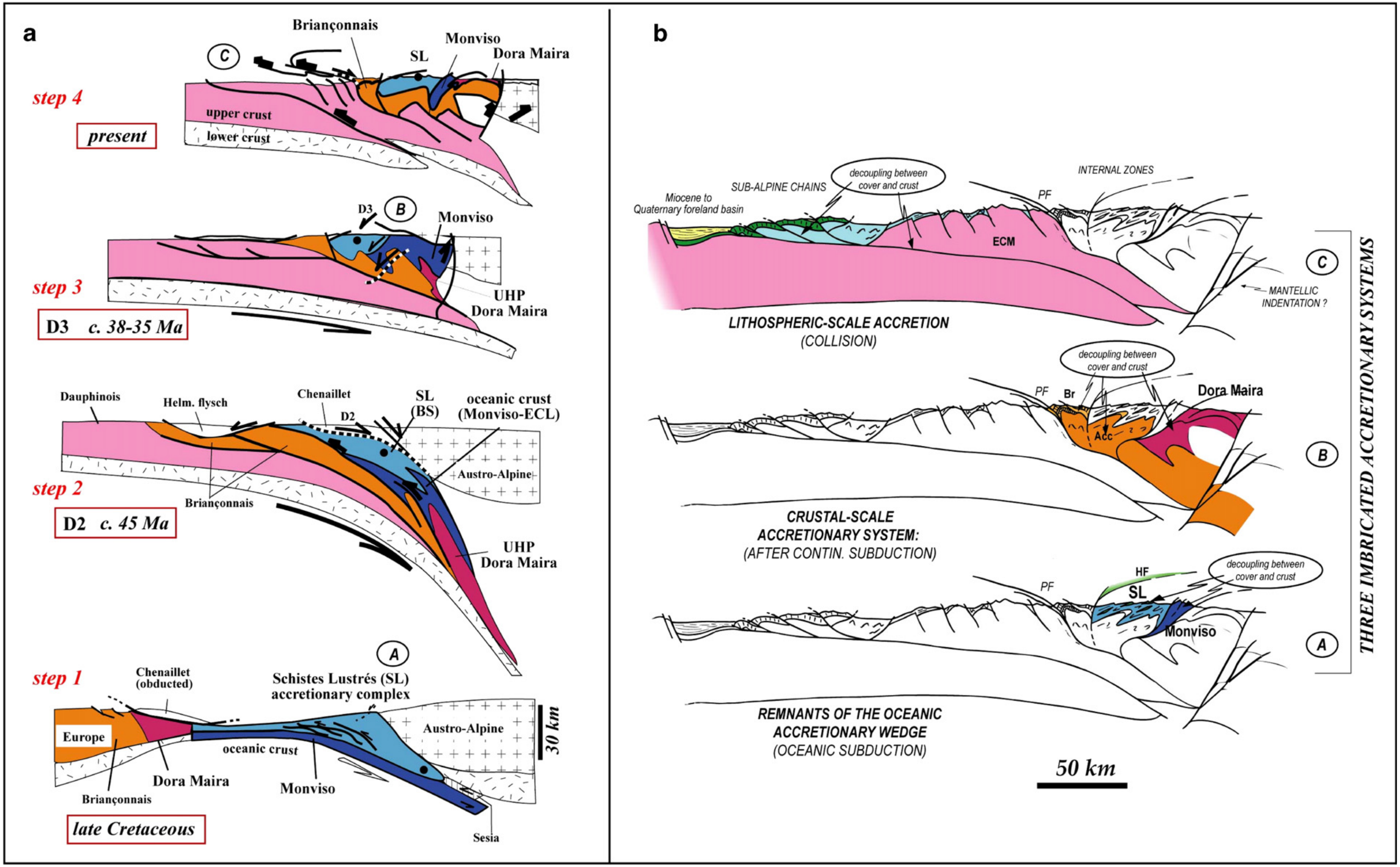

Fig. 6. (a) Geodynamic scenario for the evolution of the Western Alps (modified from Agard et al., 2002; BS: blueschist; ECL: eclogite; D2, D3: deformation stages). See text for details. (b) Lithospheric-scale cross-section (after Jolivet et al., 1996) showing the domains progressively incorporated in the Western Alpine orogen during oceanic subduction (1), continental subduction (2) and collision (3). The importance of sedimentary/crustal decoupling at each stage can also be noted. Abbreviations: Acc: Acceglio; Br: Briançonnais; HF: Helminthoid flysch; ECM: External Crystalline massifs; PF: Penninic front; SL: Schistes Lustrés complex.

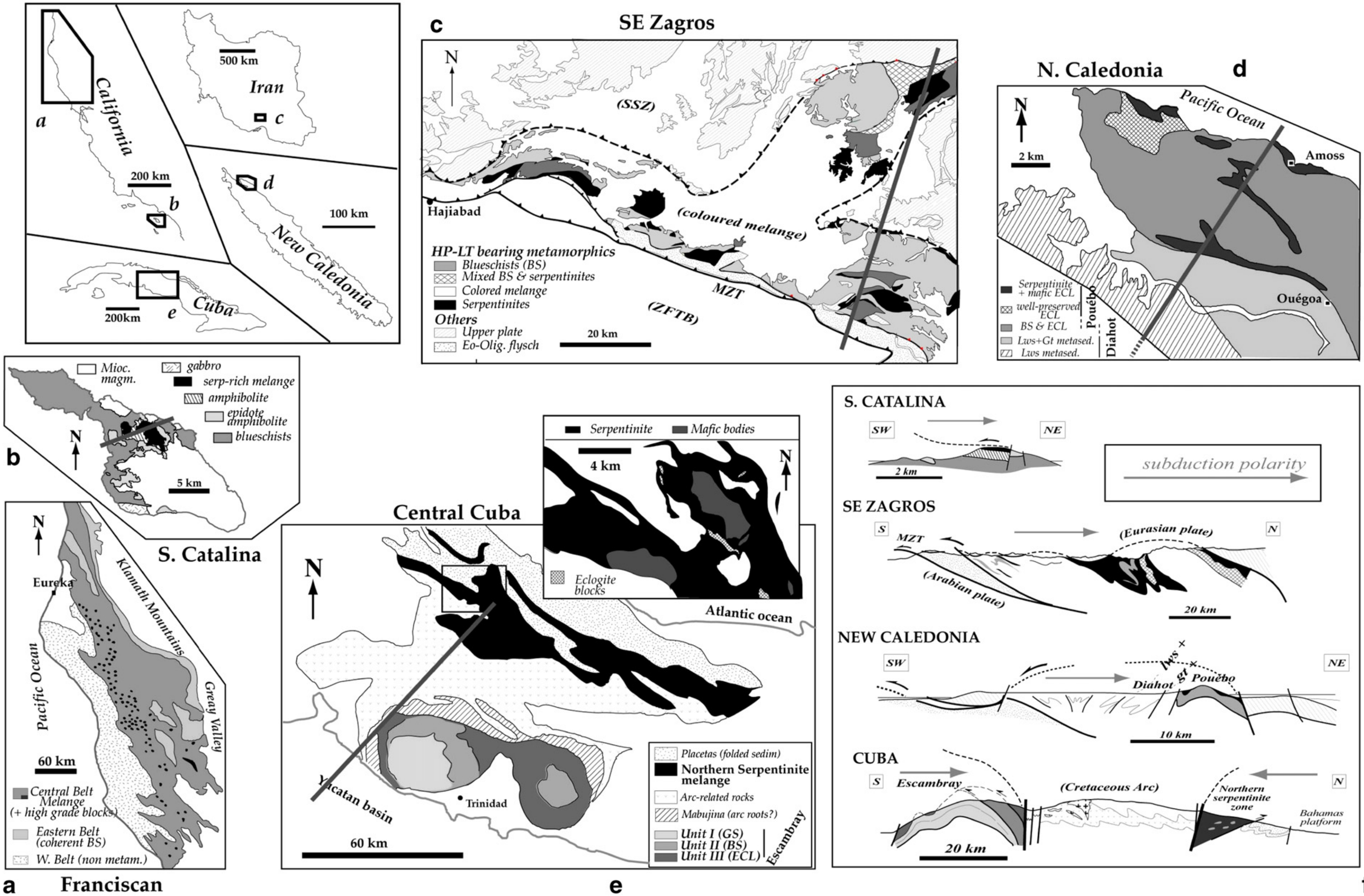

Fig. 7. Geological settings of some of the key areas considered in our study (Table 1). Profiles correspond to the cross-sections shown in panel f. (a) Main subdivisions of the Franciscan complex (after Cloos, 1985). (b) Geological map of Santa Catalina island (after Sorensen, 1988; Grove and Bebout, 1995). (c) Geological map of SE Zagros near Hajiabad (after Sabzehei et al., 1994; Agard et al., 2006). (d) Simplified geological map of NW New Caledonia (Pam Peninsula), showing the main unit subdivisions (adapted from Cluzel et al., 2001; Rawling and Lister, 2002; Fitzherbert et al., 2004, 2005). (e) Geological map of central Cuba simplified from Garcia-Casco et al. (2002) and Schneider et al. (2004). (f) Schematic cross-sections showing the position of the HP–LT rocks. Note the internal position of all mafic eclogitic units and serpentinites with respect to the subduction polarity, except for the Franciscan complex. The cross-section across NW New Caledonia was reinterpreted from available cross-sections (see above references).

2.6. Interpretation: structure and tectonic evolution of the W. Alps

This review of available *P–T–t* data therefore suggests a marked contrast between the Schistes Lustrés complex on the one hand, and the Monviso and Zermatt-Saas units on the other hand (the Voltri massif showing intermediate characteristics): the Monviso/Zermatt-Saas units show the greater *P–T* values, the greater exhumation velocities, the larger coherent tracts of mafic units, and were exhumed during the final stages of the subduction process (Fig. 5b). In addition, these two massifs crop out in the vicinity and wrap the continental units of the internal crystalline massifs (Fig. 3a). All this suggests that the exhumation of these deep-seated fragments of oceanic lithosphere is linked with continental subduction. This link is further strengthened by the exhumation rates for the UHP Dora Maira unit, which are often considered ≥1–5 cm/yr (for example, Duchêne et al., 1997b; Gebauer et al., 1997; Rubatto and Hermann, 2001).

