Review

Diamond and Other Exotic Mineral-Bearing Ophiolites on the Globe: A Key to Understand the Discovery of New Minerals and Formation of Ophiolitic Podiform Chromitite

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Abstract: Ophiolite-hosted diamond from peridotites and podiform chromitites significantly differs from those of kimberlitic diamond and ultra-high pressure (UHP) metamorphic diamond in terms of occurrence, mineral inclusion, as well as carbon and nitrogen isotopic composition. In this review, we briefly summarize the global distribution of twenty-five diamond-bearing ophiolites in different suture zones and outline the bulk-rock compositions, mineral and particular Re-Os isotopic systematics of these ophiolitic chromitites and host peridotites. These data indicate that the subcontinental lithospheric mantle is likely involved in the formation of podiform chromitite. We also provide an overview of the UHP textures and unusual mineral assemblages, including diamonds, other UHP minerals (e.g., moissanite, coesite) and crustal minerals, which robustly offer evidence of crustal recycling in the deep mantle along the suprasubduction zone (SSZ) and then being transported to shallow mantle depths by asthenospheric mantle upwelling in mid-ocean-ridge and SSZ settings. A systematic comparison between four main genetic models provides insights into our understanding of the origin of ophiolite-hosted diamond and the formation of podiform chromitite. Diamond-bearing peridotites and chromitites in ophiolites are important objects to discover new minerals from the deep earth and provide clues on the chemical composition and the physical condition of the deep mantle.

Keywords: diamond; ophiolite; chromitite; peridotite; crustal recycling into deep mantle

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1. Introduction

According to the differences in occurrence and genesis, natural diamonds are divided into various types, including diamonds occurring in kimberlite, lamproite, ophiolite, alkaline mafic rock, crust-derived ultrahigh-pressure (UHP) metamorphic rock, meteorolite-related and alluvial rocks [1–5]. Ophiolites are fragments of ancient oceanic crust and upper mantle, which is created at ocean spreading ridges and then emplaced on land [6]. Ophiolite-hosted diamond discovered in ophiolitic peridotite and chromitite is considered to be a new type that has been named an ophiolite-type in recent years [4,7].

It has been hundreds of years since the first discovery of diamonds in ophiolites [8–10]. Diamonds were successively discovered in ophiolites from Quebec in Canada [8], Kamenusha in the Urals (Russia) and Koryak in the Far East [9,10]. However, these early stage studies merely reported the occurrence of diamonds in ophiolites but did not systematically relate these discoveries to plate tectonics. Since the 1980s, the diamond group in the Chinese Academy of Geological Sciences discovered diamonds in the Luobusa ophiolite (also named Qusong) in the Indus-Yarlung Zangbo suture zone (IYSZ) and Dongqiao ophiolite

in the Bangong Nujiang suture zone (BNSZ) in Tibet, China [11–14]. Subsequently, Jingsui Yang and his group continued this research, and diamonds were also discovered in the ophiolites of Luobusa, Purang, Dongbo, Dangqiong and Dingqing in Tibet, Sartohay in Xinjiang, Hegenshan in Inner Mongolia, Myitkyina and Kalembo in Myanmar, Pozanti-Karsanti in Turkey, Mirdita in Albania and Horoman in Japan [15–33]. As these diamonds were extracted by heavy mineral separation in the early stage, their natural origin was initially doubted [7,34]. However, it was not until the *in situ* diamonds were discovered in the Luobusa and Polar Ural chromitites as well as the Nidar mantle peridotites [24,35]. Particularly, ophiolite-hosted diamonds have inclusions of fluid and Mn-Co-Ni alloy and obviously show trace elements, e.g., Ba, Pb, Th and Eu, and crust-derived carbon–nitrogen isotope compositions different from those of synthetic diamonds [5,36,37]. Following the independent discovery of diamonds in ophiolitic mantle peridotites and chromitites by Australian, Spanish and Indian geologists, the occurrence of the ophiolite-type diamond was widely accepted [35,36,38,39].

Ophiolite-hosted diamond and other UHP minerals have been globally discovered in ophiolites from various orogenic belts, suggesting that diamonds are widespread in mantle rocks [5,24,37,40]. The discovery of ophiolite-hosted diamond opens a new field for research on the genesis of ophiolitic chromitite and host mantle peridotite and for the exploration of the dynamics of crust–mantle recycling [41–43]. However, the amounts of diamonds extracted from various ophiolitic peridotites and chromitites differ greatly. For most massifs, several diamond grains were recovered per ton, while over one thousand diamond grains per ton were discovered from the Luobusa mantle peridotites and chromitites. In contrast, less than ten diamond grains per ton were recovered from mantle peridotites in the Xialu (Xigaze), Zedang, Dongbo and Myitkyina massifs [40,44]. It is still unclear what contributes to the difference in such diamond concentrations. A discussion of the spatial distribution and formation mechanism of ophiolite-hosted diamond has become a new direction of current research of ophiolites and plate tectonics [5,37,40]. This paper summarizes the geological characteristics of diamond-bearing ophiolites and discusses the four genetic models for the diamond-bearing and associated unusual minerals-bearing ophiolites, providing a basis for discussing the formation of ophiolitic chromitite and the dynamic process of crust–mantle material recycling.

2. Global Distribution of Diamond-Bearing Ophiolites

Twenty-five diamond-bearing ophiolites have been discovered across the globe, and they are mainly distributed along the Tethyan Orogenic Belt, the Ural–Central Asian Orogenic Belt, the Hidaka metamorphic belt, the Klamath–Acatlán Orogenic Belt in North America and the Andean Orogenic Belt in South America (Figure 1a). The Tethyan Orogenic Belt is a giant and long-lived plate subduction–collision system on the earth that separates the Gondwana supercontinent to the south from the Laurasia supercontinent to the north [45]. It displays a structural difference of continent–continent collision alternating with ocean–continent subduction from east to west (Figure 1b). This belt extends from the Alpine continent–continent collisional orogenic belt in Europe via the complicated Mediterranean ocean–continent subduction belt, eastwards to the Zagros continent–continent collisional orogenic belt in West Asia and the Makran ocean–continent subduction belt to the east, and to the Himalayan continent–continent collisional orogenic belt. Then it connects with the ocean–continent subduction zone formed by East Indian oceanic slab subducting below Southeast Asia [46]. The Tethyan Orogenic Belt represents an oceanic system remnant commonly composed of Prototethyan, Paleotethyan and Neotethyan oceans that developed during the Early Paleozoic, Late Paleozoic and Mesozoic, respectively [45]. The Paleotethyan ocean represented by late Paleozoic–Triassic ophiolites was bounded by the Cimmerides to the south that mainly consists of Anatolia, Persia, Afghanistan and Tibet, and Cathaysides to the north that includes South China, Indochina, Sibumasu and Qamdo blocks [45,47,48]. The Neotethyan ocean documents the Mesozoic rifting of Pangea and comprises multiple seaways formed by seafloor spreading and widening eastwards and

deep mantle circulation processes, including several stages of slab subduction and mantle plume activities [49].

