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Creation, Preservation, and Exhumation of UHPM Rocks

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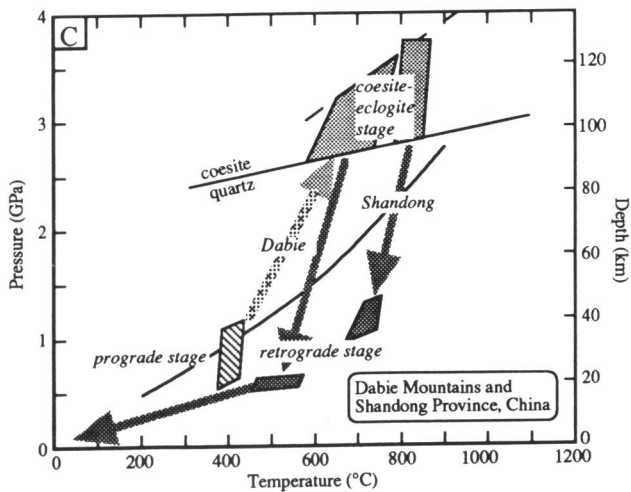
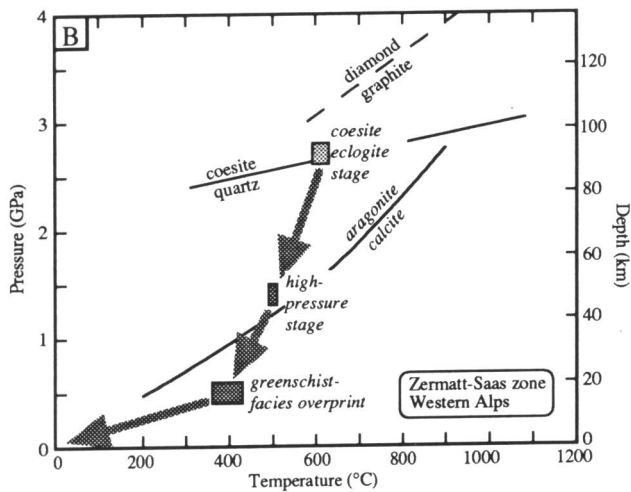
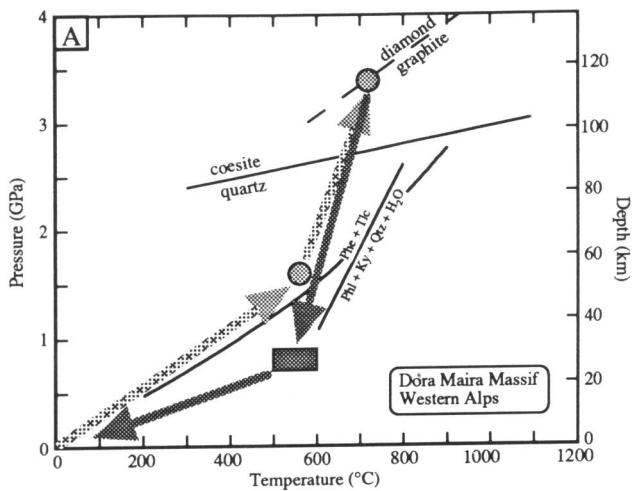
Abstract

Coesite-bearing eclogites exposed in the Western Alps and China record peak metamorphic temperatures of 550–900°C at pressures ≥ 2.5 GPa (≥ 25 kbar). Their presence as regionally metamorphosed rocks requires subduction of >400 km² of upper crustal material to depths ≥ 90 km and subsequent exhumation. Existing experimental kinetic data suggest that pure coesite rock should not survive exhumation in the presence of a fluid, and this is corroborated by field observations. Partial survival of coesite is linked to its occurrence as inclusions in porphyroblasts that can maintain high internal pressures and prevent ingress of fluids. In the absence of coesite, reliable indicators of its former presence include subparallel, sometimes curving quartz subgrains and sets of quartz subgrains that truncate other sets. Polycrystalline quartz aggregates and cracks radiating outward from inclusions are not *a priori* evidence of ultrahigh pressure (UHP) metamorphism.

Petrologic constraints demand that coesite-bearing regional metamorphic rocks cooled during exhumation. Cooling during exhumation requires either (1) continued subduction beneath the eclogites that effectively chills the overlying UHP rocks during exhumation, or (2) transport toward the surface in the lower plate of an extensional structure such as a low-angle normal fault or shear zone. Both processes may occur; neither process demands rapid exhumation rates, although radiometric dating of different portions of the Dora-Maira pressure-temperature path indicate average long-term exhumation rates of 3.00 km Ma⁻¹.

Introduction

Coesite is a high-pressure polymorph of SiO₂ stable at pressures ≥ 2.5 GPa for metamorphic temperatures ≥ 500 °C (Fig 5.1). In 1984, Chopin (1984) and



Smith (1984) reported the discovery of coesite in regional high-pressure metamorphic rocks in the western Alps and Norway, respectively. The discovery in China of coesite in metasedimentary rocks (Wang et al., 1989) of regional extent ($>400 \text{ km}^2$) demonstrates that rocks deposited at the surface are subducted to depths $\geq 90 \text{ km}$ and subsequently exhumed by erosional and tectonic processes.

In this chapter, we first review the geologic and petrologic settings of coesite occurrences in the Western Alps and China. We then examine natural coesite \rightarrow quartz reaction textures and the petrologically determined pressure-temperature (P - T) paths in light of existing kinetic data and thermal models. In this review, we document the following conclusions:

1. The coesite-bearing eclogites record relatively low temperatures at high pressures, consistent with formation in a subduction zone environment.
2. During exhumation, complete reversion of coesite to quartz can be prevented only if the coesite occurs as an inclusion within a porphyroblast capable of maintaining high internal pressures and excluding fluids that would otherwise catalyze retrogression.
3. Coesite-bearing eclogite terrains record cooling during exhumation which requires that exhumation occurred while subduction/underflow continued to chill the overlying lithosphere or that the terrains were transported to the surface in the lower plate of extensional structures.

Geologic and Petrologic Setting

There are at least six worldwide occurrences of UHP rocks, but we focus on the better described samples from the Western Alps and China. For more detailed descriptions of these rocks and the methods used to evaluate their metamorphic history, we refer the reader to other chapters in this book.