Although a detailed discussion of the tectonic evolution of the Western Alps is beyond the scope of the present paper, a synthetic view on the geodynamic evolution from subduction to collision is shown in Fig. 6a (see Agard et al., 2002; Jolivet et al., 2003). This reconstruction places emphasis on the progressive shortening and accretion (e.g., Polino et al., 1990) which occurred during oceanic subduction (step 1, Fig. 6a), continental subduction (steps 2–3) and collision (step 4). The Alpine belt can thus be viewed as the imbrication of three successive accretionary wedges of different scale with time (from A to C: sedimentary, crustal, lithospheric; Fig. 6b). Within this frame, the Schistes Lustrés complex (and possibly the Voltri group) is exhumed during convergence in the oceanic accretionary wedge and/or in the subduction channel (step 1, Fig. 6a). The faster exhumation velocities, tighter *P–T* loops (Fig. 5a) and spatial association of the large oceanic crust bodies such as Monviso and Zermat-Saas with the HP–UHP continental units suggest a different exhumation mechanism. Their exhumation occurred later, during the locking of continental subduction and the exhumation of the crustal wedge associated with continental subduction (step 2, Fig. 6a). The final exhumation of these deep-seated oceanic units takes places through major extensional shear zones (step 3, Fig. 6a), such as those separating the blueschist from eclogitic Piemonte units (e.g., Kienast, 1973; Ballèvre et al., 1990; Ballèvre and Merle, 1993; Reddy et al., 1999), and may even involve some later, minor re-burial during the early collisional stages (Beltrando et al., 2007).

3. The exhumation of oceanic rocks worldwide

3.1. Data compilation

We present in Table 1 and in the Appendix a worldwide compilation of oceanic blueschists and eclogites from subduction zones formed mainly from the Cretaceous onwards (location on Fig. 2), whose geodynamic context and oceanic origin are reasonably well known. This compilation aims at identifying key parameters controlling the burial and exhumation of oceanic (sedimentary and crustal) rocks in subduction zones, at constraining exhumation processes and, to some extent, plate interactions at subduction zones.

This review lists the main sites where HP–LT rocks are exhumed worldwide, in various geodynamic contexts: those with high or low convergence velocities, followed or not by collision and/or obduction, with protracted or short-lived exhumation histories, etc. Major data, whenever available, are given in Table 1. Note that exhumation velocities herein are either taken from the literature or calculated from averaged

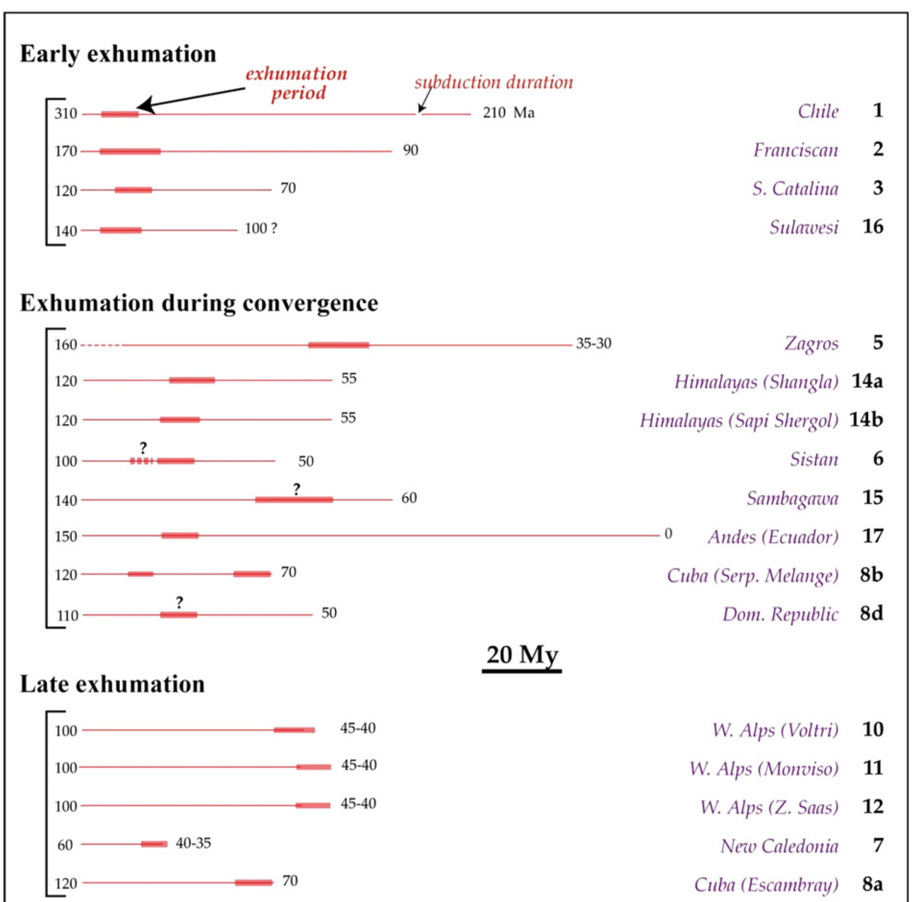

Fig. 8. Timing of oceanic crust exhumation (thick segments) with respect to the subduction history (whose duration is indicated by the thin line) in each of the investigated settings (numbers refer to ages in Ma). Numbers in the column to the right as for Table 1.

published radiometric ages pertaining to the various *P*–*T* equilibrium steps recognized on a given *P*–*T*–*t* path.

In order to both provide a brief description of the localities considered here and to ease reading, the overview of their geodynamic settings (size, nature of the material, main geological features) is placed in the Appendix. Several maps are given for the most representative localities (Fig. 7). For further details, the interested reader is referred to the selected bibliography in the Appendix and in Table 1.

3.2. Main conclusions from our survey

The following observations can be pointed out from the data of Table 1 (and Figs. 4–7):

(1) The timing of oceanic crust (+mantle) exhumation (thick lines in Fig. 8) with respect to the subduction duration is highly variable: it can either take place early (Chile, Franciscan, possibly Makran), in the midst of convergence (SE Zagros, Himalaya, Andes), or late (Western Alps, New Caledonia). Late exhumation of oceanic crust is observed in settings where continental subduction developed (New Caledonia; Zermatt-Saas and Monviso for the W. Alps), yet does not take place systematically (it is lacking in the Himalayas and SE Zagros, for example). Most importantly, Fig. 8 shows that the oceanic crust appears to be exhumed episodically, that is only during restricted, specific time windows.

(2) By contrast, sedimentary exhumation appears to be a long-lasting process: >50 My for Chile, ~50 My for the Franciscan complex, at least 25–30 My for the Alpine Schistes Lustrés (Table 1). The data also support the conclusion that exhumation returns larger volume proportions of subducted sediments (~80–90% and 30–50% for the Cascades and the W. Alps, respectively) than oceanic crust (+mantle). For example, <5% in the Western Alps, <<1% for the Himalayas, the Franciscan and SE Zagros.

(3) Oceanic crustal material is almost systematically associated with serpentinites, either as lenses wrapped in a serpentinite-rich (or exclusive) melange or as dispersed, isolated blocks. Serpentinites possibly originate from the hydrated mantle wedge (Guillot et al., 2001; Scambelluri et al., 2001; Fitzherbert et al., 2004; Guillot et al., 2004) or from the hydrothermalized slab mantle (as shown by the large oceanic pieces with gabbroic and basaltic successions of the W. Alps; Chalot-Prat et al., 2003; Li et al., 2004), or both (Hermann et

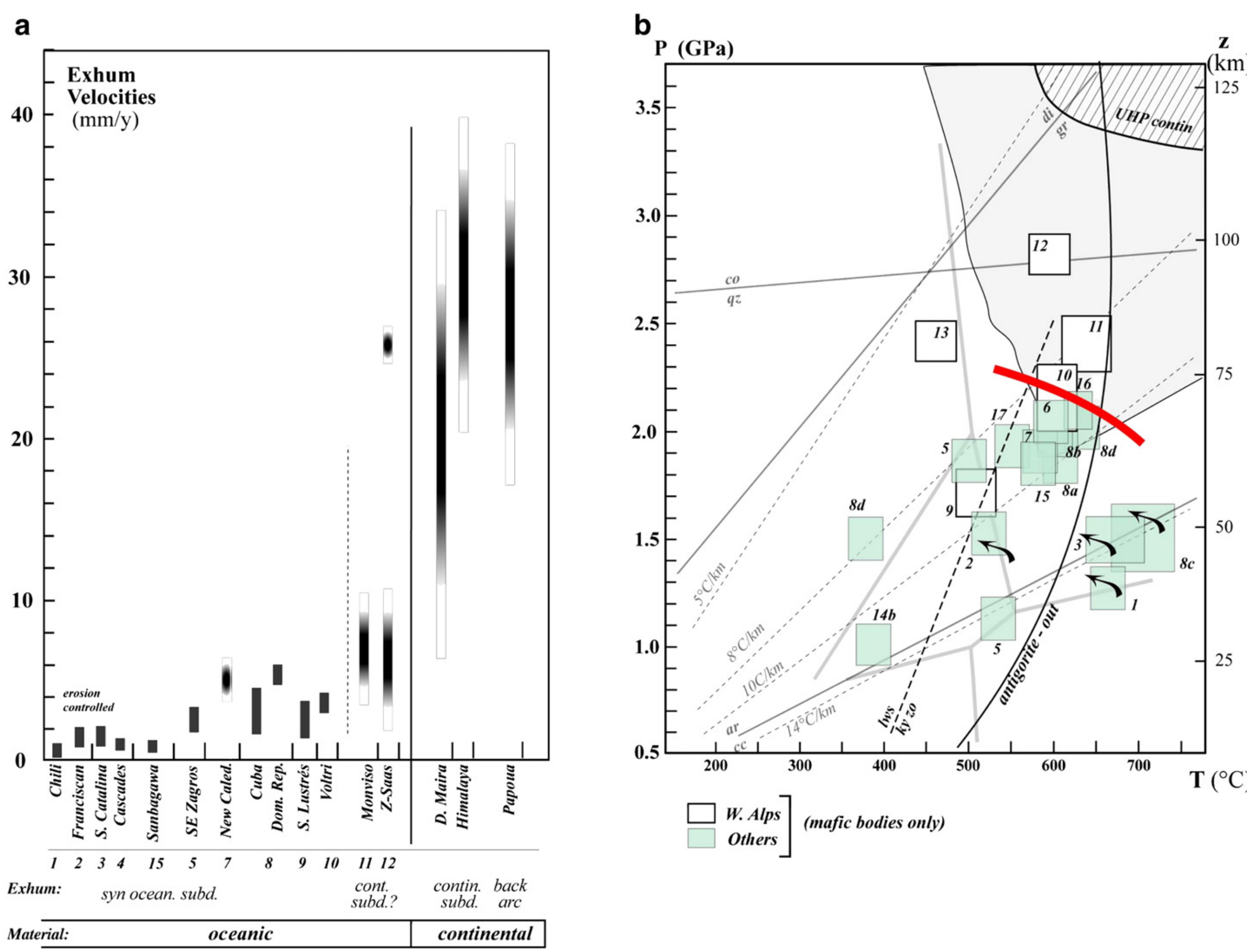

Fig. 9. (a) Plot of exhumation velocities listed in Table 1 (same numbers). Exhumation velocities herein are either taken from the literature or calculated from averaged published radiometric ages pertaining to the various *P*–*T* equilibrium steps recognized on a given *P*–*T*–*t* path. Note the sharp contrast between oceanic and continental rocks, and the highest rates for Monviso and Zermatt-Saas within the former category. (b) Plot of maximum *P*–*T* values reached by the oceanic crust in the various areas considered here (same numbers as for Table 1). Grey field: domain of negative buoyancy for the oceanic crust with respect to the adjacent mantle (see Fig. 13a and discussion). Hatched field: domain of ultra-high pressure conditions (UHP; for example D. Maira, W. Alps). Arrows: subduction settings providing evidence for cooling gradients following subduction initiation and/or for counter-clockwise *P*–*T*–*t* paths.