Figure 1. Distribution of twenty-five diamond-bearing ophiolites reported on the globe (a). Nineteen diamond-bearing ophiolites are distributed along the Neotethyan Orogenic Belt, Kunlun–Qilian–Qinling Paleotethyan Orogenic Belt (KQQ), Central Asian Orogenic Belt and Hidaka metamorphic belt (b). Ophiolites: C—Cuobuzha, D—Dongqiao, DB—Dongbo, DQ—Dingqing, DX—Dangqiong–Xiugugabu, EM—Eastern Mirdita (Albania), K—Kalamyo (Myanmar), L—Luobusa, M—Myitkyina (Myanmar), N—Nidar (Kashmir), P—Purang, PK—Pozanti Karsanti (Turkey), QL—Qilian, WM—Western Mirdita (Albania), X—Xialu, Z—Zedang. Ural–Central Asian Orogenic Belt: Sartohay, Hegenshan, Ray-Iz (Polar Urals). Hidaka Metamorphic Belt: H—Horoman (Japan). BNSZ—Bangong Nuijiang suture zone, IYSZ—Indus Yarlung Zangbo suture zone. Internal structures are modified from [50–53].

The Neotethyan Orogenic Belt, which was discovered with fifteen diamond-bearing ophiolites, includes three subbelts of the BNSZ, IYSZ and the Eastern Mediterranean zone (Figure 1b). Diamond-bearing Dongqiao and Dingqing ophiolites occur in the BNSZ,

Tibet [13,14,27]; the Myitkyina ophiolite is located in the Burmese Eastern ophiolitic belt that is regarded to be the southern continuation of the BNSZ [54]. The Luobusa, Zedang, Xialu, Dangqiong–Xiugugabu, Purang, Dongbo, Cuobuzha and Nidar ophiolites occur in the IYSZ [7,14,20,23,25,26,35,55–57]; the Kalamyo ophiolite crops out in the Indo-Burma Range that is regarded to be connected with the IYSZ [58]. The Pozanti Karsanti (Turkey), East Mirdita (Bulqiza, Albania) and West Mirdita (Skenderbeu, Albania) ophiolites occur in the Mediterranean ophiolite belt [28,30,32] (Figure 1b).

The 495–550 Ma diamond-bearing Yushigou ophiolite in the North Qilian Orogenic Belt represents the relics of the Prototethyan ocean [59]. Diamond-bearing ophiolites in the Ural–Central Asian Orogenic Belt consist of the Ray–Iz massif in the Polar Urals [24], the Sartohay ophiolite in Xinjiang and the Hegenshan ophiolite in Inner Mongolia [21,22] (Figure 1b). The diamond-bearing Horoman mantle peridotites [33], each ~10 km long, ~8 km wide and ~3 km thick, are exposed along the southern edge of the NNW–SSE oriented, ~120 km long and 10–20 km wide Hidaka metamorphic belt, which is the boundary between the North American Plate (Okhotsk block) and the Eurasian Plate [60]. The Klamath–Acatlán Orogenic Belt includes the Klamath massif in western United States and the Acatlán complex in southern Mexico [39]. Ophiolite-hosted diamonds in this belt were discovered in the Jurassic Josephine peridotite [61] and the Early Paleozoic Tehuiztingo chromitite [39], respectively. About ten diamond grains were discovered in the Moa-Baracoa ophiolitic peridotite massif in eastern Cuba, which is about 100 km long and 10–30 km wide, covering an area of ~1500 km² (Rui et al., in preparation). One diamond grain was obtained from the Taitao peridotite in Chile, the western part of the Andean Orogenic Belt (Wu et al., in preparation).

3. Geochemical Characteristics of Diamond-Bearing Ophiolites

The wall rocks of the ophiolite-hosted diamonds are podiform chromitite and mantle peridotite. Mantle peridotite is dominated by Cpx-bearing harzburgite (Figure 2). Diamond-bearing lherzolite was merely reported in the Purang and Horoman massifs [33,56]. Ophiolites in the IYSZ, including the Purang, Dongbo, Cuobuzha, Dangqiong, Xialu, Zedang and Luobusa massifs, commonly have harzburgites with low contents of Al and Ca but high contents of Mg relative to the primitive mantle. The REE and Os contents of these harzburgites are notably lower than those of the primitive mantle. These geochemical features are comparable with those of depleted abyssal peridotites, which have experienced variable degrees of partial melting [62,63]. Some samples are relatively enriched in LREE, Rb and Ba and have a comparatively large range of Cr[#] ($\text{Cr}^\# = 100 \times \text{Cr}/(\text{Cr} + \text{Al})$, 18–75) in chromite (Cr-spinel). The diamond-bearing peridotites commonly show higher contents of platinum group elements (PGE) than those of the primitive mantle. Such geochemical characteristics suggest that these depleted abyssal peridotites were widely metasomatized by sulfide-rich, incompatible element-rich and high-PGE fluids/melts [62–65].

In a rock–melt interaction process, Re–Os isotopes can reflect the impacts of infiltrating melts on the ophiolitic mantle [66,67]. Os is a highly compatible element during partial melting, while Re is a moderately incompatible element, which leads to comparatively low Re/Os ratios and increases the ¹⁸⁷Os/¹⁸⁸Os ratio in residual mantle peridotites, although the rate of growth reduces over time [66–68]. Thus, the lower ¹⁸⁷Os/¹⁸⁸Os ratios recorded in peridotites may indicate an earlier local melting event [63,68]. In addition, serpentinization, seafloor alteration and metasomatism of mantle peridotites by fluids/melts pose weak impacts on the whole rock's ¹⁸⁷Os/¹⁸⁸Os ratio, and thus, the Re depletion model ages can reflect the time of partial melting [67,68]. Compared to the average ¹⁸⁷Os/¹⁸⁸Os ratio of a primitive upper mantle (0.1296) [69], seventy-eight ¹⁸⁷Os/¹⁸⁸Os ratios (0.113–0.145, with an average of 0.125) of the Purang, Dongbo and Cuobuzha harzburgites in the western segment of IYSZ overlap the ranges of the subcontinental lithospheric mantle (SCLM, 0.105–0.129) [70] and depleted oceanic lithospheric mantle (0.123–0.129) [70,71]. The Baigang and Dazhuqu mantle peridotites in central IYSZ display a relatively large range in ¹⁸⁷Os/¹⁸⁸Os (0.118–0.130, with an average of 0.126). Additionally, sixty-eight

$^{187}\text{Os}/^{188}\text{Os}$ ratios of Luobusa and Zedang mantle peridotites in the eastern segment show a comparatively large range (0.121–0.137, with an average of 0.126 [63,72,73]. They all show hybrid signatures of depleted oceanic lithosphere and SCLM (Figure 3).