Western Alps

Coesite occurs in the Dora-Maira massif (Chopin, 1984) and in the Zermatt-Saas zone (Reinecke, 1991) of the Western Alps. The Dora-Maira massif is a

Figure 5.1. Pressure-temperature (P - T) paths for coesite-bearing eclogites from (A) Dora-Maira massif, Western Alps (Chopin, 1984; Schertl et al., 1991; Sharp et al., 1992); (B) Zermatt-Saas zone, Western Alps (Reinecke, 1991); and (C) Sino Korean-Yangtze collision zone, China (Enami and Zang, 1990; Wang and Liou, 1991, 1992). Quartz-coesite reaction from Bohlen and Boettcher (1982), calcite-aragonite reaction from Hacker et al. (1995), and graphite-diamond reaction from Bundy (1980). Phen-gite + talc = phlogopite + kyanite + quartz + H_2O reaction from Schertl et al. (1991). Depth-pressure relations based on a density of 3000 kg/m^{-3} .

domal internal crystalline massifs of the Penninic domain and may represent the southeastern edge of the European plate or a microcontinent (Platt, 1986). The coesite-bearing rocks occur as boudins in a "polymetamorphic" series that is bounded above and below by clastic Paleozoic–Mesozoic rocks also metamorphosed at high pressure (Chopin et al., 1991). The polymetamorphic series consists of the coesite-bearing unit and overlying eclogite-facies "micaschiste amygdalaire" and "cold-eclogite" units (Chopin et al., 1991). An orthogneiss derived from Hercynian granites makes up the bulk of the coesite-bearing unit (Chopin et al., 1991; Tilton et al., 1991) and forms country rock to the blocks of coesite rock and a unit of paragneiss, mafic eclogite, and marble (Chopin et al., 1991). The coesite occurs in magnesian boudins perhaps derived from evaporites or by metasomatism (Schreyer, 1977).

Coesite crystals occur in kyanite and garnet, and quartz pseudomorphs after coesite occur in omphacite, kyanite, and garnet (Chopin et al., 1991). The coesite-bearing blocks reached pressures of 3.2–3.6 GPa and temperatures of 700–750°C (Massone and Schreyer, 1989; Chopin et al., 1991; Schertl et al., 1991; Sharp et al., 1993) (Fig 5.1A). Relict minerals in garnet cores indicate that the prograde P - T path passed near $P \approx 1.5$ GPa and $T \approx 600^\circ\text{C}$, precluding the possibility of granulite-facies metamorphism immediately prior to the UHP metamorphism (Chopin et al., 1991). Because coexisting talc and phengite remained stable during decompression and did not break down to the low-pressure assemblage phlogopite + kyanite + quartz + H_2O , temperatures during decompression must have remained below 750°C at $P \approx 2.0$ GPa and below 600°C at $P \approx 1.0$ GPa (Schertl et al., 1991).

Clastic rocks beneath the polymetamorphic series reached metamorphic conditions of $P < 1.0$ GPa and $T \approx 500^\circ\text{C}$ (Chopin et al., 1991). The "cold eclogite" and "micaschiste amygdalaire" that make up the rest of the polymetamorphic series reached $P \approx 1.5$ GPa and $T = 500$ – 550°C (Chopin et al., 1991). Clastic rocks overlying the polymetamorphic series reached $P \approx 1.0$ – 1.2 GPa and $T \approx 500^\circ\text{C}$ (Chopin et al., 1991). Late deformation under greenschist-facies conditions, perhaps at $P \approx 0.7$ – 0.9 GPa and $T \approx 500$ – 600°C , caused intermixing of units and variable development of a foliation and lineation (Chopin et al., 1991; Schertl et al., 1991).

There is conflicting information about the activity of H_2O during the Alpine UHP metamorphism. Massone (1990) used the Al_2O_3 content of talc to infer that $a_{\text{H}_2\text{O}}$ was ≥ 0.8 . Probable partial melting textures in the country rock gneiss have also been used to suggest that the activity of H_2O was near unity (Schertl et al., 1991). Note, however, that H_2O partitions into silicate melt at high pressures, and thus partial melting will reduce the H_2O activity in solid phases coexisting with melt. Rossman et al. (1989) used infrared spectropho-

tometry to infer that the pyrope contains 0.002–0.003 wt% hydroxyl, a measurement that reportedly implies garnet growth at low H_2O activity. The preservation of UHP assemblages outside their stability field also argues for low aqueous fluid activity. On the basis of thermodynamic calculations of the position of the reaction pyrope + coesite + H_2O = kyanite + talc and temperatures estimated from oxygen isotope measurements, Sharp et al. (1993) inferred that a_{H_2O} was 0.4–0.75. Moreover, they proposed that the fluid diluent was silicate melt (and specifically not CO_2).

It is uncertain whether the UHP metamorphism in Dora-Maira occurred at ≈ 100 Ma or ≈ 40 Ma. High-pressure metamorphism in the Alps is generally recognized to have occurred ≈ 100 Ma (e.g., Hunziker et al., 1989). Dora-Maira coesite quartzite has yielded ≈ 100 Ma phengite $^{40}Ar/^{39}Ar$ ages (Monié and Chopin, 1991) and ≈ 121 Ma U-Pb zircon ages (Paquette et al., 1989). UHP rocks overprinted by greenschist-facies assemblages, as well as structurally adjacent greenschist-facies units, yield $^{40}Ar/^{39}Ar$ phengite ages of 30–50 Ma (Scaillet et al., 1990; Monié and Chopin, 1991). These data suggest that UHP metamorphism occurred at ≈ 100 Ma and was followed by retrogression at ≈ 40 Ma. In contrast, comprehensive Sm-Nd and U-Th-Pb dating by Tilton et al. (1991) produced markedly younger, 38–30 Ma, ages for the UHP metamorphism. Zircon inclusions in four pyrope crystals (stable only at pressures greater than ≈ 1.2 GPa; Schreyer, 1988) and one pyrope quartzite yielded a chord with a lower intercept age of ≈ 38 Ma (Tilton et al., 1991). Ellenbergerite (a phase stable only at pressures > 2.0 GPa; Chopin, 1986) and monazite from the same rock types yielded U-Pb and Th-Pb ages of 30–34 Ma (Tilton et al., 1991). Sm-Nd dating of 13 pyrope, zircon and phengite crystals produced an errorchron of 38 Ma (Tilton et al., 1991). Other ultrahigh-pressure rocks in the Alps from the Cima Lunga nappe have yielded five Sm-Nd garnet-pyroxene-whole rock ages of 38–42 Ma (Becker, 1993) and two U-Pb zircon ages of 32 and 36 Ma (Gebauer et al., 1991). Thus, there is an apparent 60 Ma discrepancy among different radiometric attempts to date the UHP metamorphism. The discrepancy may be attributable to an effect of deformation, pressure, phase chemistry, or some other factor on isotopic system(s) in one or more high-pressure phase(s). Given the increasing recognition that phengite may incorporate excess ^{40}Ar (e.g., Hacker and Wang, in review), we favor the Sm-Nd and U-Th-Pb ages of ≈ 38 Ma.