al., 2000). The data of Hattori and Guillot (2007) suggest that most Himalayan serpentinites originate from the mantle wedge, whereas Alpine serpentinites come from the slab mantle. Table 1 also shows that the exhumed oceanic crust generally occupies a structurally high, internal position, near the buttress (Fig. 7f). This is not the case, however, for the Franciscan complex, where coherent blueschist tracts are found at the rear of the wedge (Fig. 7a).

(4) Exhumation velocities range mainly between 1 and 5 mm/yr (Fig. 9a). Faster exhumation (>5–10 mm/yr) is only documented for Zermatt-Saas and Monviso in the Western Alps, and possibly for the Beigua unit in Voltri (see above). Exhumation velocities for New Caledonia, which lie in the upper bound too, further suggest that there is a link between faster exhumation rates and continental subduction. A word of caution is necessary, however, with respect to isolated blocks wrapped in serpentinites or in a weak matrix, for which exhumation velocities are not known with precision but were inferred to be low (Anczkiewicz et al., 2004; Glodny et al., 2005). Reappraisal of P–T conditions for the Franciscan blocks (Tsujimori et al., 2006) could lead to higher exhumation velocities than previously thought, since it is yet unclear at which P–T conditions the blocks were juxtaposed and later dispersed within the accreting metasediments.

(5) P–T gradients are in general ~8–10 °C/km, and in any case <15 °C/km (Fig. 9b). A number of settings provide evidence for cooling gradients following subduction initiation (Chile, Franciscan, Santa Catalina; arrows in Fig. 9b). P–T conditions are scattered but usually lower than 1.8–2.0 GPa for the metasediments and <2.0–2.3 GPa for the oceanic crust (Fig. 9b). A part of the Zermatt-Saas unit clearly went further down that this limit, and some of the Monviso and Voltri units (all of which belong to the W. Alps too) could have equilibrated 0.1–0.2 GPa higher. To date, the Zermatt-Saas occurrence is the deepest piece of subducted oceanic crust returned to the surface (Reinecke, 1991, 1998). Unlike continental rocks, however, no oceanic crustal rocks known to have returned from ultradeep conditions (that is 120–300 km, Fig. 1b; Green, 2005; Scambelluri et al., 2008 and references therein) have been found so far. Note that type B protoliths, as defined by Maruyama et al. (1996), experience pressure conditions largely >1.2 GPa (Table 1), contrary to the statement of these authors.

(6) P–T paths (shown in Fig. 10) essentially fall into two types, as noticed by Ernst (1988). Cooling P–T paths (Type I; Fig. 10a) characterize Chile, the Franciscan complex, Santa Catalina and most units in the Antilles, whereas isothermal (to slightly cooling) decompression paths characterize the W. Alps, Iran, the Himalayas and the Northern serpentinite melange from Cuba (Type II; Fig. 10b). All type I P–T paths (Fig. 10a), plus Corsica (Fig. 10b), stay on the low temperature side of the reaction identified by Zack et al. (2004; dotted line in Fig. 10) as the stability limit of lawsonite eclogite. Note that less typical paths are found for New Caledonia (Fig. 10c), and that the P–T paths for Sambagawa, despite the lack of subsequent collision, rather resemble those of Fig. 10b. Although these plots broadly

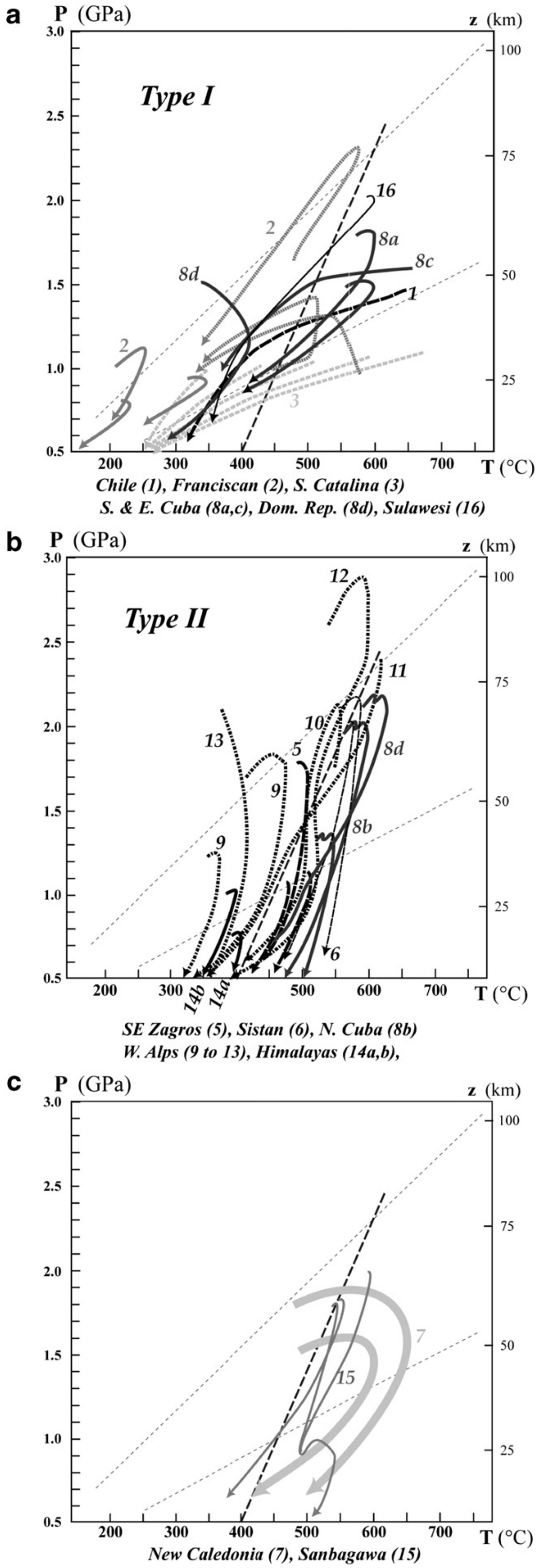

Fig. 10. P–T paths for the various HP–LT areas considered here (same numbers as for Table 1), grouped according to shape. Thin doted lines refer to the metamorphic gradients of 8 °C/km and 14 °C/km, respectively, as shown in Fig. 9. Thick dashed line: reaction lawsonite = kyanite + zoisite + SiO_2 + H_2O (after Schmidt and Poli, 1998). P–T conditions to the left of this curve favor the preservation of lawsonite eclogite according to Zack et al. (2004). (a) Type I: cooling paths. Several of them have allowed for the preservation of lawsonite eclogite (no. 2, 8d, 14). (b) Type II: isothermal to slightly cooling decompression paths (most of them lie to the right of the lawsonite eclogite preservation curve, apart from Corsica — no. 13). (c) Other paths contrasting with type I and type II (presented independently here for the sake of clarity): Sanbagawa (no. 18) and New Caledonia (no. 7).