Figure 2. Compiled stratigraphic-lithological columnar sections of the main diamond-bearing ophiolites shown in Figure 1b along the Neotethyan Orogenic Belt. Red triangles labeled as diamonds hosted in podiform chromitite, harzburgite and lherzolite. (a) Dingqing ophiolite from the eastern part of the BNSZ, Tibet [27,44]; (b) Dongqiao ophiolite from the central part of the BNSZ, Tibet [13,14,44]; (c) Dongbo, (d) Purang and (e) Xiugugabu ophiolites from the western part of the IYSZ [44]; (f) Luobusa ophiolite from the eastern part of the IYSZ [14,16,17]; (g) Xialu ophiolite from the central part of the IYSZ [26,44]; (h) East Mirdita (Bulqiza massif) from the eastern Mediterranean ophiolite belt [28].

Figure 3. Osmium isotopic values of diamond-bearing chromitite and peridotite in different ophiolites. Osmium isotopic values of primitive upper mantle, depleted mid-ocean-ridge-basalt mantle, Ocean-island basalt (OIB), Pico and Faial OIB from the Azores after [71]. Osmium isotopic data references: Purang [63,74,75]; Dongbo [62]; Cuobuzha [65]; Baigang [63]; Dazhuqu [76]; Zedang [63,77]; Luobusa [72,78,79]; Bangong Lake [80]; Yunzhug [81]; Dongqiao [67,79,82]; Sartohay [79]; Yushigou [59].

Podiform chromitite is defined as lenticular aggregates of chromite formed in Alpine-type peridotites and oceanic crust–mantle transition zones [83], and it normally contains over 20 vol.% chromite, and preserves abundant magma rheological and high-temperature deformation structures that are distinct from those of layered chromitite [83,84]. Podiform chromitite, by occurrences and Cr values of chromite, may be classified as follows: (1) According to the exposed locations of chromitite orebodies in a well-exposed ophiolite sequence, ophiolite can be divided into subvarieties occurring below and within the MOHO transition zone [83,85,86]. The former is mostly enclosed in harzburgite with a thin dunite-envelope and usually occurs as large-sized deposits of economic value. The latter is generally distributed in thin-bedded dunite and usually forms small-sized, disseminated and banded chromitite deposits, which are commonly interlayered with cumulate dunite [86]. (2) According to the structural deformation features of podiform chromitite orebodies, chromitite deposits are mainly divided into discordant, subconcordant and concordant types compared with permeable structures of foliation and lineation developed in host mantle peridotites [87]. Discordant ore bodies are irregular and obviously cut through the foliation and lineation of peridotites, subconcordant orebodies usually have angles of 10°–25° with the foliation, and concordant ore bodies are commonly parallel to the foliation and lineation [87]. (3) According to the chromite aggregation forms, podiform chromitites are divided into massive, nodular, antinodular, densely/sparsely disseminated, banded and vein-like types [85,88,89]. (4) According to geochemical compositions of chromite in podiform chromitite, it can be divided into high-Cr ($\text{Cr}^\#$ value > 60) type and high-Al ($\text{Cr}^\#$ value < 60) type [90]. High-Al chromitite is usually considered to be precipitated from basaltic melts, which might be formed at a mid-ocean ridge (MOR) or derived from backarc or forearc settings at a suprasubduction zone (SSZ). High-Cr type is generally related to boninitic melt metasomatism in subduction zones [91,92]. Recently, Li et al. (2019) further subdivided chromitite from the Dingqing ophiolite in the BNSZ, China, into high-Cr ($\text{Cr}^\# = 78\text{--}86$), medium-high-Cr ($\text{Cr}^\# = 60\text{--}74$), medium-Cr ($\text{Cr}^\# = 30\text{--}51$) and low-Cr ($\text{Cr}^\# = 9\text{--}14$) types [89].

Ophiolite-hosted diamonds occur in both high-Cr chromitite, e.g., the Ray-Iz and Luobusa ophiolites [24,93], as well as in high-Al chromitite, e.g., the Sartohay and Hegen-

shan ophiolites [21,22]. They even have been recovered from high-Al and high-Cr type chromitites in the same ophiolite massif such as Dongbo, Purang, Dingqing, Mirdita and Pozanti Karsanti [29,32,89,94]. Current statistics show that the amount of diamond grains in high-Cr type chromitite is much higher than those in high-Al type chromitite [40]. For example, the number of diamond grains from the Luobusa, Kangjinla and Ray-Iz massive high-Cr chromitites reaches 1000 per ton in total [20,24,57], while in both high-Al type and high-Al-high-Cr type chromitites from, e.g., the Dongbo and Purang ophiolites, there are commonly only several to dozens of diamond grains per ton [7,56,95]. The reason for these differences is unclear.

$^{187}\text{Os}/^{188}\text{Os}$ ratios of chromitites are significantly higher in high-Cr chromitites compared to high-Al chromitites. For example, for Dongbo and Purang, the high-Cr chromitites have $^{187}\text{Os}/^{188}\text{Os}$ compositions varying from 0.128 to 0.133 and 0.123 to 0.132, respectively, whereas their high-Al chromitites have values ranging from 0.120 to 0.126 and 0.124 to 0.127 [62,63,74,94]. It is noted that compared to the Purang and Dongbo high-Cr chromitites, the Luobusa high-Cr chromitites exhibit generally lower $^{187}\text{Os}/^{188}\text{Os}$ ratios of 0.104–0.127 [72,78,79] (Figure 3). The latter chromitite massif has been interpreted to have been formed by decompression partial melting of garnet and orthopyroxene (Opx) in the SCLM [37,79]. Alternatively, the Luobusa podiform chromitite was produced by mixing primitive asthenospheric Cr-rich melt and boninitic magma, triggering the saturation and crystallization of chromite [96].

4. Ultra-High-Pressure Minerals in Ophiolites

The main minerals separated from ophiolitic peridotites and chromitites include native elements (e.g., diamond, Si), carbides, metal alloys, oxides, sulfides, silicates and fluorides [5,40,97]. Ophiolite-hosted diamonds are usually light yellow and yellowish-green (Figure 4a–c), with rare grayish-green or black colors; some may even be colorless. They mainly occur as monocrystals with grain sizes generally ranging from 0.1 mm to 0.5 mm (Figure 4a–d). The largest diamond grain discovered so far comes from Dongqiao and has grain size up to $0.7\text{ mm} \times 0.6\text{ mm} \times 0.6\text{ mm}$, weighing $\sim 0.43\text{ mg}$ in total [13]. Diamonds are mostly euhedral and subhedral monocrystals, with minor schistose or polycrystals (Figure 4d) for those ophiolitic in situ diamonds (Figure 4e,f), and grains extracted from the chromitites and peridotites (Figure 4a–d). Morphologically they are characterized by octahedrons, rhombic dodecahedrons and octahedron-cube aggregates. Diamonds with a mixed crystal habit exhibit growth sectors with various levels of darkness and brightness under cathodoluminescence (CL) light (Figure 4g,h) [98]. The bright part commonly represents an octahedral growth sector with a layered growth characteristic, and a dull sector represents a cuboid growth sector [5,98]. Locally, irregular corrosion pits or branching etched grooves are observable [4]. Fourier Transform Infrared (FTIR) spectra studies show that the Luobusa diamonds are IaA, mixed IaA–IaB and Ib types [4,36,99–101].