Coesite also occurs in the Zermatt-Saas zone, the uppermost part of the Pennine nappes, which consists of HP meta-sedimentary and meta-ophiolitic rocks representing portions of the Tethyan ocean basin subducted southward beneath the Apulian (African) plate in late Mesozoic time. Coesite inclusions occur in garnet and tourmaline in piemontite-bearing quartzite of the oceanic

“schistes lustrés” (Barnicoat and Fry, 1986). Conditions for coesite crystallization are estimated as $T \approx 600^\circ\text{C}$ and $P \approx 2.7$ GPa. The retrograde P - T path is judged to have passed first through $T \approx 500^\circ\text{C}$ and $P \approx 1.4$ GPa, and then through $T \approx 400^\circ\text{C}$ and $P \approx 0.5$ GPa (Reinecke, 1991) (Fig 5.1B).

China

Coesite-, aragonite-, and diamond-bearing eclogite occur in the collision zone between the Yangtze and Sino-Korean cratons in China (Ernst et al., 1991; Wang and Liou, 1991; Xu et al., 1992). Mafic eclogite occurs as blocks and rare layers in marble, schist, and quartzofeldspathic gneiss (Wang and Liou, 1991). Coesite pseudomorphs occur in garnet in the latter three rock types, and in omphacite and epidote in eclogite (Wang et al., 1990; Wang and Liou, 1991; R. Zhang, personal communication, 1992). Peak pressures and temperatures for coesite-bearing rocks are estimated as $T \approx 740$ – 840°C , $P > 2.8$ GPa on the Shandong Peninsula at the eastern end of the collision zone (Enami and Zang, 1990; Hirajima et al., 1990) (Fig 5.1C), and $T \approx 550$ – 700°C , $P > 2.6$ – 2.8 GPa in the Dabie Mountains farther west in the collision zone (Wang et al., 1990). Partial melting of gneiss in the northern Dabie Mountains may indicate that the H_2O activity was near unity at ultrahigh pressures or the partial melting may be related to decompression following the UHP metamorphism. Retrograde parageneses indicate amphibolite-facies and greenschist-facies overprints (Wang and Liou, 1991) at $T \approx 500^\circ\text{C}$ and $P \approx 0.6$ GPa in the Dabie Mountains (Wang et al., 1990), and $T \approx 670$ – 750°C and $P \approx 0.9$ – 1.4 GPa in Shandong Province (Enami and Zang, 1990). These rocks also contain partially transformed aragonite inclusions in garnet and omphacite (Wang and Liou, 1991; 1993) and diamond inclusions in garnet (Xu et al., 1992). Zircons from UHP gneisses yield U-Pb ages of ≈ 209 Ma (Ames et al., 1993), and hornblende, muscovite and biotite record cooling through 500 – 300°C by 180 Ma (Hacker and Wang, in review). Hornblende barometry on metamorphism associated with emplacement of Cretaceous plutons (≈ 105 – 135 Ma; Li and Wang, 1991) suggests crystallization at pressures of ≈ 500 MPa (Liou et al., 1992). Thus, the UHP rocks reached depths as shallow as ≈ 15 km by ≈ 125 Ma.

Regional Extent of UHP Metamorphism

The distribution of key mineral parageneses and phase compositions indicates that the area metamorphosed at ultrahigh pressure in China exceeds 400 km². UHP metamorphism may have occurred on a similar scale in the Alps, even

though the outcrop area is limited by overlapping postorogenic sediments. In the Dora-Maira massif, coesite-bearing boudins are distributed over an area of 10×15 km. Jadeite pseudomorphs and garnet compositions suggest that the gneisses containing the coesite-bearing blocks were also metamorphosed at UHPs (Chopin et al., 1991). Tectonic models for these areas must account for exhumation of "blocks" exceeding 20 km in two dimensions. In the Zermatt-Saas zone, where the coesite quartzite unit is only several hundred meters long (Reinecke, 1991), further work is needed to determine the regional extent of UHP metamorphism.

Coesite \rightarrow Quartz and Aragonite \rightarrow Calcite Retrogression Textures

Silica inclusions in garnet from UHP regional metamorphic rocks are as large as $300 \mu\text{m}$ (Smith, 1984; Enami and Zang, 1990; Reinecke, 1991). Partially preserved coesite inclusions are invariably single crystals. The amount of retrogression from coesite to quartz varies from ≈ 25 to 100 vol%. Partially retrogressed coesite inclusions are surrounded by as many as several thousand elongate quartz grains (ranging down to less than $1 \mu\text{m}$ wide) arranged radially in an ellipsoidal shell around the coesite (Chopin, 1984; Smith, 1988). The quartz grains occur in groups of similarly shaped crystals with long axes subperpendicular to the host coesite crystal grain boundary (Fig 5.2). Individual thin sections show not only relict coesite crystals surrounded by thousands of quartz grains, but also coesite pseudomorphs composed of tens to only a few coarse grains (Smith, 1984). These textures are taken to indicate that after a host coesite crystal was consumed, the quartz grains coalesced and coarsened (Smith, 1988).