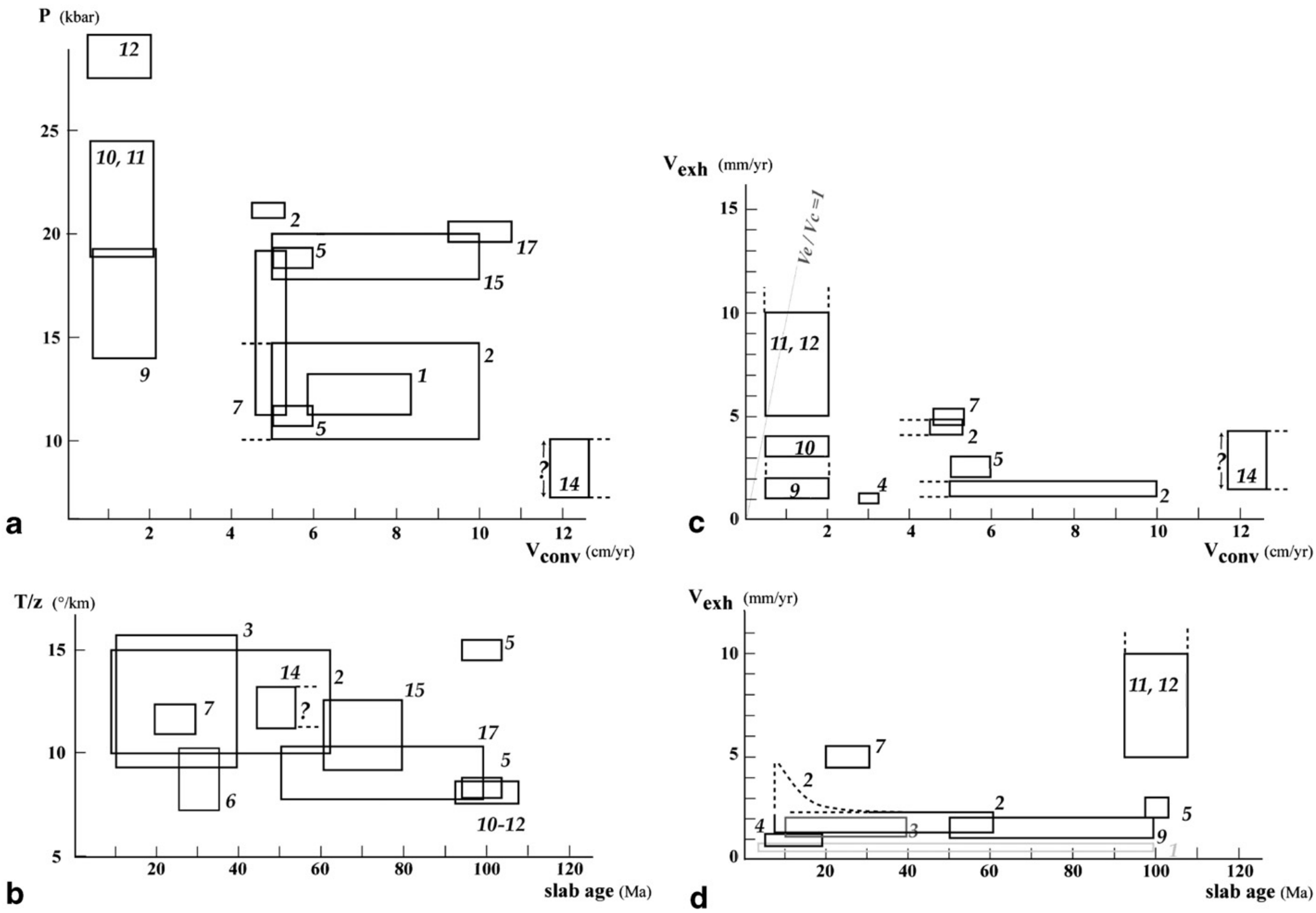

Fig. 11. Plots evidencing a weak correlation of convergence velocities or slab age on pressures and metamorphic gradients, respectively, and showing the lack of influence of these parameters on exhumation velocities (see also Table 1). (a) Plot of peak pressures versus convergence velocities. (b) Plot of metamorphic gradients versus slab age. (c) Plot of exhumation velocities versus convergence velocities. (d) Plot of exhumation velocities versus slab age.

confirm the conclusion that *P*–*T* paths fingerprint the tectonic history of subduction zones (Ernst, 1988), individual *P*–*T* paths should thus not be regarded as univocal (see also Matsumoto et al., 2003).

(7) Peak pressures seem to negatively correlate with convergence velocities (Fig. 11a), as broadly do metamorphic gradients with slab age (Fig. 11b). Yet data are very scattered and these two rather predictable correlations are not very robust at present. On the other hand, a number of parameters seem to have little or no influence on exhumation patterns and velocities (Table 1):

- Convergence velocities taken from the litterature do not affect exhumation rates (Fig. 11c). This observation points to the existence of strong decoupling between the upper, essentially sedimentary wedge and the subduction channel.
- No relationship is observed betwen subduction duration and exhumation velocities. This suggests that an exhumational steady state is reached at depth (as documented by Ring and Brandon, 1999 for the Franciscan complex).
- There is no correlation between the age of the subducting plate and exhumation velocities (Fig. 11d). The nature of the material, whether originating from a slow or fast spreading environment (with or without much ocean-floor serpentinites, respectively; Karson, 1998) could nevertheless be important (see below). On the other hand, young, warm plates (e.g. Franciscan complex, Santa Catalina; Table 1) obviously favor warm initial subduction gradients (15–16 °C/km).
- Assessing the role of slab dip is hampered by the lack of data.
- The presence (Franciscan, Himalayas; type B of Maruyama et al., 1996) or absence (W. Alps; type A) of arc-related magmatism in the upper plate does not seem to play a significant role.

4. Insights from subduction modelling

This section summarizes available data from experiments (mainly thermomechanical models) addressing the exhumation of HP–LT rocks in subduction zones during oceanic convergence, and emphasizes on the recent results obtained by our group (Yamato et al., 2007).

Analog modeling provides few insights on the exhumation of HP–LT rocks because it either focused on slab behavior during oceanic subduction (Faccenna et al., 2001; Regard et al., 2003; Funiciello et al., 2004) or dealt with continental subduction and the positive buoyancy of continental material (Chemenda et al., 1995, 2000, 2001; Boutelier et al., 2004). Similarly, most thermomechanical modelling studies dealt with the mechanisms of exhumation during continental subduction (Bousquet et al., 1997; Beaumont et al., 1999; Ellis et al., 1999; Pfiffner et al., 2000; Doin and Henry, 2001; Burov et al., 2001; Warren et al., 2008; Yamato et al., 2008; Burov and Yamato, 2008) rather than with oceanic subduction proper (Cloos, 1982; Gerya et al., 2002).

Reproducing the thermal regimes of blueschists and eclogites and their evolution through time was achieved in the early 1990s by Peacock (1987, 1990, 1996), based on the Franciscan and Santa Catalina examples, and refined until recently (for example, Peacock et al., 2005). Early thermomechanical models (Cloos, 1982, 1984; Allemand and Lardeaux, 1997; Schwartz et al., 2001) explored the circulation of particles in a deformable wedge (Dahlen, 1984). These models, however, have a kinematically constrained geometry which strongly prescribes the geodynamic setting, and yield velocities too fast (1–4 cm/yr). Recent, essentially self-consistent numerical schemes (Burov et al., 2001; Gerya et al., 2002) allow to more realistically take into account the complex interplay of the different exhumation mechanisms (erosion, tectonic thinning, buoyancy, delamination, slab detachment: both superficial and lithospheric, as emphasized by Platt, 1993). These models also take into account the feedback between *P*–*T* dependent mineralogical phase changes and the mechanical response of the system (e.g., to density and rheological changes).

Burov et al. (2001) explored the behavior of the subduction channel during continental subduction and pointed out the existence of two distinct circulation levels at 0–30 km and 30–70 km, respectively, with slower exhumation rates for the upper level. Gerya et al. (2002) successfully modeled oceanic subduction and subduction channel processes with full phase and density changes and variable hydration of the mantle wedge. Their experiments were also applied to the Alpine case (Stöckhert and Gerya, 2005). Their models, however, predict too fast exhumation velocities for the metasediments and the oceanic crust (>1 cm/yr). Serpentinites are also randomly dispersed in the metasediments, contrary to the Western Alps and to most natural settings reported above (Figs. 4 and 7; Table 1).

Yamato et al. (2007) developed a model based on subduction experiments without pre-defined kinematic assumptions in the model (yet, with no hydration of the mantle wedge), and showed that:

- Exhumation velocities are ~1–2 mm/yr in the wedge (Fig. 12a). There is a turn-over circulation lasting ~15 My, which is consistent with constraints from the Schistes Lustrés and part of the Voltri units (W. Alps; Agard et al., 2002; Federico et al., 2005) but also with the Franciscan, Cascades and Chilean examples (Brandon et al., 1998, Ring and Brandon 1999; Batt et al., 2001; Glodny et al., 2005).
- The exhumation of oceanic crust is obtained only when serpentinites are present in the slab mantle, which is consistent with the Alpine case (Li et al., 2004; Hattori and Guillot, 2007), and exhumation velocities are 3–4 mm/yr (Fig. 12b).
- Oceanic crust is exhumed towards the rear of the subduction channel, in an internal position (Fig. 12b).
- No overpressures build up in the subduction channel.

5. Discussion: implication for subduction processes and exhumation concepts

5.1. Oceanic versus continental rocks: contrasting exhumation modes

From a number of recent studies in various settings worldwide, it is now apparent that the buoyant continental crust is exhumed during continental subduction, with velocities comparable to those of plate tectonics at mantle depths (1–5 cm/yr; Fig. 9a) and later decelerates (~mm/yr) in the upper crust (W. Alps: Duchêne et al., 1997a; Rubatto and Hermann, 2001; Papoua: Baldwin et al., 2004; Himalayas: De Sigoyer et al., 2000; O'Brien, 2006; Dabie Shan: Liu et al., 2006; Kokchetav: Hermann et al., 2001; Hacker et al., 2003a). As a first approximation, the buoyancy-driven exhumation of continental rocks takes place through extrusion (Chemenda et al., 1995), although in detail HP–UHP nappes are notably thinner (~100 m–1 km) than the thickness of a normal crust (~30–35 km; Jolivet et al., 2005). Note that slab breakoff could also provide an additional help for the buoyant rise of continental rocks (Von Blanckenburg and Huw Davies, 1995), although it is not required for exhumation (Warren et al., 2008; Yamato et al., 2008).

In contrast to the continental crust, our survey reveals that subducted ocean-floor sediments are exhumed at velocities ~mm/yr (Fig. 9a; Table 1). The sediments are efficiently decoupled from the oceanic crust and tend to accumulate in the accretionary wedge, as shown by the respective volumes for the W. Alps (Table 1; see also Fig. 6b). They circulate in the wedge under the combined effects of underthrusting (Gutscher et al., 1998; Glodny et al., 2005), erosion and tectonic thinning, whose respective contributions vary from one geodynamic setting to the other (highly erosive in the Franciscan complex; Platt, 1986, 1993; Ring and Brandon, 1999; highly tectonic in the W. Alps; Agard et al., 2002).

Most oceanic crust is to be irreversibly buried, yet it occasionnally returns to the surface at velocities ~mm/yr too, either wrapped in serpentinites or in a muddy/shaly, mechanically weak matrix melange. One finds m- (Franciscan) to hm- (Iran) or km-sized blocks (New Caledonia; W. Alps; Escambray, Cuba), which are either brought back early (Franciscan, Chili), during (SE Zagros, Himalayas) or late (W. Alps, New Caledonia; some parts of Cuba) in the subduction history (see next section). The dominantly internal position of oceanic crust (+mantle) with respect to sediments (Table 1), and frequently of eclogites on blueschists, either points to specific dynamics in the subduction channel and/or to the influence of the buttress and hydrated mantle wedge in channelling oceanic crust towards the surface (see also Fitzherbert et al., 2004). Mafic and ultramafic slices occupying such an internal position were also reported from Turkey (Okay et al., 1998).

Exhumation velocities for oceanic rocks are thus, contrary to continental ones, largely inferior to plate velocities. This is also the case when looking back at the few available Paleozoic examples: rates are <3 mm/yr for the ~470–460 Ma suture zones from Svalbard (Motalafjella; Agard et al., 2005a,b) and China (Qilian; Song et al., 2007-in review).