Figure 4. Cont.

Figure 4. Photomicrographs of ophiolite-hosted diamonds from chromitite and peridotite. (a) Diamonds from the Ray-Iz chromitite, Russia [17]; (b) diamonds from the Pozanti-Karsanti (Al-adag) chromitite, Turkey [32]; (c) diamonds from the Ray-Iz chromitite, Russia [24]; (d) backscattered secondary electron image showing polycrystal striations on the diamond surface [20]; (e,f) chromite hosted in an in situ diamond [24]; (f) in situ diamonds hosted by chromitite from the Luobusa and Ray-Iz, respectively [24]; (g) cathodoluminescence (CL) image of a Kalamyo harzburgite-hosted diamond [58]; (h) CL image of a Luobusa harzburgite-hosted diamond.

Diamond is a typical UHP mineral that crystallizes at depths above 150 km (~5 GPa) [3,102]. Ophiolite-hosted diamonds usually have irregular spots or a thin metal film on the surface, e.g., the Ray-Iz diamonds display a 0.01 µm film on the surface, whose compositions are dominated by Zr and ZrO₂, with minor Cu sulfides, NaCl, SiO₂, Ta, Fe and Cr [103]. This metal film is considered to be the product of diamond growth under a high-temperature, high-pressure mantle environment [104]. The microdiamonds are usually associated with other UHP and/or highly reduced minerals, such as elemental Fe, Cr, Ni and W and alloys of SiC, OsIr and FeTi, indicating that these minerals were formed in a highly reduced environment [37]. An Os-Ir alloy separated from the Luobusa chromitites contains in situ diamond inclusions of 1–2 µm and an FeTi alloy rimmed by kyanite and coesite that shows pseudomorphs of stishovite, which indicates a formation pressure of >10 GPa (300 km) [17]. Subsequently, a new mineral of qingsongite (c-BN) was discovered in the coesite, which is suggested to form at a comparable pressure to stishovite

formation [105]. Yamamoto et al. (2009) discovered clinopyroxene (Cpx) and coesite exsolution assemblages in the Luobusa chromitite, which is interpreted to be derived from Ca-ferrite, suggesting a depth of formation of the chromitite in excess of 410 km [106]. Chromite in the Luobusa chromitites has a high Fe^{3+} content and abundant primary iron spheres. This combination may represent a high-pressure unbalanced reaction ($\text{Fe}^{2+} \rightarrow \text{Fe}^{3+} + \text{Fe}^0$) or a Ca-ferrite structured polymorph under deep mantle conditions [107]. The octahedral silicate mineral separated from Luobusa chromitites displays all Si atoms at octahedral coordination and has a cubic structure thR is similar to that of ringwoodite. This cubic structure is different from the cubic spinel structure of tetrahedral coordination but instead consistent with an inverse-spinel structure. Their equilibrating conditions are generally consistent with the formation of ringwoodite or wadsleyite, which are stable at $\sim 2000^\circ\text{C}$ and $\sim 15\text{ GPa}$ [38].

Evidence for UHP conditions is also observed in mantle peridotites. A symplectite texture composed of spinel + Opx $\pm$ Cpx was discovered in Luobusa and Purang harzburgites [38,75,108,109]. It is interpreted as a product of decompressional decomposition of high-pressure khoharite, indicating a formation depth of above 13.5 Gpa ($\sim 400\text{ km}$). Moreover, several lines of evidence, including the presence of C2/c clinoenstatite and high-pressure $\beta\text{-Mg}_2\text{SiO}_4$, reported in situ diamond and coesite in the Nidar peridotites in the NW Himalaya [35,110], abundant clinoenstatite as lamellae associated with diopside exsolution in orthoenstatite from the Luobusa mantle peridotites [111], assemblages of carborundum, wustite and native iron inclusions, as well as FeNi and FeCr alloys observed in olivines and chromites in the Luobusa peridotites, all indicate a UHP environment where ophiolitic chromitite and host peridotite formed [112–114].

The other two significant UHP minerals associated with diamond are carborundum (SiC) and corundum. SiC, also called moissanite, is usually light green, pale blue, dark blue and colorless (Figure 5a), with grain sizes of 50–200 μm [5]. Corundum is a common mineral in eclogite and occurs as an unusual inclusion within diamonds [115]. Recent research on corundum in the Luobusa chromitite mine, which is subdivided into, from west to east, Luobusa, Xiangkashan and Kangjinla ore zones in Tibet, has made fruitful and significant progress [20,57,115,116]. The Kangjinla corundums are translucent to transparent, mostly pinkish red, and in rare cases colorless, light yellow and light brown. They are characterized by grain sizes of $>200\text{ }\mu\text{m}$, Al_2O_3 contents of 96–99.57% and minor Ti, Fe and Al oxides [115]. Corundums contain a variety of mineral inclusions, including rutile, native Ti, and Ti alloys of Ti-N, Ti-Si, Ti-C, Ti-Si-P, Ti-B and Ti-Zr-Al, and REE-bearing (e.g., La, Ce, Nd) silicate minerals and other unknown minerals [57,100] (Figure 5b). Five nano-micron-sized titanium-bearing inclusions called jingsuite (TiB_2), zhiqinite (TiSi_2), badengzhuite (TiP), kangjinlaite ($\text{Ti}_{11}\text{Si}_{10}$) and wenjiite ($\text{Ti}_{10}\text{Si}_x\text{P}_y$, $x > y$, $6 \leq (x + y) \leq 7$) have recently been approved by the IMA [116,117]. These discoveries are significant and are another great breakthrough made on the basis of the discovery of eight new minerals of qingsongite (c-BN), qusongite (WC), luobusaite ($\text{Fe}_{0.83}\text{Si}_2$), yalungite ($(\text{Cr}_4\text{Fe}_4\text{Ni})_9\text{C}_4$), zangboite (TiFeSi_2), lingzhite (FeSi_2), naquite (FeSi) and native titanium [105,118]. These minerals, stable at different depths of the earth's interior, were formed under highly reducing and UHP conditions. Moreover, petrological experiments indicate that the melting temperature of corundum is higher than that of Fe, Ni and Co, and the temperature increases with the increasing pressure, with a maximum up to 25.4 Gpa and 3500 K. Corundum tends to be stable under the p–T conditions of the mantle transition zone (410–670 km) [119,120]. In summary, ophiolite-type diamond, corundum and their massive hosting chromitite and mantle peridotite most likely originated at the mantle transition zone or even in the lower mantle.

Figure 5. Mantle- and crust-derived minerals associated with diamond in ophiolitic chromitite and peridotite. (a) Moissanites from the Dingqing chromitite [27]; (b) Corundum and its inclusions from the Kangjinla chromitite [115]; (c) Rutiles from the Dangqiong harzburgite [23]; (d) zircons from the Dingqing chromitite [27].