Ingrin and Gillet (1986) reported a lack of topotaxy between coesite and quartz from the Dora-Maira massif, based on transmission electron microscopy (TEM). We measured, with a universal-stage microscope, the *c*-axis orientations of quartz crystals grown on the largest coesite crystal reported by Wang and Liou (1991; their Figs 5.2A and 5.2B). The fine grain size and pervasive undulatory extinction make measurement difficult, but the *c*-axes of measurable grains are neither orthogonal to the host grain boundary nor parallel to the quartz crystal long axes. Thus, the crystallographic orientation of the quartz crystals appears not to have been controlled by the host garnet and coesite crystals, and the shape of the quartz crystals is related chiefly to the retreat of the unstable coesite crystal rather than to crystallographic controls on growth rate. The high density of atopotactic quartz crystals likely reflects a high nucleation to growth rate ratio. Such rate ratios are favored during transformations by rapid changes in reaction driving potential (Rubie,

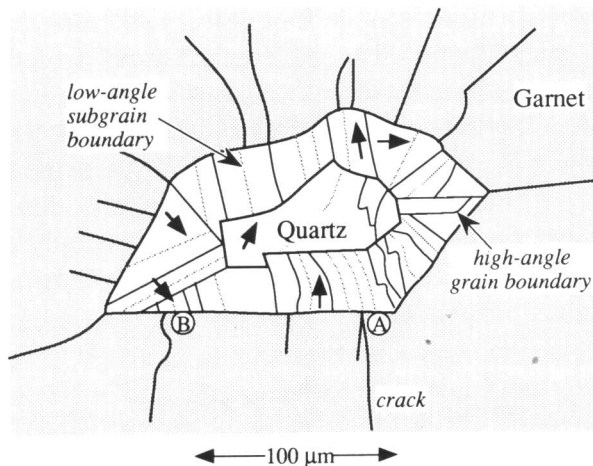


Figure 5.2. Line drawing of garnet (gray) including coesite pseudomorph retrogressed to quartz (unlabeled). Solid lines within garnet are cracks formed during expansion of the inclusion. Solid lines within quartz are high-angle grain boundaries and pale lines are low-angle subboundaries. Most grain boundaries are straight, but a minority are gently or sharply curved (A). Note the areas of grain impingement near the corners of the garnet grain boundary, where grains of one orientation truncate grains of another orientation (B). Not shown is a pervasive undulatory extinction within the quartz grains. Small arrows indicate traces of *c*-axes of grains measured with a u-stage microscope; the *c*-axes show no consistent relationship with the host coesite-garnet grain boundary. This is sample MW37 of Wang and Liou (1991; see their fig. 2A, 2B).

1983). The coesite \rightarrow quartz reaction has relatively low $\Delta P/\Delta T$, thus, changes in potential are more easily effected by decompression, rather than by heating. If decompression caused coesite \rightarrow quartz reaction, it might have been the result of erosional or tectonic removal of overlying rocks or cracking of the enclosing garnet crystal.

Aragonite single-crystal inclusions in garnet and omphacite are surrounded by aggregates of calcite crystals, much like the textures associated with the breakdown of coesite to form quartz (Wang and Liou, 1993). The calcite occurs as fine-grained aggregates surrounding single aragonite crystals.

Minerals that host inclusions of coesite and aragonite invariably contain radiating cracks (Fig 5.2; Wang and Liou, 1993). These radiating cracks are typically inferred to have formed during the polymorphic phase transformation to form the low-pressure polymorph quartz or calcite (van der Molen and Paterson, 1986). However, the expansion may instead be a result of differential volume change of the host and inclusion phases during decompression or cooling (Wendt et al., 1993). In the absence of coesite or other supporting evidence, radial cracks emanating from quartz inclusions do not mandate UHP metamorphism. The most reliable indicators of the earlier presence of

coesite are the textures illustrated in Fig 5.2, particularly the subparallel, sometimes curving quartz subgrains (Fig 5.2 label A) and the truncation of one set of subgrains by another (Fig 5.2 label B).

Experimental Kinetic Studies

There have been a modest number of experimental studies on the equilibrium boundary and rate of the quartz-coesite transformation (see Bohlen and Boettcher, 1982, for a review); however, these data are not generally applicable to natural metamorphism. The physical state under which quartz grew from coesite in these experiments differs substantially from nature. In nature, the transformation was of coesite single crystals enclosed in single-crystal garnet, whereas the laboratory experiments were conducted on powdered coesite often in the presence of powdered quartz and H₂O. Powdering results in an experimental transformation rate that is faster than the natural transformation, and can provide only an upper bound on the natural case (e.g., Rubie and Thompson, 1985; Hacker et al., 1992).

However, experimental data do suggest that the coesite → quartz reaction is sluggish in the absence of water; Dacheville et al. (1963) found that dried, powdered coesite persisted for at least 550 hours when heated at 1070°C and room pressure. Transformation experiments on more “natural” material must be conducted to provide data that can be applied quantitatively to the retrogression and exhumation of coesite-bearing rocks. One possibility is to conduct experiments on coesite single crystals contained within garnet crystals, but synthesizing such material is fraught with difficulty.

Significantly better experimental data are available for the transformation of aragonite to calcite. Carlson and Rosenfeld (1981) measured the growth rate of calcite from aragonite single crystals at 1 atm, and Carlson (1983) calculated a nucleation rate from Kunzler and Goodell's (1970) experiments on powdered aragonite. Extrapolation of Carlson's data to UHP metamorphic rocks meets with difficulty, however. Even for small reaction free energies (<1 kJ mole⁻¹), Carlson's data predict that 1000 μm aragonite grains will revert completely to calcite in less than 1 Ma at temperatures as low as 200°C. In obvious conflict with this is the local preservation of aragonite in UHP rocks in China that evidently entered the calcite stability field at temperatures in excess of 500°C (Wang and Liou, 1993). Fundamental differences must therefore exist between the experimental configuration and the natural situation. Possible differences include:

1. The retrograde *P-T* path following UHP metamorphism may have been such that the free energy potential to drive nucleation and growth was very

small or zero. In other words, the depressurization and cooling path may have followed the conditions of the calcite–aragonite equilibrium boundary to low temperatures. Calculations based on Carlson's data suggest that reaction free energies of 100 J are sufficient to drive complete transformation in less than 1 Ma at temperatures of 200°C. Such a small reaction free energy means that the pressure during decompression cannot have deviated from the equilibrium pressure by more than ≈ 50 MPa. However, field study (Wang and Liou, 1991, 1993) indicates that retrogression occurred in the calcite stability field at pressures ≈ 0.5 GPa below the equilibrium boundary and at temperatures of $\approx 500^\circ\text{C}$.