One exception is when the exhumation of oceanic crust (+mantle) directly results from the locking of subduction processes by continental subduction (W. Alps; Fig. 6a). Only in this case do we find large volumes preserved (Fig. 4), either as continuous slab pieces (Zermatt-Saas; Bucher et al., 2005) or as slices from contrasting depths (Monviso; Schwartz et al., 2000a,b). Similarly, in this case only are rocks brought back from UHP depths (Fig. 9b). We note that earlier, smaller subducted continental pieces (such as Sesia; Dal Piaz et al., 2001; Babist et al., 2006), some of which may represent extensional allochtons (Froitzheim and Manatschal, 1996; Manatschal, 2004), may have been too small to choke the subduction and provoke the exhumation of oceanic slices (see also Brun and Faccenna, 2008).

The scarcity of lawsonite eclogites worldwide (Fig. 2; Zack et al., 2004; Whitney and Davis, 2006; Tsujimori et al., in press), due to overprinting on exhumation, is consistent with the prevalence of rather low exhumation velocities (Fig. 9a). Fig. 10a,b, however, suggest that even at slow exhumation rates (Franciscan complex, Corsica), the preservation of lawsonite eclogite is possible if refrigerent conditions are maintained on exhumation.

This survey thus further strengthens the concept of subduction channel (Shreve and Cloos, 1986; Jolivet et al., 2003; Federico et al., 2007), by showing that a large amount of decoupling exists within the upper and lower parts of the channel. The detailed exhumation pathways of the slices of oceanic crust detaching from the sinking slab remain poorly constrained, however.

5.2. Continuous versus discontinuous oceanic crust exhumation?

A priori, two end-members situations can be expected for the exhumation of the oceanic crust:

- a continuous, steady state exhumation process. A good candidate would be a 'corner-flow' accretionary process (Cloos, 1982; Shreve and Cloos, 1986; Gerya et al., 2002; Gorczyk et al., 2007) with the incorporation of oceanic crust in the wedge, possibly enhanced by serpentinization (hydration of the slab mantle and/or serpentinite diapirs).
- a discontinuous, incidental exhumation process. In this case the oceanic crust would be exhumed only during specific time windows, for example in response to the entrance of buoyant

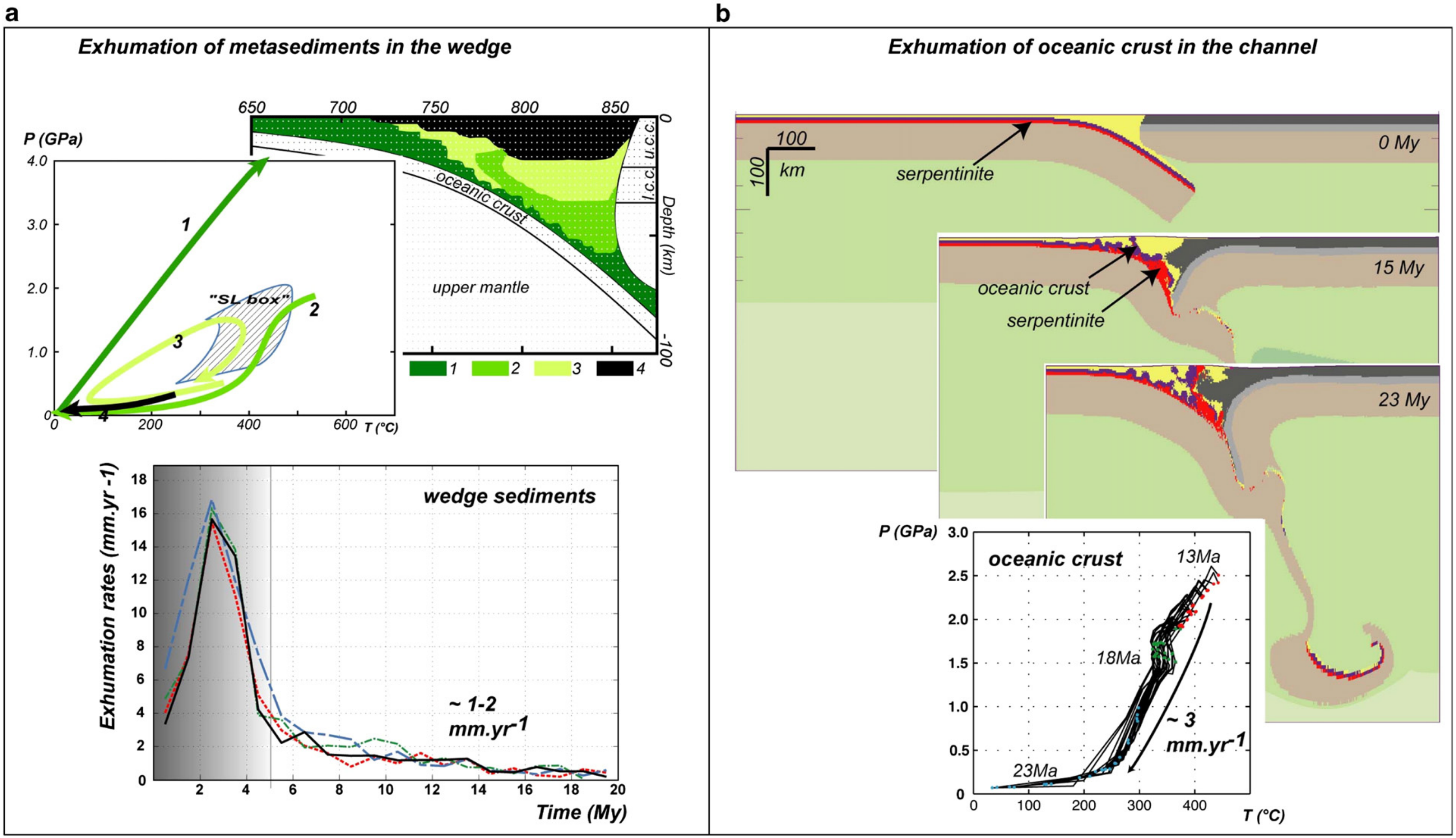

Fig. 12. Overview of the main results of the thermomechanical numerical study of Yamato et al. (2007) with application to the Western Alps. This large-strain model accounts for visco-elastic-plastic properties of the rocks, is thermally coupled, includes methamorphic phase changes and surface processes. Tracers are used to track particles and reproduce synthetic *P–T–t* paths that are compared with the observations (see text and Yamato et al., 2007 for details). (a) Schematic *P–T* paths followed by the sedimentary particles (grouped by colour) in the accretionary wedge and the subduction channel, showing that a number of them effectively enter the *P–T* field of the Schistes Lustrés (SL box). Exhumation velocities are a few mm/yr after a transient period. (b) Morphology obtained in the case of oceanic crust (blue) exhumation facilitated by the presence of an underlying serpentinized slab mantle (red). Note that the oceanic crust and serpentinites are exhumed at the rear of the wedge, at velocities ~3–4 mm/yr. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

material in the subduction wedge (for example thinned continental margin, oceanic plateau, slab asperities; Ernst, 1988; Cloos, 1993; Arculus et al., 1999; Ernst, 2001), to the modification of the convergence setting (velocities, obliquity; Agard et al., 2006), or to the cessation of subduction (for example delamination of lower crust, slab-breakoff; Von Blanckenburg and Huw Davies, 1995).

Our worldwide survey of past subduction zones reveals the following contrasting behaviors (Fig. 8; Table 1):

(1) Short-lived early exhumation (Franciscan complex, Chile).In both cases, the geodynamic setting advocates for a warm, rapidly cooling subduction zone associated with the consumption of young oceanic lithosphere (Anczkiewicz et al., 2004; Glodny et al., 2005). The short-lived exhumation (lasting no longer than 10–15 My) suggests the existence of a transient state, both thermal and mechanical, following the initiation of subduction, as documented in our oceanic subduction models (Yamato et al., 2007). It could also be hypothetized that exhumation is no longer possible after a certain time period, once the mantle wedge has become too lubricated by water hydration (Iwamori, 1998; Arcay et al., 2005).
(2) Short-lived exhumation in the midst of convergence (SE Zagros, N. Cuba).In the case of SE Zagros, this short-lived exhumation (~15–20 My versus >120 My of subduction) coincided with a sharp rise of convergence velocities and with the onset of regional-scale obduction movements (Agard et al., 2006, 2007). The same ages are in fact obtained from Eastern Turkey (Sherlock et al., 1999) to Makran (Delaloye and Desmons, 1980) and the Himalayas (Anczkiewicz et al., 2000; Maluski and Matte, 1984), which indicates that this exhumation was contemporaneous and short-lived across thousands of km. This observation points to a major, regional-scale change in interplate mechanical coupling, which affected the whole Neotethyan subduction zone below Eurasia (Agard et al., 2006). In the case of Cuba, *P–T* fluctuations were deduced from oscillatory garnet zoning in eclogites taken hundreds of kilometers apart in the Northern Serpeninite melange zone (Garcia-Casco et al., 2002). These oscillations were interpreted to reflect slab rupture and episodic ascent of eclogitic slices incorporated into the mantle wedge serpentinites (Garcia-Casco et al., 2002).
(3) Exhumation towards the end of the oceanic subduction (W. Alps, New Caledonia).The role of continental subduction in triggering crustal exhumation has already been emphasized for the Western Alps (see above). The situation is probably similar for the rather short-lived New Caledonia subduction zone, except that collision did not fully develop. Some of the oceanic rocks were underplated below the mantle wedge (Fitzherbert et al., 2004), but were not significantly exhumed until the entrance of buoyant material in the subduction zone (Northfolk ridge; Cluzel et al., 2001; Spandler et al., 2005).
(4) Configurations in which the exhumation is (almost) impossible. This is obviously the case for the Andes and Java, and for the Himalayas and SE Zagros throughout most of their subduction history (Fig. 8). These settings correspond to gently-dipping, relatively fast subduction zones capped by well-developped calc-alkaline magmatic arcs. Where collision developed (Himalayas, SE Zagros), no oceanic rocks were brought back towards the end of subduction with UHP continental slices, unlike the W. Alps (and HP–UHP continental rocks are even lacking in SE Zagros). The reason is unclear, yet it can be noted that in such settings: 1) the extensive hydration of the mantle wedge, as indicated by the presence of the calc-alkaline arcs at the surface, may efficiently lubricate the subduction channel and prevent the detachment of oceanic slices from the slab, 2) in low-angle slabs, the positively buoyant continental material will less efficiently counterbalance the dense, rapidly subducting oceanic material during continental subduction.