5. Crustal Minerals in Ophiolites

In addition to the unusual UHP minerals, crust-derived minerals have been extracted by heavy mineral separation in ophiolitic peridotites and chromitites. These minerals mainly include oxides (e.g., quartz and rutile, Figure 5c), sulfides (e.g., pyrite, chalcopyrite and arsenopyrite) and silicate minerals (zircon, kyanite, sphene and almandine, etc.). Zircons have been discovered in peridotites and chromitites by many geologists in various ophiolites (Figure 5d). For example, Robinson et al. (2015) implemented in situ U-Pb dating on zircons from the Luobusa and Dongqiao ophiolites in Tibet and the Semail ophiolite in Oman. By secondary ion mass spectroscopy (SIMS), they obtained ages with large variations, i.e., 549–1675 Ma for the Luobusa zircons, 645–2695 Ma for the Dongqiao zircons and 84.3–1411 Ma for the Semail zircons [121]. Yamamoto et al. (2013) reported an even more complex age range of 100–2700 Ma for the inherited zircons from the Luobusa chromitites [122]. Moreover, McGowan et al. (2015) obtained an average zircon U-Pb age of 376 ± 7 Ma, $\epsilon_{\text{Hf}}(t)$ values of 9.7 ± 4.6 and δ^1 values of 4.8–8.2‰ for zircons from the Luobusa chromitite [55]. In comparison, inherited zircons from the Horoman peridotites in Japan yielded a wide range of ages between 159 and 3131 Ma with four peaks at 2385, 1890, 1618 and 1212 Ma [33]. The aforementioned inherited zircons have roughly similar inclusions of the crustal minerals: quartz, potash feldspar, muscovite, apatite, ilmenite and rutile, and no mantle minerals such as olivine, pyroxene and spinel. Combined with zircon REE patterns and trace elements abundances and ratios, it can be concluded that these peridotite and chromitite-hosted inherited zircons are of continental origins and mostly derived from granitic rocks [121,122].

Mineral inclusions within diamonds can also be used to reveal details of the environment of their formation and, thus, gain insight into the genesis of the host ophiolitic chromitite and peridotite. Mn-bearing mineral inclusions are commonly observed in

ophiolite-hosted diamond, e.g., Ni-Mn-Co alloy, Mn-bearing garnet, Mn spinel, Mn olivine, MnO and elemental Mn, which have been attributed to mainly indicate an ancient shallow origin as a ferromanganese crustal component subducted to over 200 km in depth [37]. Other inclusions in diamond, including graphite, coesite thought to be pseudomorphic after stishovite, and fluids with REE, H₂O, Cl, Ca and Si, are comparable with those of crust-derived fluid inclusions enveloped in kimberlitic diamonds [21,97]. The various suites of surficial inclusions indicate that crustal materials have been recycled into the deep mantle.

6. Isotopic Compositions of Carbon and Nitrogen in Diamond

The isotopic values of carbon ($\delta^{13}\text{C}$) and nitrogen ($\delta^{15}\text{N}$) in ophiolite-hosted diamond also record the recycling of various mantle and surficial materials. Ophiolite-hosted diamonds have negative carbon isotope ($\delta^{13}\text{C}$) values of -29.2‰ to -17.2‰ [5,6,24,31,33,40,123] (Figure 6a). These light $\delta^{13}\text{C}$ values are roughly identical to those of recycled carbon in subducted sediments (-42‰ to $+4.1\text{‰}$, -25‰ on average) [102], of common organic matter-like $\delta^{13}\text{C}$ compositions (-30‰ to -20‰) [124] and of HPHT synthetic octahedral diamonds (-42‰ to -16‰) [125]. Their shares are also comparable to $\delta^{13}\text{C}$ values from mantle peridotite-type (P-type) diamonds (-26.4‰ to $+0.2\text{‰}$) and eclogite-type (E-type) diamonds (-41.3‰ to $+2.7\text{‰}$) in kimberlites. They roughly coincide with those of some super-deep continental diamonds from Monastery (-18‰ to -16‰) and Jagersfontein (-24‰ to -16‰) from the Kaapvaal Craton in South Africa [126] and overlap with those of carbonado (-32‰ to -5‰) as well as those of diamonds (-30‰ to -3‰) hosted in ultra-high-pressure subduction-related metamorphic rocks [3,123,127]. In contrast, these light $\delta^{13}\text{C}$ values distinctly differ from those of diamonds in meteorites (-38.8‰ to -32.5‰) and coated/fibrous diamonds (-8.1‰ to -4.9‰) that are a special type of monocrystalline-polycrystalline diamond, where monocrystals have been overgrown by a thick, cloudy, polycrystalline coat or rod-blade-shaped diamond laden with microinclusions of fluid [3,128]. The aforementioned $\delta^{13}\text{C}$ range of ophiolite-hosted diamonds is much lower than that observed in diamonds from either kimberlites (mainly between -8‰ and -2‰) or UHP metamorphic rocks (mainly between -18‰ and -5‰), indicating a natural growth environment at special oceanic mantle conditions.

Figure 6. $\delta^{13}\text{C}$ values (a) and $\delta^{15}\text{N}$ values (b) for various types of diamonds and recycled carbon. Data mainly from [3,5,102,123,128,129].

The ophiolite-hosted diamonds have a large range of $\delta^{15}\text{N}$ values from -19.1‰ to $+28.6\text{‰}$ (Figure 6b), which is roughly comparable with that of P-type (-24‰ to $+12\text{‰}$) and E-type diamonds (-12‰ to $+18\text{‰}$) [5,129], and nearly overlap with those of synthetic HPHT diamonds, which show a large range from -10‰ to $+30\text{‰}$ [130]. Conversely, the $\delta^{15}\text{N}$ values of ophiolite-type diamonds are inconsistent with those of the bulk primi-

tive mantle (-10% to -4%) [129], fibrous diamonds (-12% to -0.3%) [5,102,123] and metamorphic diamonds (up to 0%) [5] (Figure 6b). It is important to note that the octahedral sectors of a large synthetic HPHT diamond were shown to have $\delta^{15}\text{N}$ values close to the atmospheric value (0%), while the cube sectors have higher $\delta^{15}\text{N}$ values of $+30\%$ to $+42\%$ [130,131], indicating that $\delta^{15}\text{N}$ values from different types of diamonds cannot simply be used to discriminate between natural and synthetic diamonds unless the crystal morphologies of the analyzed diamonds are considered [36]. Nevertheless, ophiolite-hosted diamonds generally have high contents of LREEs and are typically enriched in Ba, Pb, Th and Eu with typical negative anomalies of Eu, Sm, Yb and Y, and low concentrations of Fe. Their REE and trace element patterns of the ophiolite-hosted diamonds are parallel to those of fibrous diamonds [36,38]. They contain an admixture of nitrogen in an unaggregated form, which is commonly interpreted to indicate that the diamonds did not reside for a long time in the deep mantle [36,132].

7. Genetic Models of Diamond-Bearing Ophiolite and Chromitite

The discoveries of UHP minerals (e.g., diamond) and crust-derived minerals (e.g., zircon) in ophiolitic peridotites and chromitites have been gradually recognized. However, genetic models of these unusual minerals and their host chromitites are still under great debate. Four main genetic models are as follows.