2. The natural nucleation or growth rates may be very different from the experimental situation, because the natural aragonite crystals are enclosed as single crystals within host garnet and omphacite single crystals. A large rate difference is plausible because nucleation in powders and single crystals is controlled by the free energy of the aragonite crystal surface in contact with vapor or fluid, whereas nucleation in the natural case must occur at the crystalline interface between the aragonite and host crystal. Surface and grain-boundary free energies differ by an order of magnitude in metals (Murr, 1975; Porter and Easterling, 1981), and similar differences between aragonite–fluid and aragonite–garnet interfaces would yield notable differences in transformation kinetics. Evaluating the importance of such effects is experimentally possible but difficult.

3. The transformation of aragonite to calcite involves dilatation, and the host mineral may act as a pressure vessel that confines the expanding inclusion. The aragonite \rightarrow calcite transformation can be inhibited if the pressure buildup causes the inclusion to remain in its HP stability field, even while the outside of the host crystal is at substantially lower pressure (Gillet et al., 1984; van der Molen and van Roermund, 1986).

Gillet et al. (1984) and van der Molen and van Roermund (1986) postulated that the transformation of coesite to quartz was inhibited during cooling and decompression because the garnet surrounding the coesite acted as a pressure vessel that maintained high pressure on the silica inclusion even as the rock pressure and temperature moved well into the quartz stability field. Modeling isotropic, linear elastic behavior of garnet and coesite, they calculated the rate of quartz formation based on the assumption that any increase in inclusion pressure required to stay out of the quartz stability field was instantly achieved by the volume increase produced when coesite transformed to quartz. In spite of the assumptions required to make the problem tractable, the models lead to the interesting prediction that the pressure vessel effect restricts the coesite \rightarrow quartz retrogression to only moderate amounts (≈ 25 – 30 vol%) before the garnet fractures. Van der Molen and van Roermund (1986) calculated that

when the temperature has decreased to $\approx 400^\circ\text{C}$, the inclusion pressure will have built up to more than three times the external pressure, and the garnet may fracture, allowing ingress of fluid and more complete retrogression. Their model requires that the host crystal be able to sustain tensile stresses of ≈ 500 MPa at temperatures of $\approx 500^\circ\text{C}$. Of the phases that include coesite and aragonite, only the mechanical behavior of clinopyroxene has been investigated at high temperatures. Kollé and Blacic (1982) found that clinopyroxene single crystals can deform by twinning at significantly lower differential stresses, about 100–140 MPa, at strain rates of 10^{-4} – 10^{-8} s^{-1} and temperatures of 500–1000°C. This implies that clinopyroxene crystals might not be able to preserve coesite inclusions, but would instead deform by mechanical twinning.

P-T Paths and Thermal Evolution

The occurrence of coesite-bearing eclogite derived from sedimentary protoliths requires (1) subduction of crustal material to depths of ≈ 100 km, (2) detachment from the downgoing slab (transfer to the overriding plate), and (3) exhumation by a combination of erosional and tectonic processes.

Prograde P-T Path

Coesite-bearing eclogites record relatively cool temperatures of 550–900°C for depths of 90–125 km (Fig 5.3A). Typical geotherms from stable continental and oceanic settings suggest that temperatures exceed 1000°C at 100 km depth (Pollack and Chapman, 1977). In convergent plate margins, temperatures at depth can be substantially cooler than in intraplate settings because the subduction of oceanic lithosphere carries cold near-surface rocks downward, depressing isotherms at depth. The peak metamorphic pressures and temperatures recorded by coesite-bearing eclogites are consistent with *P-T* paths calculated for subduction zones (e.g., Peacock, 1990, 1993).

In thermal steady state, the *P-T* path followed by the top of the subducting oceanic crust coincides with the *P-T* conditions along the subduction shear zone. In the absence of significant radiogenic heating, steady-state temperatures along a subduction shear zone are approximated by (Molnar and England, 1990, eqns. 16 and 23):

$$T = \frac{(Q_0 + \tau V) z_f / k}{S} \quad (5.1)$$

where T = temperature (K), Q_0 = heat flux into the base of the subducting lithosphere (W m^{-2}), τ = constant shear stress (Pa), V = convergence rate (m s^{-1}), z_f = depth to the fault (m), k = thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), and