A major conclusion is therefore that the exhumation of the oceanic crust metamorphosed under HP–LT conditions is highly discontinuous (or 'episodic'; Fig. 8). Further insights should perhaps be sought from adequate geodynamic settings such as the Antilles (particularly Cuba) and from geophysical data and modeling (section below).

5.3. Serpentinites: a great help down to limited depths?

Major unknowns in the deeper parts of the subduction channel are whether there are maximum depths for the exhumation of oceanic crust (+mantle), and whether oceanic crustal rocks are returned from specific depths (e.g., ~100 km, according to Bucher et al., 2005) or sampled at various depths along the subduction zone (Guillot et al., 2004; Fotoohi Rad et al., 2005). Evaluation is hampered by the difficulty of determining whether the inferred *P–T* conditions pertain to the entire series or to isolated blocks only (see the discussion by Wallis and Aoya, 2000; Fitzherbert et al., 2005; Stipska et al., 2006). In any case, the compilation from Fig. 9b clearly shows that exhumation is normally prohibited beyond 60–70 km (~2.0–2.3 GPa), except in the event of continental subduction (Zermatt-Saas, Monviso).

It was previously proposed that the exhumation of oceanic rocks was facilitated by their association with serpentinites, which would counterbalance their negative buoyancy and enhance mechanical decoupling (Hermann et al., 2000; Guillot et al., 2001; Schwartz et al., 2001; Pilchin, 2005). For example, Hermann et al. (2000) estimated that the density of the Beigua unit (Fig. 4d; Voltri, Western Alps), made of 90% of serpentinites and 10% of mafic eclogites, was lowered to 2.9 g/cm^3, and that of the Erro-Tobio unit, with variably serpentinized lherzolites and minor eclogites, to 3.1 g/cm^3 (Fig. 4d), hence lower than that of the mantle.

Our Fig. 9b, by showing that the exhumed oceanic crust from the various settings studied here lies within the mostly temperature-dependent antigorite stability field (<~650 °C; Hermann et al., 2000; Guillot et al., 2001; Hacker et al., 2003b), further supports the conclusions of these workers.

We also provide new insights by calculating, in the range 0–4 GPa and 0 to 1000 °C, the net buoyancy of the oceanic crust, which corresponds to the density difference between the oceanic crust and the surrounding mantle at a given *P–T* value (Fig. 13). This net buoyancy of the oceanic crust was computed using stables phases and equilibrium densities calculated by the Theriak thermodynamic code (De Capitani, 1994). Fig. 13a shows that all exhumed oceanic crust lies outside of the negative buoyancy domain, apart from the Alpine Zermatt-Saas and some of the Monviso and Voltri massifs, which nevertheless lie inside the antigorite stability field (Fig. 9b and 13b,c).

We therefore suggest that beyond maximum depths around 70 km there are in general either not enough serpentinites (because both the slab mantle and the mantle wedge are dehydrated) and/or they are not light enough to compensate the negative buoyancy of the crust. This depth range is consistent with the scarcity, in such settings, of garnet peridotites (typically equilibrated at depths >70–80 km) associated with oceanic crust (and, in those rare cases, the depths at which they were juxtaposed is unclear; e.g., Gorczyk et al., 2007). The existence of maximum depths explains the contrasting behavior of the oceanic crust from Zermatt-Saas (continuous slab with uniform *P–T* conditions) and Monviso (slices with contrasting *P–T* conditions) more satisfactorily than the concept of a specific return depth point (Bucher et al., 2005). We note that incidental exhumation from greater depths should nevertheless be occasionnally possible with the help of continental material (continental subduction demonstrably exhumed mantle rocks from depths >300 km!; Green, 2005 and references therein), whose buoyancy is much greater than that of serpentinite at T>600° and P>2.0–2.3 GPa (Fig. 13c; Hermann, 2002).

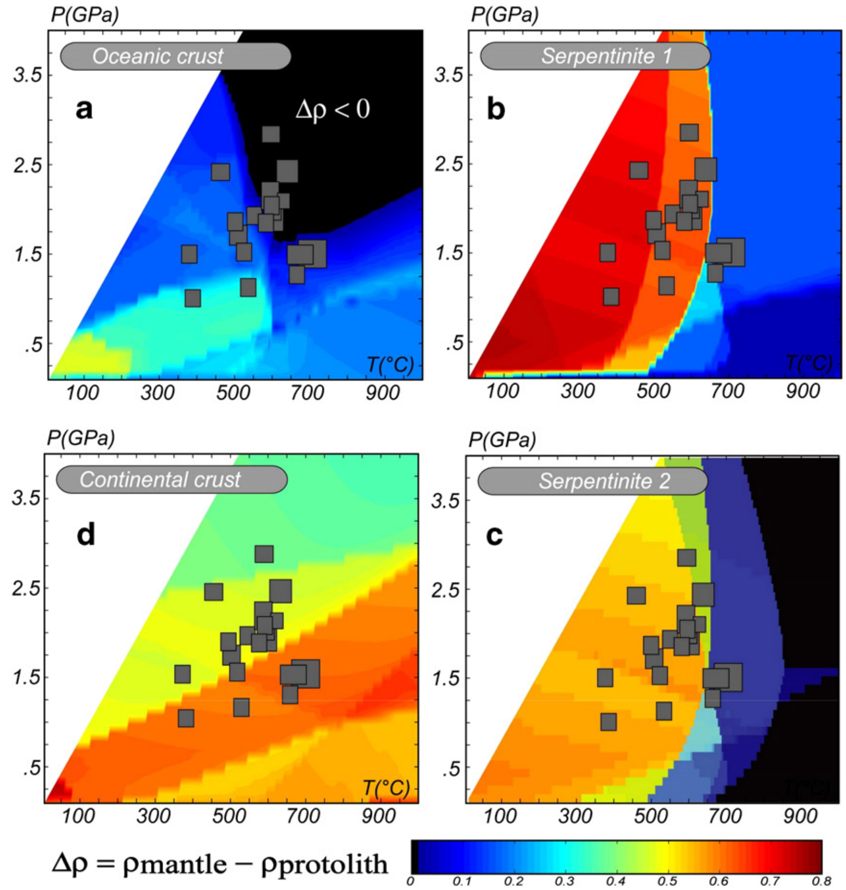

Fig. 13. Estimation of the net buoyancy of the oceanic crust (a), serpentinite (b, c) and continental crust (d), defined as the density difference between each of them, respectively, and the adjacent mantle. Serpentinite 1 corresponds to the pure Mg-rich end-member and average compositions for the oceanic and continental crust are taken from Yamato et al. (2007). Note that similar density contrasts are found for buoyancies calculated with other serpentinite compositions (Ulmer and Trommsdorff, 1995; Bromiley and Pawley, 2003). The most contrasting results are obtained for serpentinite 2 (plot c), which corresponds to the slightly Al-richer composition of Bromiley and Pawley (2003; their sample BM1913,87). Whatever the serpentinite composition, however, note that the sharp drop in buoyancy always occurs at the antigorite/forsterite reaction, whose position remains stable within 30 °C. Boxes as for Fig. 9b: maximum *P*–*T* values recorded by oceanic rocks in the subduction zones reviewed here. These diagrams, using stables phases and equilibrium densities calculated by the Theriak thermodynamic code (De Capitani, 1994), show that all exhumed oceanic crust lies outside of the negative buoyancy domain, apart from Monviso and Zermatt-Saas, which nevertheless lie inside the antigorite stability field.

5.4. Interplate mechanical coupling and exhumation processes at depth in the subduction channel

Although mantle wedge or slab mantle serpentinites obviously play a great role in the exhumation of oceanic crust, they alone do not explain how the crust (+mantle) detaches from the subducting slab. This problem of localization and detachment also exists for continental rocks (Jolivet et al., 2005). The great difference, however, is that continental rocks are intrinsically buoyant and effectively detach from the subducting plate (except perhaps in Zagros, where no HP–UHP continental rocks crop out), whereas oceanic lithosphere only rarely makes it to the surface (Fig. 8), despite the help from serpentinites.

Mechanisms based on mineral reactions triggering earthquakes were recently proposed for localization and detachment at depth in subduction zones. They include dehydration embrittlement (Kirby et al., 1996; Hacker et al., 2003b and references therein) and localized melting (Kanamori et al., 1994; pseudotachylites: Austrheim and Andersen, 2004; Andersen and Austrheim, 2006) followed by localization (Jolivet et al., 2005). Note that dehydration embrittlement will occur deeper in fast, cool subduction zones than in warm, moderately rapid ones (for example NE Japan v. SW Japan, respectively; Peacock and Wang, 1999). Once detached, the crust will encounter major decoupling interfaces such as the mantle wedge (Guillot et al., 2001; Gerya et al., 2002) and the 5–10 km thick hydrated crustal layer imaged by seismology in the upper part of the slab (Abers et al., 2006).

The fact that exhumation is discontinuous/episodic demonstrates that during most of the subduction process either the detachment from the slab and/or the later upward migration along the slab interface is inhibited. Examples of short-lived oceanic crust exhumation during convergence suggest that the interplate mechanical coupling conditions (Conrad et al., 2004) drastically changed, possibly in response to increased velocities (SE Zagros: Agard et al., 2006; Himalayas?), thermal instabilities (N. Cuba: Garcia-Casco et al., 2002) or slab rollback triggered by the entrance of continental blocks (Brun and Faccenna, 2008).

It is beyond the scope of the present paper to address in detail mechanical coupling and force balance at subduction zones (Scholz and Campos, 1995 and references therein). We point out, however, that the discontinuous exhumation of oceanic blueschists and eclogites, could be used as a proxy for interplate mechanical coupling and the mechanics of subduction zones.

From the preceeding examples, unfavorable interplate coupling conditions for exhumation seem to be that the oceanic crust is formed at fast spreading ridges (hence with a less hydrated slab mantle; for example Himalayas, Andes), with relatively low dip (too much friction and/or underplating below the upper plate at depth), at relatively fast rates (hence a cooler thermal regime: dehydration and detachment will take place deeper, where a counter-flow is no longer possible).