Model I: The SCLM carrying low-pressure chromitites formed at a shallow depth was subducted into the deep mantle and then rapidly exhumed to a shallow depth [38]. Podiform chromitites that formed under low-pressure conditions in the shallow part of the SCLM were gradually detached and entered the upper mantle and the mantle transition zone due to the dragging of the subducted oceanic lithosphere (Figure 7a). High-pressure chromitites were then formed, and subducted slabs were metasomatized by carbon-bearing fluids to produce UHP and highly reduced minerals. Subsequently, the subducted slab retreated and gave rise to a shallow extension of the lithospheric mantle and the upward migrating of UHP chromitite to shallow levels beneath MORs and/or forearc and backarc spreading ridges (Figure 7b).

Figure 7. Cont.

Figure 7. Different genetic models of diamond, peridotite and podiform chromitite in ophiolites. (a,b)—after [38]; (c)—after [91]; (d)—after [92]; (e)—modified after [37,133]. BAB—backarc basin; Cont.—continental; MOR—mid-ocean ridge; OLM—oceanic lithospheric mantle; OP—oceanic plateau; SCLM—subcontinental lithospheric mantle.

This model is mainly based on the research on the Luobusa diamond-bearing chromitite. Re-Os isotopes of the Luobusa mantle peridotites display SCLM signatures [72]. The Re-Os T_{RD} ages of the peridotites show a wide range from 0.28 to 3.37 Ga. Among the younger ages is the peak Re-Os model age (325 Ma) of Ru-Os-Ir sulfides, and slightly older than that is the peak Re-Os model age (234 Ma) of Os-Ir nuggets in the chromitites [55]. While the minimum age (~376 Ma) of inherited zircons in chromitite was interpreted as the formation age of chromitite in the shallow part (<30 km) of the SCLM [55]. The 325–375 Ma may represent the time of ancient SCLM rifting and subsequent the subducted SCLM wedge materials detaching into the transition zone taken place in the Paleo-Tethys Ocean [38]. The age of 234 Ma was thought to represent the early crystalline phase of Os-Ir alloys enclosed in chromitites [72] and, thus, was explained to be the time of the delaminated chromitites being metasomatized within the transition zone [38,55]. Two subduction episodes at 150–170 Ma and 120–135 Ma in the Neo-Tethys Ocean may have disturbed the previously subducted lithosphere materials within the mantle transition zone, then excavated the chromitites that have undergone UHP metamorphism to shallow levels via mantle convection (Figure 7a,b). This model illustrates the formation of podiform chromitite and helps to gain insight into the process of crust-derived minerals and mantle-derived UHP and highly reduced minerals, but further consideration of the following issues may be required: (1) Needle-shaped diopside exsolutions are commonly observed in chromites from massive, nodular and disseminated chromitites within the Luobusa and Kangjinla massifs, and they were considered to mean that these three types of chromitites share the same evolution [38]. This point of view is not consistent with the interpretation that massive chromitite-hosted UHP and ultra-reduced minerals formed in the mantle transition zone, while disseminated and nodular chromitites crystallized at shallow depth through the reaction between tholeiitic/boninitic melts and host peridotite [4,93,97]. (2) Elements, such as Si and Zr, incompatible in the parent magma that formed chromite unlikely to crystallize to form zircon. Therefore, it is far-fetched to take 376 Ma as the age of magmatic formation of chromitite that falls within the range of the detrital zircon ages of 100–2700 Ma [121,122]. These detrital zircons with a relatively large age range could have resided for a long time in the mantle transition zone, implying that the process of crustal materials transported to the mantle transition zone is complicated and multi-sourced.

These subducted materials likely originate from relics of other ancient subducted oceanic lithospheres, such as those of the Panthalassa, Paleo-Asian, Neotethyan, Paleotethyan and even Prototethyan oceans. (3) Re-Os isotopes of chromitites and their host peridotites show SCLM signatures (Figure 3), suggesting that SCLM materials likely have participated in the formation of chromitite. This otherwise feasible model needs to be supported by additional geological evidence before it can be accepted.

Model II: Low-pressure chromitites were initially formed in MOR or SSZ settings at shallow depths of the oceanic mantle lithosphere, then delivered through mantle convection or subduction downward into the mantle transition zone where carbon-bearing fluids that were generated from the dehydration of subducted slabs produced diamonds in a high-pressure and strongly reducing environment. PGE alloys in UHP chromitites may have formed by the desulfuration of PGE-enriched sulfides from low-pressure-type chromitites. Diamond-bearing UHP chromitites are transported to a shallow depth by mantle convection and coexist with newly formed low-pressure chromitites [86,91] (Figure 7c). This model is generally similar to the first one, with the exception that in model II, the low-pressure chromitites formed in a MOR or SSZ setting rather than in the SCLM. Model II argues that the compositions of host peridotites control the formation and scale of podiform chromitite. Podiform chromitite mainly occurs in harzburgite, particularly in moderately depleted harzburgite ($\text{Cr}^\# = 40\text{--}60$), while lherzolite and highly depleted harzburgite have almost no or very small-scale chromitite deposits. This model is based on whether Si, Na-rich hybrid magmas could cause chromite oversaturation. In a lherzolite-magma system, where the Cr-depleted magma mixed with Si-rich magma formed by Opx decomposition, as lherzolite Opx is low in $\text{Cr}^\#$, the mixed magma will, at most, be only slightly oversaturated in chromite. In contrast, in highly depleted harzburgite, Opxs have much higher $\text{Cr}^\#$, and a melt generated by Opx dissolution will also be much richer in Cr. However, as these depleted harzburgites have a low modal abundance of Opx, the mixed magma can only supply low amounts of Cr and Al, which is also not favorable for producing large volumes of chromitite. Accordingly, moderately depleted harzburgites in MOR and SSZ, which provide both high enough $\text{Cr}^\#$ and high (Cr + Al) amounts for the precipitation of chromite, are the optimal hosts for large-scale podiform chromitites [86,91].