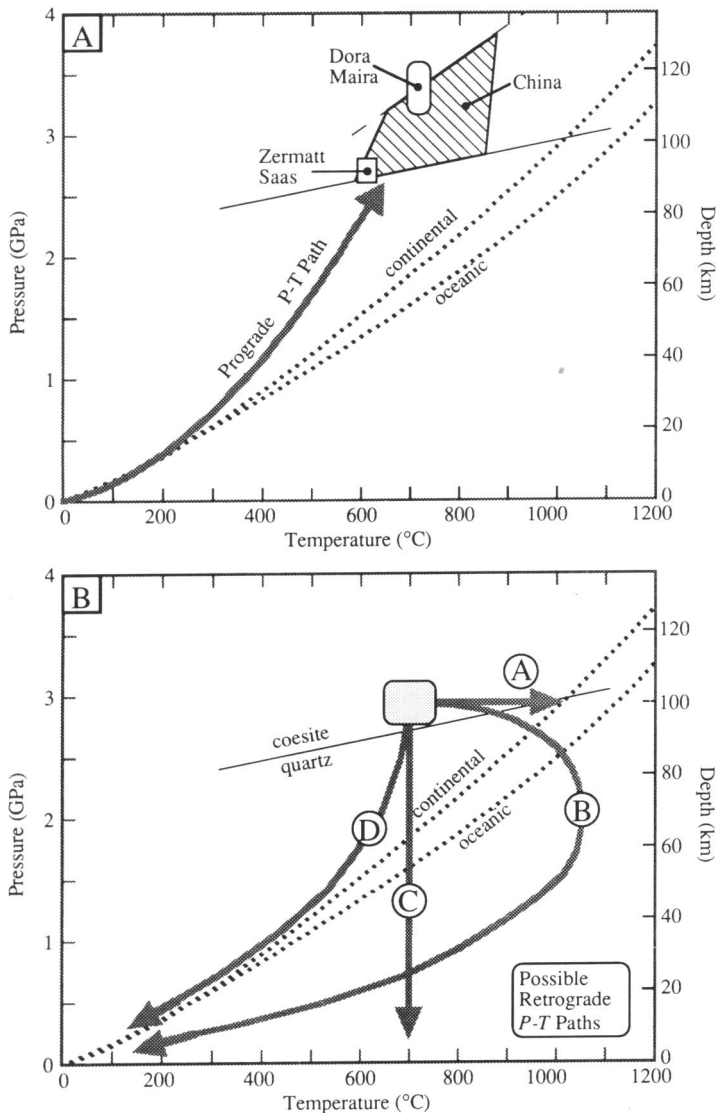


Figure 5.3. (A) Representative prograde subduction zone P - T path consistent with metamorphic conditions recorded by coesite-bearing eclogites; this path is one of a family of transient and steady-state subduction-zone P - T paths consistent with coesite-eclogite-facies metamorphism. This particular P - T path was calculated for a steady-state subduction zone with a convergence rate = $3 \text{ cm}\cdot\text{yr}^{-1}$, a constant shear stress = 80 MPa , a dip of 25° , and average thermal parameters. See text for discussion. (B) Four possible P - T paths subsequent to coesite-eclogite facies metamorphism at $P \approx 2.9 \text{ GPa}$ and $T \approx 700^{\circ}\text{C}$. Dotted lines represent average oceanic and continental geotherms [surface heat flow = 50 mW m^{-2} (for both cases), Pollack and Chapman, 1977]. Depth-pressure relation based on a density of $3000 \text{ kg}\cdot\text{m}^{-3}$. Path (A): No exhumation after peak metamorphism results in isobaric heating toward the steady state geotherm. Path (B): Exhumation at a moderate rate ($\approx 1 \text{ km}\cdot\text{Ma}^{-1}$) results in

S = a divisor that accounts for advection (heat transported by the subducting oceanic lithosphere):

$$S = 1 + b \sqrt{\frac{V z_f \sin \delta}{\kappa}} \quad (2)$$

where b = a constant (≈ 1 based on numerical experiments), δ = angle of subduction, and κ = thermal diffusivity ($\text{m}^2 \text{s}^{-1}$).

The prograde P - T path in Fig 5.3A represents the calculated steady state P - T path for the case of $Q_0 = 0.05 \text{ W m}^{-2}$, $\tau = 80 \text{ MPa}$, $V = 3 \text{ cm yr}^{-1}$, $k = 2.5 \text{ W m}^{-1} \text{ K}^{-1}$, $\delta = 25^\circ$, and $\kappa = 10^{-6} \text{ m}^2 \text{ s}^{-1}$. This representative P - T path is but one of a family of subduction zone P - T paths that pass through the P - T conditions recorded by coesite-bearing eclogites; similar steady-state P - T paths may be calculated for different combinations of convergence rate, shear stress, and other parameters.

Alternatively, the coesite-bearing eclogites may have formed during the early stages of subduction prior to the establishment of thermal steady state. At 100 km depth and $V = 3 \text{ cm yr}^{-1}$, achievement of thermal steady state requires several tens of Ma (see equations in Molnar and England, 1990). If coesite-bearing eclogites form during the first few tens of Ma of plate convergence while the subduction zone is still warmer than the steady-state thermal structure, then lower rates of shear heating (and lower shear stresses) are required to achieve the conditions of coesite-eclogite metamorphism. Shear heating may have been negligible (shear stresses ≈ 0), if the coesite-bearing eclogites formed during the first few million years of subduction while the hanging wall was still hot (Peacock, 1992, 1993).

Exhumation Mechanism

Exhumation of metamorphic rocks, defined as the displacement of rocks toward the Earth's surface (England and Molnar, 1990), requires removal of the rock overburden that existed at the time of metamorphism. Exhumation may involve sedimentary erosion, tectonic thinning, or some combination of the two processes.

Caption for Fig. 5.3 (cont.) moderate heating during decompression followed by cooling as the rocks approach the surface. Path (C): "Infinitely fast" exhumation results in nearly isothermal (adiabatic) decompression. Path (D): Exhumation occurring at moderate rates while subduction continues beneath the exhuming rocks results in cooling during decompression. Continued subduction chills the system, allowing rocks to be returned to the surface without heating toward the steady-state geotherm at high pressures. See text for discussion.

Large syn- to postorogenic Alpine sedimentary deposits document the importance of sedimentary erosion in exhuming Alpine metamorphic rocks. However, it is unlikely that sedimentary erosion alone could have been responsible for the removal of ≈ 100 km of overburden that existed above the Dora-Maira massif. The volume of Alpine syn- to postorogenic sedimentary deposits is dramatically smaller than that predicted by an erosion model (Platt, 1986), although some of the sedimentary record may have been subducted. Moreover, the pressure difference (≈ 1.5 GPa) between the coesite-bearing unit and adjacent units cannot be explained by local differential erosion but must record displacement between units.