5.5. A general model for the exhumation of the oceanic crust

Although further carefully designed petrological and geophysical studies are needed, a provisional picture of processes operating at depth in subduction zones is shown in Fig. 14. It emphasizes that in general no steady state exhumation takes place, and that no exhumation is possible beyond a certain depth (Fig. 14a). Slab mantle and mantle wedge serpentinites play an important role on exhumation. By contrast, upper levels of the subduction channel are largely decoupled from the oceanic crust and circulate in the accretionary wedge (Fig. 14b; Yamato et al., 2007).

Early exhumation, in the first 5–15 My of oceanic subduction (Fig. 14c; Franciscan complex) occurs for relatively warm, rapidly cooling subduction zones (e.g. Anczkiewicz et al., 2004). Note that exhumation velocities associated with this process are still unclear. Given the characteristics of a young, warm subduction zone such as in SW Japan (Peacock and Wang, 1999), we propose that exhumation takes place through weakening resulting from the partial melting of the oceanic crust (as reported by Sorensen and Barton, 1987 for California), leading to boudinage and detachment of blocks in a weaker, sedimentary melange (Cloos, 1985; Anczkiewicz et al., 2004).

Once the subduction has cooled (particularly in subductions with velocities >~3 cm/y), dehydration reactions take place deeper down the subduction zone. In a relatively low-angle subduction channel, detachment will occur well below the upper plate and, given the pinch-out geometry, rocks can not be flushed back towards the surface along the slab interface (Fig. 14d). Mantle wedge hydration by continuous influx from the subducting slab, which results in increased interplate mechanical decoupling, further inhibites exhumation. Incidental exhumation, however, may happen in response to a major modification of boundary conditions (Himalayas, SE Zagros). Slab rollback developping in response to the subduction of large enough continental pieces can probably also significantly enhance the exhumation of oceanic material in some settings (Brun and Faccenna, 2008).

Finally, late exhumation associated with oceanic closure and continental subduction (and/or slab breakoff) will take place in relatively slow, steeply dipping subduction settings with a highly hydrated slab mantle (for example the W. Alps; Fig. 14e).

6. Conclusions

This study outlines the existence of three contrasting exhumation modes for sedimentary, oceanic crustal and continental rocks.

(1) Whereas the buoyancy-driven exhumation of continental rocks, possibly plucking ultradeep mantle rocks, proceeds at relatively fast rates at mantle depths (≥cm/yr; Duchêne et al., 1997a; Rubatto and Hermann, 2001; Ernst, 2001; Green, 2005), oceanic exhumation velocities for oceanic rocks, whether sedimentary or crustal, are usually on the order of the mm/yr.
(2) Underthrusting, detachment faulting and erosion, which are the driving exhumation mechanisms for the sediments, often preserve the continuity of the *P–T* conditions within the accreted sedimentary material (e.g. Agard et al., 2001a). In contrast, blueschist and eclogite mafic bodies are systematically associated with serpentinites and/or a mechanically weak matrix, and eclogites wrapped in serpentinites mostly crop out in an internal position in the orogen.
(3) Oceanic crust rarely records P conditions >2.0–2.3 GPa, which suggests maximum depths for the sampling of slab-derived oceanic crust. We propose that beyond such depths (~70 km), there are either not enough serpentinites (because both the slab mantle and the mantle wedge are dehydrated) and/or they are not light enough to compensate the negative buoyancy of the crust.
(4) Most importantly, this survey demonstrates that short-lived, discontinuous exhumation is the rule for the oceanic crust and associated mantle rocks: exhumation takes place either early (group 1: Franciscan, Chile), late (group 2: New Caledonia, W. Alps) or incidentally (group 3: SE Zagros, Himalayas, Andes, N. Cuba) during the subduction history. These conclusions are tentatively accounted by a comprehensive model in which exhumation is permitted by the specific thermal regime following the onset of a young, warm subduction (group 1), by continental subduction (group 2) or by a major, geodynamic modification of convergence across the subduction zone (group 3; change of kinematics, subduction of asperities, etc).

We finally stress the importance of understanding parameters controlling this short-lived exhumation and the detachment and migration of oceanic crustal slices along the subduction channel. In this respect, blueschists and eclogites are expected to shed light into the interplate mechanical coupling in subduction zones.

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Appendix A

This section provides a brief geological description of the localities considered in our compilation of oceanic blueschists and eclogites from subduction zones across the world (location on Fig. 2; selected maps in Fig. 7; data in Table 1). For further details, the interested reader is referred to the selected bibliography (see also Table 1).

A.1. Chile

Within the metamorphic basement of the Coastal Cordillera of central Chile, the Western Series constitutes the HP–LT paleoaccretionary wedge of a fossil-paired metamorphic belt dominated by metagreywackes (Kato, 1985; Kato and Godoy, 1995). This subduction zone experienced a protracted accretion history (~50–100 My; Glodny et al., 2005) and moderate exhumation velocities (0.6 mm/yr; Glodny et al., 2005). In its eastern part, blocks derived from small lenses of garnet amphibolite with a blueschist facies overprint are locally intercalated and associated with serpentinite and garnet mica-schist (Willner et al., 2004). Those mafic blocks, dated at ~305 Ma, preserve a counter-clock-wise *P–T* path indicative of cooling early in the subduction history (Willner et al., 2004). Note that this is the only Paleozoic wedge considered here, because it escaped from later deformation (unlike Svalbard, for example; Ohta et al., 1986; Agard et al., 2005a).

A.2. California and Cascades

A.2.1. Franciscan complex

HP–LT rocks of the Franciscan Complex represent the prototypical, sediment-rich, fossil accretionary wedge. It was associated with the E-dipping subduction zone operating from c. 170 to 100 Ma (Cloos, 1985; Ring and Brandon, 1999; Anczkiewicz et al., 2004), which gave birth to the Sierra Nevada batholith (Ernst, 1970). A marked contrast exists between the essentially metasedimentary Eastern belt (Fig. 7a), composed of coherent tracts of blueschist facies metagraywackes (Cloos, 1982, 1985; Jayko et al., 1986; Ernst, 1993; Kimura et al., 1996), and the Central Belt, where meter-scale mafic high-grade blueschist and eclogite blocks (knockers) are found in a serpentinite- or shale-matrix melange (Oh and Liou, 1990; Wakabayashi, 1990; Krogh et al., 1994).

The Eastern belt blueschists reached *P–T* values <1.0–1.2 GPa and 400 °C, on average (Ernst, 1971; Jayko et al., 1986; Ernst, 1993), and were exhumed at velocities around 0.5–1 mm/yr (Ring and Brandon,

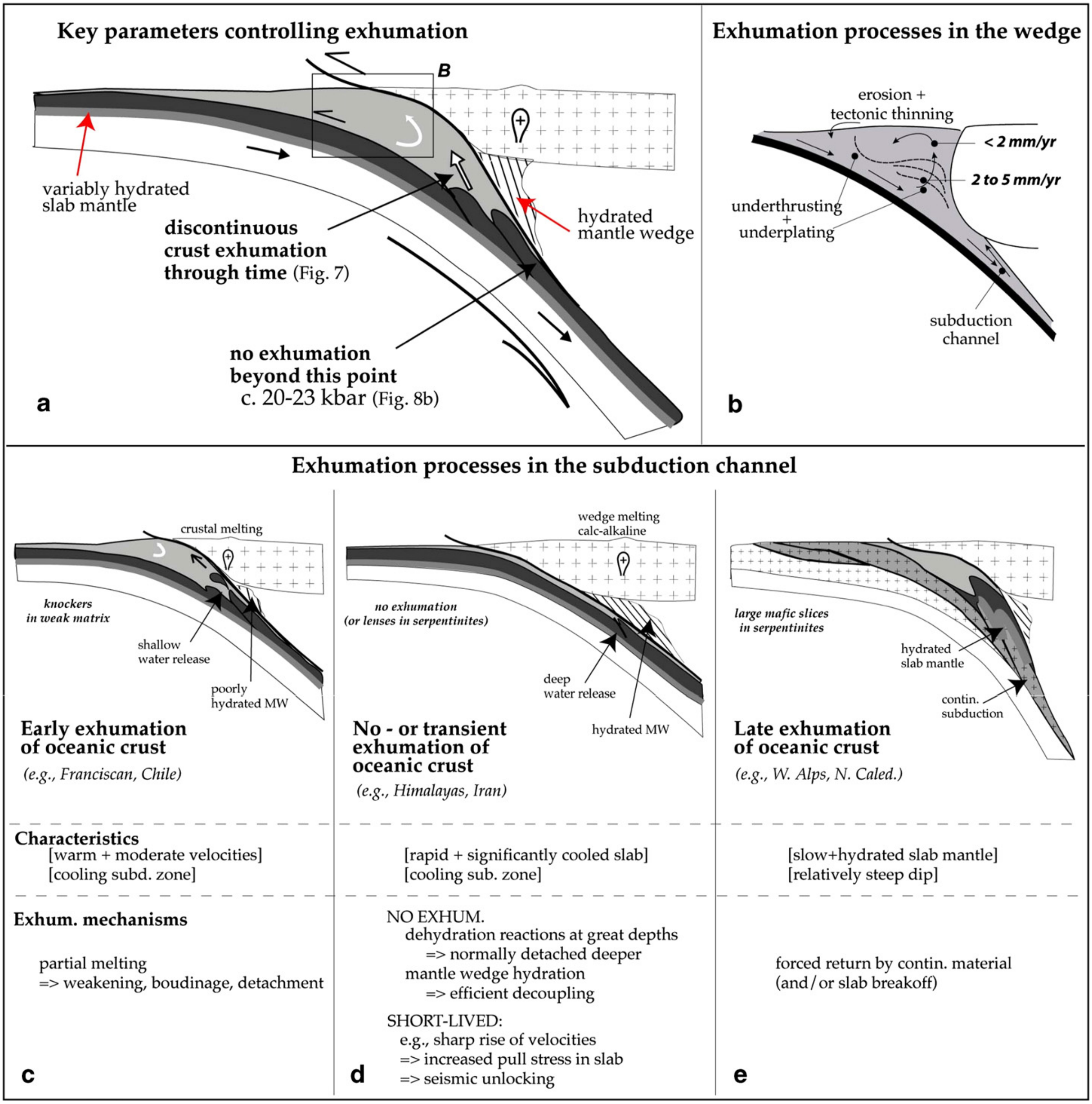

Fig. 14. Tentative model of processes controlling the exhumation of oceanic material in subduction zones. (a) In general no steady state exhumation of the oceanic crust takes place, and no exhumation is possible beyond a depth of ~70 km, corresponding to 2.0–2.3 GPa. Slab mantle and mantle wedge serpentinites play an important role on exhumation. (b) Sedimentary input is largely decoupled from the oceanic crust in the upper levels of the subduction channel and circulates in the accretionary wedge (dotted lines and dots: successive position of sedimentary interfaces). (c) Early exhumation, in the first 5–15 My of oceanic subduction may occur for relatively warm, rapidly cooling subduction zones. We propose that exhumation takes place through weakening resulting from the partial melting of the oceanic crust (Sorensen and Barton, 1987), leading to boudinage and detachement of high grade blocks in a weaker, sedimentary melange. (d) No steady-state exhumation of oceanic material takes place once the subduction has cooled (see text for details). Incidental exhumation, however, may happen in response to a major modification of boundary conditions (Himalayas, Zagros, Ecuador). (e) Late exhumation may be associated with continental subduction (and/or slab breakoff) in relatively slow, steeply dipping subduction settings with a highly hydrated slab mantle (such as the W. Alps).