If the aforementioned arguments are correct, they are significant to guide the exploration of chromitite deposits. The mantle underlying mid-ocean ridges with small spreading rates is dominated by lowly depleted peridotites. According to the statistics on components of modern oceanic mantle peridotites, those oceanic lithospheres at low spreading rates are generally dominated by highly enriched lherzolite other than by depleted harzburgite [134]. In contrast, the mantle under the East Pacific Rise has high spreading rates of 128–157 mm/y dominated by harzburgite, where large-scale chromitite deposits are most likely to occur. However, this model does not systematically evaluate the genetic inconsistent relationship between degrees of host mantle peridotite depletion and precipitation of podiform chromitite. Chromitite formation is physicochemical not in equilibrium with their host peridotites. Many disseminated and dyke-shaped chromitites enclosed by thin envelopes of dunite are evidence for their simultaneous formation and a genetic relationship between them. Nevertheless, podiform massive chromitite normally cutting through the foliation of surrounding peridotites and widespread ductile deformation indicates that these chromitites formed much later than their host harzburgites [92]. In addition, high- $\text{Cr}^\#$ and high- $\text{Al}^\#$ (or low- $\text{Cr}^\#$) chromitite bodies are widely developed in the same ophiolite massif. For example, both high- $\text{Cr}^\#$ and high- $\text{Al}^\#$ chromitite orebodies are observed about 1–2 km apart from each other in the Purang ophiolite in the western segment of the IYSZ [56,92,94]. Tectonic discrimination diagrams of $\text{Cr}^\#$ versus TiO_2 and Al_2O_3 versus TiO_2 of chromite from various chromitites show that high-Al chromites plot in the field of MORB, although they are not likely derived from MORB magmas, because they display subduction-related signatures evident in depletion of Sc, Ti, V and Mn in MORB-normalized trace element patterns [92]. Actually, the usefulness of the compositions of chromite as indicators to trace the tectonic settings is questionable because the

composition of chromite can be modified by fluid/melt-rock interactions in both SSZ and MOR environments [135]. The interpretation of the tectonic setting of chromite formation is still ambiguous and needs further clarification.

Model III: The upwelling of the asthenosphere caused by the rollback, tearing and breakoff of subducted slabs in intra-oceanic subduction zones promoted the formation of UHP chromite-encapsulated diamond [92] (Figure 7d). The continuous subduction of oceanic lithospheric slabs carried oceanic and continental materials into the deep mantle. Subducted slabs successively experiencing greenschist, amphibolite and eclogite facies metamorphism may break off between the boundary of amphibolite and eclogite facies rocks, where a hybrid magma was produced through the mixing of a Cr-rich mafic magma originating from the asthenosphere with melts/fluids derived from subducted slabs. These magmas became more oxidized, more siliceous and more hydrous and rapidly triggered chromite crystallization. Tiny chromite grains suspended in the upward migrating mixed magma passed through the overlying modified mantle wedge, eventually precipitated and accumulated in the uppermost mantle near the MOHO. Some highly reduced and UHP minerals, enclosed in chromite and host mantle peridotites, were transported upwards by the uprising asthenosphere through mantle convection [92] (Figure 7d). The key to this model is that asthenosphere-sourced magmas interacted with melts/fluids from subducted slabs, leading to the formation of a water-rich, high-Mg, Si-rich and chromite-saturated hybrid magma [136]. Zhou et al. (2014) highlight that some of the UHP and highly reduced minerals were enclosed within crystallizing chromites. The large-scale chromitites, typically surrounded by envelopes of dunite, are only found in intra-oceanic SSZ ophiolites, and no major podiform chromitite bodies are developed in MORs or back-arc basins [92].

This model is of significance for understanding the multi-stage magmatic evolution in ophiolite and its relationship with the genesis of chromitite. The following issues must be resolved before this model can be considered valid: (1) can large-scale chromitite deposits form in MOR environments, and can the known small-scale chromitite bodies in modern oceans contain UHP and highly reduced minerals or not? (2) The dunite envelopes of chromitite orebodies are not always observed. For example, the chromitite bodies are in contact with their hosting harzburgites in the Luobusa and Purang ophiolites [93,94]. This observation is seemingly not accounted for by model III. (3) Hydrous mafic magmas are important for the crystallization of chromite and the formation of podiform chromitite. However, H₂O may not only come from the dehydration of subducted slabs but also from the mantle transition zone, where, at least locally, the water content hosted in wadsleyite and ringwoodite may amount up to about 1 wt% [137]. (4) The field and petrological differences between UHP chromitite and low-pressure chromitite, as well as their various hosting peridotites, should be studied in more detail.

Model IV: Ophiolite-hosted diamonds crystallized from organic carbon in subducted slab-related fluids/melts; subsequently, they were encapsulated in chromite at different depths between 150 and 660 km, and then these diamond-bearing chromites were transported to a shallow depth under MOR and SSZ environments via mantle convection or mantle plumes [4,37]. Subducted materials from oceanic lithospheres that experienced multiple episodes of subduction were incorporated into the mantle transition zone or lower mantle, resulting in the formation of diamonds and other UHP and highly reduced minerals. These minerals were captured and enclosed in high-pressure chromite at the top of the mantle transition zone. Chromitite-hosted diamond and other UHP minerals and some crustal minerals were transported upward to shallow depths by mantle convection through the action of either a mantle plume or channel flow within subduction zones (Figure 7e). Mantle peridotites experienced a widespread rock-melt reaction at shallow levels to create new chromite grains and finally coalesced with UHP chromites to form podiform chromitites in the mantle wedge of an SSZ setting [37].

This model accounts for the widespread existence of UHP and highly reduced minerals in ophiolitic mantle peridotites and chromitites and their dynamic mechanism of upward migration. It suggests that chromitite formation may initially begin within or near

the mantle transition zone, and thus podiform chromitites commonly record a crust–mantle recycling process [5,37,138,139]. Yang et al. (2021) pointed out, however, that there are still many questions remaining in the interpretation of ophiolite-hosted diamond, chromitite and their hosting peridotite formation [37]. For example, (1) how are peridotites containing UHP minerals and recycled crustal materials transported from the mantle transition zone depths to the shallow mantle where they experienced processes of partial melting and metasomatism in various geodynamic settings? Re-Os isotopes of the Luobusa chromitites show no signatures of the lower mantle and core-mantle boundary [72] and do not seem to support the hypothesis of a mantle plume that caused chromitite upward-migration. However, Middle Triassic–Early Cretaceous seamounts widely developed in the IYSZ may point to mantle plume magmatism developed in the Neo-Tethys Ocean [140–142]. Moreover, inclusions in inherited zircons from mantle peridotites and chromitites comprise typical crustal minerals, e.g., quartz, muscovite, biotite, feldspar, ilmenite, rutile and apatite [121,122]. If these zircons were enclosed in chromite grains in the mantle transition zone and upper mantle, they should have preserved some UHP-HP information on the recycling process. Until now, however, no UHP-HP records, such as coesite, garnet, etc., have been reported for the zircons. These records may still be discovered, but other mechanisms must be investigated also. (2) What are the differences between UHP and crustal mineral assemblages in the different ophiolitic belts? These differences could provide important information on the mantle dynamics in various tectonic settings. The amounts of diamonds in ophiolitic chromitites and peridotites located in the same IYSZ are obviously different; for example, over one thousand diamond grains have been extracted from each ton of the Kangjila massive chromitite and its hosting mantle peridotites in the Luobusa ophiolite [20,57]. In contrast, less than ten to dozens of diamond grains were recovered from the Dongbo, Purang, Dangqiong and Zedang ophiolites along the IYSZ [40,44] (Figure 1b). This variation is likely due to the Yarlung Zangbo Neotethyan oceanic lithosphere having experienced various degrees of mantle plume influence and SCLM contamination in multiple stages of oceanic tectonic evolution. A detailed interpretation remains a subject of debate and ongoing research. (3) What are the carbon sources of ophiolite-hosted diamonds? Their $\delta^{13}\text{C}$ values of -28.7‰ to -18.0‰ fall within the range documented for kimberlite-hosted E-type diamond (-41.3‰ to $+2.7\text{‰}$) and those diamonds discovered in UHP metamorphic zones (-30‰ to -3‰) [3,5,40,123]. Combined with Mn-bearing inclusions within diamonds, these light carbon isotopic compositions are considered as evidence of recycled crustal materials in the mantle portions of ophiolites [5,37,138,139]. Nevertheless, it is worth noting that these low $\delta^{13}\text{C}$ values also match those of some super-deep diamonds from kimberlites (-24‰ to -17‰) and carbonados that have been linked to komatiites (-27‰ to -21‰) formed in the transition zone and lower mantle [143]. Particularly, ophiolite-hosted diamonds are comparable with those diamonds from primitive chondrites (-32‰ to -38‰) [128]. In view of the lower mantle diamonds being commonly associated with Mn-bearing inclusions [144], carbon sources of the lower mantle and even the outer core cannot be ignored [6].