Metamorphic rocks can also be exhumed by tectonic thinning of overburden. Extensional pure shear of the lithosphere displaces rocks toward the Earth's surface. In extensional simple shear, rocks in the footwall are exhumed as they are drawn out from beneath the hangingwall. Several workers have proposed that thrust faulting alone can exhume metamorphic rocks, but in the absence of erosion or tectonic thinning of the hangingwall, rocks in the upper plate do not approach the surface (i.e., they are not exhumed). In contrast to the views of some workers, exhumation of UHP rocks by extensional tectonics can occur within an active convergent plate margin. Platt (1986, 1992) has argued for extensional exhumation of high-pressure metamorphic rocks, pointing out that extension is occurring today in modern continental collision zones such as Tibet (Molnar and Lyon-Caen, 1989), and in modern accretionary wedges such as the Hikurangi prism of New Zealand (Walcott, 1987). Paleo-extensional structures have been reported from virtually every orogenic belt. Recently, Ballevre et al. (1990) and Blake and Jayko (1990) documented a west-dipping ductile normal fault overlying the Dora-Maira massif that appears to have accommodated some of the Dora-Maira exhumation in mid-Tertiary time, and Avigad (1992) similarly interpreted data collected by Henry (1990). Wheeler (1991) reported that thrusting may have occurred during exhumation as well, as HP rocks in the Dora-Maira massif are in thrust contact with underlying lower pressure rocks. Fossen and Rykkelid (1992) proposed that exhumation of Norwegian eclogites occurred partly by subhorizontal extension. Thus, tectonic thinning by subhorizontal extension, perhaps with the aid of thrusting at deeper levels and erosion above, seems the most likely mechanism for exhumation of coesite-bearing eclogite.

Cooling During Exhumation

The great depth and high P/T ratio of UHP metamorphism indicates that the deep burial of coesite-bearing rocks must have occurred by subduction. Away

from plate boundaries, inferred oceanic and continental geotherms pass through temperatures in the range 1000–1120°C at depths of ≈ 100 km (Fig. 5.3; Pollack and Chapman, 1977; surface heat flux of $50 \text{ mW}\cdot\text{m}^{-2}$). The coesite-bearing eclogites record markedly cooler temperatures (550–900°C) at similar depths. As discussed above, these relatively cool metamorphic conditions are consistent with transient and steady-state P - T paths calculated for subducting lithosphere.

A greater problem is posed by the preservation of low temperatures during exhumation of coesite-bearing eclogites, as illustrated in Fig 5.3B. Most exhumation scenarios result in heating during decompression, in contrast with the petrologic evidence discussed above, which requires that the coesite-bearing eclogites cooled during decompression.

P - T path A (Fig 5.3B) represents the trajectory followed by coesite-bearing eclogites if subduction ceases and no exhumation occurs. In this scenario, the eclogites heat isobarically as the thermal structure of the former subduction zone relaxes toward a new, warmer steady state. Once subduction ceases, the isotherms rise toward the surface because the isotherms are no longer deflected downward by the subducting lithosphere. If the coesite-bearing eclogites followed P - T path A, new, higher temperature metamorphic assemblages would be expected to form.

If subduction ceases and the coesite-bearing eclogites are exhumed by erosion at a moderate rate of ≈ 0.5 – 1.0 km Ma^{-1} , they will follow a P - T trajectory like path B (Fig 5.3B). This “clockwise” P - T path is similar to the calculated paths followed by metamorphic rocks exhumed by erosion subsequent to crustal thickening (e.g., England and Thompson, 1984). During the early stages of decompression, the rocks heat up as the thermal structure reequilibrates and approaches the post-subduction steady-state geotherm. As exhumation continues, the rocks cool as they approach the surface. Because the rocks become hotter during decompression, new mineral assemblages may form to reflect the higher temperatures achieved at lower pressures. For P - T path B, a new mineral assemblage in the eclogites is likely to form near the maximum temperatures reached during decompression: temperatures of $\approx 1050^\circ\text{C}$ at pressures of $\approx 2 \text{ GPa}$.

P - T path C (Fig 5.3B) represents the hypothetical trajectory followed by coesite-bearing eclogites if subduction ceases and the rocks are exhumed instantaneously (i.e., an infinitely rapid erosion rate). In this end-member situation, the eclogites will undergo adiabatic (nearly isothermal) decompression. During decompression, the rocks would cross many metamorphic reactions that have a positive P - T slope, such as the reaction phengite + talc \rightarrow phlogopite + kyanite + quartz + H_2O (Fig 5.1A). There is no petrologic evidence

for such reaction having occurred during decompression of Dora-Maira rocks (e.g., Chopin, 1984).

The retrograde P - T paths determined for coesite-bearing eclogites suggest cooling during decompression (Fig 5.1) in contrast to the heating (paths A and B) and isothermal (path C) decompression paths depicted in Fig 5.3B. Maximum temperatures appear to have been reached during peak coesite-eclogite facies metamorphism and not during decompression. Cooling during decompression requires that the coesite-bearing eclogites lost heat to the surrounding rocks throughout the exhumation process. Heat may have been conducted downward into cooler underlying rocks if subduction continued beneath the coesite-bearing eclogite terrains during exhumation of the eclogites. Alternatively, if the coesite-bearing eclogites were exhumed by tectonic extension of overlying rocks, then eclogites lying beneath the normal fault (i.e., in the upper part of the lower plate) might cool as heat is conducted into the cooler upper plate of the detachment system. Cooling during decompression would be most pronounced if subduction continued to occur during exhumation and if eclogites were transported toward the surface in the lower plate of extensional structures.

P - T path D (Fig 5.3B) represents the trajectory followed by coesite-bearing eclogites if exhumation occurred while subduction continued beneath the eclogites, rather than after subduction ceased. Continued subduction or underplating of additional units would keep isotherms depressed in the region of the subducting slab and rocks exhumed during this period would cool during decompression (e.g., Rubie, 1984; Davy and Gillet, 1986; Gillet et al., 1986; Ernst, 1988). If coesite-bearing eclogites formed and were exhumed in a steady-state subduction zone thermal structure, then the retrograde, decompression P - T path could closely track the prograde, subduction P - T path. In thermal steady state, temperatures within the lithosphere of the upper plate increase monotonically from the surface down to the subduction shear zone; no inverted thermal gradients exist in the upper plate. If exhumation took place under approximately steady-state thermal conditions, we cannot constrain the location of the eclogites with respect to the subduction shear zone during exhumation. Alternatively, if the eclogite protoliths were subducted prior to the development of thermal steady state, then the retrograde P - T path could be cooler than the prograde P - T path.