1999). Mafic knockers were classically considered to have equilibrated at $P<1.2–1.5$ GPa, but recent findings point to pressures in excess of 20 kbars and exhumation velocities possibly as high as 5 mm/yr (Tsujimori et al., 2006). The Franciscan subduction zone is known to have cooled during a transient period which lasted ~10–15 My (Peacock, 1987; Anczkiewicz et al., 2004). Radiometric constraints for the high-grade blocks of oceanic crust suggest, as for Chile, that they were only exhumed during these early subduction stages (Cloos, 1985; Anczkiewicz et al., 2004).

A.2.2. Santa Catalina

This island corresponds to a small part of a later accretionary wedge formed after an outboard jump of the Franciscan subduction zone (Platt, 1986; Sorensen, 1988; Bebout, 1991; Grove and Bebout, 1995; Fig. 7b), and preserves evidence of transient, cooling metamorphic gradients associated with the onset of subduction (Peacock 1987). Age constraints suggest exhumation velocities from depths of 30–40 km at 1–2 mm/yr (Grove and Bebout, 1995).

A.2.3. Cascades

This Miocene accretionary wedge is only briefly alluded to here as a modern analogue of the Franciscan complex. Based on fission-track data, exhumation appears to be mostly erosion-driven (Brandon et al., 1998) and steady-state accretion and particle flow within the wedge, returning 80–90% of the sedimentary input, were deduced for the last 15 My (Batt et al., 2001). No deeply seated rocks have been exhumed so far.

A.3. Iran

A.3.1. Zagros

The Tertiary collisional Zagros orogen formed after a long-lived subduction of the Neotethys below Eurasia (~150–35 Ma; Agard et al., 2005b). HP–LT remnants are scarce, however, as for the Himalayas (see below). Blueschist facies rocks are only found in the south-easternmost corner of Zagros, near Makran (Fig. 7c; Sabzehei, 1974; Agard et al., 2006). They represent elongate hm- to km-sized mafic and volcanoclastic bodies, intimately associated with serpentinites, dispersed in a 'coloured' melange zone. Some serpentinite units contain lawsonite–omphacite–glaucophane-bearing blueschist knockers pointing to deeper *P*–*T* conditions (c. 1.8 GPa/500 °C). Age constraints point to a short-lived exhumation during the period 95–85 Ma, which coincided with regional-scale obduction movements (Coleman, 1971; Ricou, 1971) and followed a sharp rise of convergence velocities (Agard et al., 2006, 2007). This period also coincided with blueschist exhumation along adjacent segments of this subduction zone, in Turkey (Sherlock et al., 1999), Makran (Delaloye and Desmons, 1980) and Pakistan (Anczkiewicz et al., 2000).

A.3.2. Makran

Several blueschist facies outcrops of dominantly mafic material have been reported (McCall, 1997) at the rear of the Tertiary accretionary wedge (Platt et al., 1985). Radiometric dating pointed to ages of 87.9 ±5 Ma (Delaloye and Desmons, 1980) but *P*–*T* constraints are lacking.

A.3.3. Sistan (Eastern Iran)

Two serpentinite-rich accretionary complexes resulting from the closure of a small branch of the Neotethys (<600 km; Tirrul et al., 1983) are found in Eastern Iran. BS and eclogites, which form small lenses (metric to 200×100 m) exhibiting variable *P*–*T* conditions (Fotoohi Rad et al., 2005), were exhumed during convergence at least 20 My before collision (Tirrul et al., 1983). The lack of radiometric constraints unfortunately precludes the calculation of exhumation velocities for this geodynamic setting.

A.4. New Caledonia

Blueschist and eclogites found in New Caledonia represent one of the most extensive such outcrop in the world (Fig. 7d; Yokohama et al., 1986; Clarke et al., 1997). Its detailed structure and isograds, however, are still disputed (Rawling and Lister, 2002; Fitzherbert et al., 2005). Diahot blueschists and incipient eclogites correspond to clastic metasediments associated with an oceanic melange, whereas the eclogitic Pouébo terrane is dominated by mafic and volcanoclastic rocks intimately associated with serpentinites (Cluzel et al., 2001; Rawling and Lister, 2002; Fitzherbert et al., 2004, 2005). Mafic blocks in serpentinites preserve evidence for a 100–150 °C heating thought to result from the contact with the mantle wedge (Fitzherbert et al., 2004). This rather short-lived subduction zone (<15–25 My) is thought to have experienced the incorporation of several terranes (Diahot, Pouébo and Poya units) before choking when the continental Norfolk ridge attempted subduction (Cluzel et al., 2001; Fitzherbert et al., 2004; Spandler et al., 2005). This event resulted in the fore-arc obduction and exhumation of the HP–LT rocks (Cluzel et al., 2001), which later suffered contractional refolding (Rawling and Lister, 2002).

A.5. Antilles

The Antilles is a complex geodynamic setting with evidence of several subduction zones associated with the convergence between North and South America, whose timing and dip directions are still debated, either for Cuba (Iturralde-Vinent, 1994; Kerr et al., 1999; Schneider et al., 2004; Cobiella-Reguera, 2005; Garcia-Casco et al., 2006; Stanek et al., 2006) or for the Dominican Republic (Goncalves et al., 2000). These subductions have given rise to a variety of settings, from extensive, >500 km long serpentinite-rich zones with eclogite blocks (the so-called Northern serpentinite melange; Fig. 7e; Garcia-Casco et al., 2002), to highly deformed nappe stacks grading from blueschist to eclogite facies conditions, with clastics and mafic bodies wrapped in serpentinites (Escambray: Schneider et al., 2004; Stanek et al., 2006).

Contrasting *P*–*T* regimes are found, ranging from high temperature regimes associated with nascent subduction zones (for example Eastern Cuba; Garcia-Casco et al., 2006), to cold regimes associated with lawsonite–eclogite forming subduction zones (Dominican Republic; Zack et al., 2004; Krebs et al., 2008). One of the peculiar aspects of the subduction record is the evidence for transient thermal (and/or mechanical) instabilities in the subduction process, inferred from oscillatory zoning in garnets (Garcia-Casco et al., 2002, 2006).

A.6. The Himalayas

Sparsely distributed blueschists, mostly metavolcanics and volcanoclastics with minor serpentinites, crop out along the Indus–TsangPo suture zone, in the vicinity of the NW Himalayan syntaxis (Shangla: Anczkiewicz et al., 2000; Sapi-Shergol: Honegger et al., 1989). Age constraints point to exhumation periods similar to the one reported above for SE Zagros, at around 90–80 Ma (Maluski and Matte, 1984; Tonarini et al., 1993; Anczkiewicz et al., 2000). Compared to the vast amount of subducted oceanic lithosphere, the Himalayas have preserved only a very small percentage of HP–LT rocks. Besides, eclogites reported so far correspond exclusively to the metamorphosed continental Indian crust dragged below Asia and exhumed in the NW Himalayas (Pognante and Spencer, 1991; Guillot et al., 1997, 1999; De Sigoyer et al., 2000; O'Brien et al., 2001; Guillot et al., 2007), apart from one mafic eclogite from the Indus suture zone whose significance is still unclear (Le Fort et al., 1997).

A.7. Sambagawa

The Sanbagawa belt represents the fossil HP–LT part of one of Japan's paired metamorphic belts (Miyashiro, 1961; Banno, 1986; Takasu et al., 1994). It is mostly made of high grade blueschists and comprises rare eclogitic blocks encased in a mainly epidote–amphibolite weak pelitic matrix (Takasu et al., 1994; Enami, 1998), with only minor serpentinites. Recent petrological studies suggest, however, that eclogitic conditions may neither be as restricted as previously thought and may not be confined to isolated mafic blocks only (Wallis and Aoya, 2000; Ko et al., 2005). Whether this belt represents or not an accretionary melange is debated (Takasu et al., 1994; Wallis, 1998). *P*–*T* estimates yield 2.0–2.1 GPa and 600 °C in the Beshi area for isolated blocks and interleaved metapelites (Ko et al., 2005). Similar *P*–*T* values but contrasting exhumation paths were obtained in the Kotsu area (Matsumoto et al., 2003). Exhumation velocities are ≤1 mm/yr (Wallis et al., 2004), although few radiometric constraints are available, unfortunately, for the HP stages.

A.8. Others

A.8.1. Sulawesi

The Bantimala complex represents the exhumed wedge of a Cretaceous subduction and is mainly made of mafic bodies and volcanoclastics, with eclogite blocks embedded in serpentinites. Ultramafic units occupy an internal position. Miyazaki et al. (1996) inferred a cold *P–T* gradient of 8 °C/km and a cooling exhumation path.

A.8.2. Andes

The case of the Andes is considered here as an example where almost no exhumation of oceanic material takes place during convergence (Central Andes; Jaillard et al., 2002) despite active subduction along more than 4000 km from the Jurassic onwards. Accreted terranes testifying to restricted exhumation episodes are found, however, in the Northern Andes (Ecuador; Feininger, 1980; Kerr et al., 2002; Gabriele et al., 2003; Venezuela: Stöckhert et al., 1995; Maresch et al., 2000). The exhumation of the Raspas complex of Ecuador, in particular, is thought to result from the entrance of an oceanic plateau jamming the subuction zone during the lowermost Cretaceous (Arculus et al., 1999; Bosch et al., 2002).

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