8. Controversy and Future Perspectives

The controversial genetic interpretation of diamond, chromitite and their host peridotite in various ophiolites not only includes the aforementioned four main models but also involves other viewpoints, e.g., UHP and highly reduced mineral assemblages possibly formed by lightning strikes [145], contamination by artificial abrasive materials [146], or formed during serpentinization under low pressure [39]. These issues have recently been hotly debated [147–150]. It is noted that Farré-de-Pablo et al. (2018) reported in situ microdiamonds (diameter = 1–8 μm) in chromite from the Tehuiztzingo chromitite in the Acatlán Orogenic Belt in southern Mexico. These diamonds were described as fracture-filling inclusions associated with serpentine, quartz, clinocllore and amorphous carbon. They were considered to have precipitated metastably at low pressure from reduced C-O-H fluids that infiltrated during the serpentinization process of the host peridotite. Massonne

(2019) argued instead that these diamonds represent contaminations introduced during the preparation of thin sections because five diamond grains are all xenomorphic in shape. Yang et al. (2019) concluded that such microdiamonds are of natural origin because they have been not only observed in serpentinized peridotite but also globally in un-serpentinized chromitites and peridotites. They proposed alternatively that microdiamonds associated with some silicates were enclosed by chromite in the deep mantle, as shown in Model IV. Farré-de-Pablo et al. (2019) replied that they had not found UHP minerals other than diamond in the Tehuiztingo chromitites, serpentinized peridotites and other rocks from the Acatlán Complex and claimed that the metamorphism of the Acatlán Complex did not even reach the stability of coesite. They argued that the aforementioned abundant so-called UHP and highly reduced minerals or oriented lamellae of Cpx in chromite developed during cooling and serpentinization rather than under UHP conditions [147].

Actually, several lines of UHP-HP evidence of the Tehuiztingo chromitite bodies and host Acatlán complex have been well studied [151–153]. The Acatlán complex, being the largest Paleozoic metamorphic complex in Mexico, records a continuous evolution from opening, subduction to the final closure of three global Paleozoic oceans, namely Iapetus, the Rheic Ocean and the Paleo-Pacific. It consists of metamorphic lithologies, including eclogite, granulite, blueschist and greenschist facies rocks, as well as phengite-garnet-rutile granitic rocks [153]. The Tehuiztingo eclogite occurs as lenses within serpentinites and meta-argillaceous rocks, which are similar to UHP metamorphic rocks documented in serpentinized peridotites, enclosed eclogite lenses within metapelites in the UHP unit from the southwestern part of the Chinese Tianshan [154]. The rock assemblages suggest that the entire assemblage of chromitite, eclogite and wallrock serpentinite likely underwent a common (ultra)high-pressure metamorphic process [152,153,155]. Moya-Carreras (2017, cited in [153]) discovered UHP metamorphic rocks associated with diamonds and metal alloys in chromitites enclosed by the Xayacatlán serpentinites in the Tehuiztingo area of the Acatlán Complex [153], implying that evidence of UHP metamorphism probably exists in the Tehuiztingo chromitite and host serpentinite.

The Tehuiztingo chromitite orebodies commonly occur as lenses with a length of several meters and a width of <2 m in serpentinite. The grain size of diamond enclosed in chromite from massive chromitite is very small (1–8 μm), which is much smaller than ophiolite-hosted diamond enclosed in other serpentinites worldwide (200–500 μm). The serpentinite is closely interlayered with eclogite and blueschist and locally contains eclogite lenses, indicating that the genesis of the diamond may differ from other ophiolite-type diamonds shown in Figure 1. Alternatively, this diamond may have formed during oceanic lithospheric slab subducted to depths where rocks experienced eclogite facies conditions and microdiamonds crystallized from C-rich fluids produced by metamorphic dehydration of slabs. This process is consistent with the formation of UHP minerals hosted in an ophiolitic podiform chromitite mélange from the Neoproterozoic metamorphic belt of the Central (Taihang) Orogenic Belt, Northern China [156]. We name this kind of microdiamond oceanic lithospheric subduction-type diamond to distinguish it from diamonds categorized as the mantle transition zone-type discovered in fresh ophiolitic mantle peridotite and chromitite. Both types of diamond are closely related to ophiolite and ascribed to the category of ophiolite-hosted diamond and differ from those of diamonds discovered in kimberlite and continental collision-related UHP metamorphic rock.

In conclusion, as outlined in the previously described four models, a consensus on the deep-mantle (over 150 km) genesis of ophiolitic massive chromitite and host peridotite was reached. However, there is still debate on the exact details of ophiolite-hosted diamond and ophiolitic chromitite formation. Potential issues include the source of the chrome, crystallization condition and position of chromite, the transport mechanism of diamond-bearing mantle peridotite and chromitite to a shallow depth, the source of carbon and formation depth of diamond, the mechanism by which crust-derived and mantle-derived materials became trapped by mantle peridotite and chromitite, the subduction zone initiation of plate tectonics, and the accumulation process of chromite as podiform chromitite. Moreover, it is

unclear whether present-day oceanic mantle peridotite and chromitite have UHP minerals, whether large-scale chromitite deposits develop in modern MOR environments, and why do some ophiolites contain chromitite orebodies while others do not, even in the same suture zone.

In summary, ophiolite-type diamond and other UHP, highly reduced unusual minerals, as well as crustal materials that originated from the mantle transition zone, have been recovered in ophiolitic chromitite and host peridotite. Peridotite and chromitite in ophiolites are not only important mineral reservoirs in the deep earth but also provide fantastic natural samples for understanding the composition of the deep earth, physical and chemical conditions and dynamic processes of recycling of crustal materials to the earth's deep mantle through subduction and their emplacement at shallow depths by mantle convection. These findings, particularly the discovery of new minerals, open a vast new area of research on the genesis of ophiolitic chromitite, tectonic setting of ophiolites in orogenic belts worldwide and crust–mantle materials recycling.

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