In the second heat loss scenario, coesite-bearing eclogite terrains in the lower plate of an extensional structure, such as a normal fault or shear zone, cool during exhumation as a result of conduction of heat into the upper plate. Prior to exhumation, rocks lying above the eclogites will be generally cooler than a steady-state geotherm because of the depression of isotherms by the

subduction process. If the eclogites are exhumed by movement along a low-angle normal fault that cuts through these cooler rocks, then the upper portions of the footwall will lose heat to the cooler hangingwall during movement along the fault. Rocks located at the top of the footwall will cool more than rocks located deeper in the footwall. If exhumation occurs after subduction ceases, then there will be competition between cooling due to contact with the cooler hangingwall and heating from below as isotherms move toward the surface during postsubduction thermal relaxation.

In the case of the Western Alps, the hypothesis of exhumation during continued subduction is supported by plate reconstructions which suggest that the Mesozoic Tethyan ocean basin was slowly subducted beneath the Apulian (African) plate between 120 and 40 Ma (e.g., Dewey et al., 1989; Hsü, 1989). Eoalpine HP metamorphism in the Dora-Maira region occurred during the early stages of plate convergence, followed by thrusting of the Pennine nappes beneath the Dora-Maira. Consumption of the Tethyan ocean basin led to collision of the European and Apulian plates at ≈ 40 Ma, emplacement of the Helvetic nappes, and widespread regional metamorphism. At ≈ 40 Ma, according to Monié and Chopin (1991), the Dora-Maira massif was undergoing greenschist-facies retrogression at depths of 15–20 km. Subduction of the Tethyan ocean basin was apparently occurring beneath the Dora-Maira massif from 120 to 40 Ma while the eclogites were being exhumed from depths of 100–120 km to 15–20 km.

While this scenario appears to be accepted by most workers, the recent radiometric data of Tilton et al. (1991), Gebauer et al. (1991), and Becker (1993) raise the possibility that UHP Dora-Maira metamorphism may be as young as 40 Ma, and would thus correspond to the onset of continental collision. In this case, the exhumation of UHP rocks must have occurred without the refrigerative effects of continuing, deeper-level subduction, and cooling the eclogites during decompression is even more problematic.

Exhumation Rates

In certain situations, the shape of the decompression portion of a metamorphic P - T path can be used to constrain the exhumation rate of the metamorphic rock. For example, Draper and Bone (1981) showed that if subduction ceases, blueschist-facies rocks in a subduction zone will heat up and be overprinted by greenschist- or amphibolite-facies assemblages unless very rapid exhumation rates of 10–100 km Ma^{-1} prevail. For the case of the coesite-bearing eclogites, as well as the many noncollisional blueschist terrains that cooled during exhumation (e.g., Ernst, 1988), the thermal structure during

exhumation was probably controlled by deeper-level subduction coincident with exhumation. Therefore, it is not possible to constrain the exhumation rate solely on the basis of the retrograde P - T path (Rubie, 1984, 1990). Exhumation may have been relatively rapid or relatively slow as long as subduction continued to chill the overlying rocks.

Average rates of exhumation may be calculated by dating different portions of a metamorphic P - T path. For example, if the Dora-Maira coesite-bearing eclogites underwent UHP metamorphism at ≈ 38 Ma (Tilton et al., 1991) and metamorphic pressures of ≈ 3.2 GPa (Schertl et al., 1991), the average long-term exhumation rate was a moderate 3.0 km Ma^{-1} ($3.0 \text{ mm}\cdot\text{yr}^{-1}$). Exhumation rates for Dabie Mountains coesite-bearing eclogites constrained by data discussed earlier are $\approx 0.75 \text{ km m}\cdot\text{y}^{-1}$. Fault-zone parallel displacement rates of 10 – $20 \text{ mm}\cdot\text{yr}^{-1}$ have been inferred for Basin and Range extensional faults by Davis (1988), Hacker et al. (1991), and Spencer and Reynolds (1991). With a displacement rate of 10 – $20 \text{ mm}\cdot\text{yr}^{-1}$, an extensional fault dipping 30° could bring rocks from a depth of 100 km to the surface in 10 – 20 Ma .

Conclusions

Coesite-bearing eclogites record relatively low-temperature, HP metamorphic conditions consistent with formation in a subduction zone environment. Existing kinetic data suggest that in the presence of fluid the retrograde coesite \rightarrow quartz and aragonite \rightarrow calcite reactions should progress to completion during exhumation of UHP rocks. The preservation of coesite and aragonite only as inclusions in other minerals suggests that retrogressive phase changes can be inhibited within a porphyroblast capable of maintaining high internal pressures and excluding catalytic volatiles. More experimental kinetic data on the coesite \rightarrow quartz and aragonite \rightarrow calcite reactions are needed; the effects of fluids, of powdering, and of inclusions within other phases must be assessed quantitatively before experimental data can be extrapolated to the Earth.

Careful examination of natural samples that reached different peak temperatures at UHPs may reveal additional information about the rate and mechanisms of the coesite \rightarrow quartz and aragonite \rightarrow calcite transformations. UHP occurrences in central China are a good candidate for such study, because peak temperatures vary from $\approx 550^\circ\text{C}$ to 850°C along the length of the collisional orogen. Cathodoluminescence and microscale isotopic and chemical determinations might also be used to identify different stages in the formation of retrogressive quartz or calcite.

Polycrystalline quartz aggregates and cracks radiating outward from inclusions are not, by themselves, reliable indicators of UHP metamorphism. In

the absence of relict coesite, subparallel, sometimes curving quartz subgrains and the truncation of one set of quartz subgrains by another are probable indicators of the earlier presence of coesite.

Petrologic evidence for cooling during decompression from depths in excess of 90 km requires that the exhumation of coesite-bearing terrains occurred while subduction/underflow continued to chill the overlying plate or that the eclogites were located near the top of an extensional structure's footwall. On the basis of radiometric data, the average rate of exhumation of the Dora-Maira coesite-bearing eclogite was $\approx 3.0 \text{ km Ma}^{-1}$, assuming that high-pressure metamorphism occurred $\approx 38 \text{ Ma}$. Exhumation from depths $>90 \text{ km}$ demands extensional tectonic thinning in addition to sedimentary erosion. Extensional features above the Dora-Maira eclogites have been described by Ballèvre et al. (1990); we predict that similar structures will be found above other UHPM terrains. More field work is required to reveal whether such extensional features exist in other terrains and to determine the age, magnitude, and sense of displacement along these structures.

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