INVITED REVIEW

Origin of hotspots in the South Pacific: Recent advances in seismological and geochemical models

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In this study, we review recent seismological and geochemical studies of the hotspots in the South Pacific superswell. Extensive studies on global-scale seismic tomography have revealed the presence of slow seismic-velocity anomalies in the lower mantle beneath the superswell region although the size and vertical extent of these anomalies are not well constrained. In the last decade, regional seismic networks deployed on islands and the seafloor on the superswell have enabled detailed imaging of the mantle structure. Results show that the low-velocity superplume extended from the core–mantle boundary to a depth of 1000 km, and the low-velocity narrow plumes extend from the top of the superplume toward the South Pacific hotspots. Geochemical studies have suggested that dehydrated subducted oceanic crust is involved in the formation of the HIMU lavas, but the trace element and isotopic composition of HIMU lavas cannot be explained by the inclusion of this crust. Recent studies have revealed that the oceanic crust should sink into the lowermost mantle and melt, which would metasomatize the ambient mantle to form the HIMU reservoir. We present a model incorporating a thermochemical superplume and secondary narrow plumes generated from the superplume; our model can explain seismic structure, geochemical heterogeneities of ocean island basalts in the superswell region, age progression of the South Pacific hotspots, and massive eruption and formation of large oceanic plateaus in the Cretaceous period.

Keywords: hotspot, South Pacific, seismic tomography, geochemical heterogeneities, HIMU, thermochemical superplume

INTRODUCTION

The South Pacific seafloor is characterized by a broadly elevated seafloor (the South Pacific superswell) (McNutt and Judge, 1990; McNutt, 1998; Adam and Bonneville, 2005) and a concentration of volcanic island chains (Fig. 1; in the present study, we define the following regions of volcanism in the South Pacific: Marquesas to the north, Pitcairn to the east, Macdonald to the south, and Rarotonga to the west). In these oceans, the rate of volcanism is 3–4 times greater than that in other oceans, and 14% of the active hotspots of the Earth are concentrated in an area covering less than 5% of the globe. Most of the island chains have been recognized as products of areas of intraplate volcanism, known as hotspots. A hotspot chain is characterized by the spatial age progression along a chain, anomalously shallow topography around the chain, and geochemically enriched basalts distinct from mid-oceanic ridge basalts (MORBs). The age progressions of hotspot chains in the South Pacific cover relatively short periods (<20 Myr according to Bonneville

et al., 2002, 2006). The hotspots produce oceanic island basalts (OIBs) with an enriched mantle signature (e.g., Hart, 1984), spanning a wider range of radiogenic isotope compositions than basalts in other regions (e.g., Vidal *et al.*, 1984). Seismological studies have commonly shown large areas of pronounced low seismic velocity in the lower mantle beneath the South Pacific Ocean and the African continent (e.g., Lay, 2007). All these observations suggest the presence of mantle upwelling beneath the South Pacific region.

Morgan (1971) proposed that hotspot volcanism is generated by mantle plumes, involving the rapid ascent of unusually hot and buoyant mantle materials from deep thermal boundary layers in the lower mantle. In the present study, the boundary between the upper and lower mantle is assumed to be at a depth of 660 km, with the mantle delineated by the Moho discontinuity and the core mantle boundary (CMB). The mantle transition zone (MTZ) comprises the lower third of the upper mantle (410–660 km depth). Several models, modified from the classical-plume model, have been proposed to explain the South Pacific and African hotspots (e.g., Davaille, 1999; Torsvik *et al.*, 2006; Suetsugu *et al.*, 2009); such models are commonly based on the notion that the South Pacific hotspots

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Fig. 1. The South Pacific superswell region. The stars denote hotspots. Bathymetry contours of 1000 m interval are shown. The superswell is indicated by dashed curve, which represents seafloor elevation of 300 m (Adam and Bonneville, 2005).

have a deep origin in the lower mantle. Many non-plume hypotheses have suggested mechanisms that allow intraplate volcanism to be attributed to the shallower mantle: small-scale convection cells in the asthenosphere (e.g., Ballmer *et al.*, 2010); upwelling induced by asthenospheric shear (e.g., Conrad *et al.*, 2011); the influence of the structure and stress state of the lithosphere (e.g., McNutt *et al.*, 1997; King and Anderson, 1998; King and Ritsema, 2000); and volcanism as a precursor to future breakup of the Pacific plate (e.g., Clouard and Gerbault, 2008). Although some of the South-Pacific volcanism requires the influence of lithosphere and asthenosphere, deep-seated mantle plumes are still the most plausible explanation for the hotspot volcanism in this region (e.g., Konter *et al.*, 2008) as we will elucidate on the basis of recent advances in seismic imaging and geochemical studies.

Volcanism in South Pacific was far more extensive in the past than that at present, thereby producing large oceanic plateaus in the Cretaceous period (e.g., Utsunomiya *et al.*, 2007). These phenomena led to the assumption of a large mantle plume (superplume) in the deep mantle beneath the South Pacific region (Larson, 1991) although its size, depth as well as presence remain controversial. If the superplume proves to exist, it would have a great impact on our understanding of mass transfer in the mantle, the heat balance of the Earth, and the Earth's evolution. Previous studies on superplumes have been limited by sparse geophysical, petrological, and geochemical data

in this remote oceanic region. However, in the last decade, geochemical and seismological observations and analyses have advanced substantially in terms of accuracy and spatial resolution.

The purpose of this study is to review recent advances in understanding the origin of the South Pacific OIB and recent refinements of seismological images of the mantle plumes beneath the South Pacific superswell region. In addition, we will propose a hypothesis that links the geochemical and seismological anomalies of the superswell region.

HOTSPOT VOLCANISM AND SEAFLOOR SWELL IN SOUTH PACIFIC

In the South Pacific, at least four volcanic chains are recognized to have been formed by hotspot activity (Fig. 1): Society, Marquesas, Pitcairn–Gambier, and Cook–Austral. The Society Islands consist of aligned volcanic islands and seamounts that exhibit linear relationships between age and locality. Maupiti, on the northwest end of the island chain, is the oldest island (4–5 Ma ago). The present position of the hotspot is manifested by the active Mehetia Seamount, located approximately 150 km to the southeast of Tahiti (Devey *et al.*, 1990; Binard *et al.*, 1993; Guillou *et al.*, 2005). The Pitcairn–Gambier hotspot has been active for 13 Ma, expressing its volcanic activity in the Mururoa and Fangataufa atolls, Gambier, Pitcairn (Island) and the presently active Pitcairn Seamount (Guillou *et al.*, 1994; Devey *et al.*, 2003). The age–distance relationships of these volcanic chains are consistent with a theory in which fixed mantle plumes formed volcanic islands and seamounts on the moving Pacific plate. The Marquesas Islands exhibit an age progression from 5.8 Ma at the northwest end to 0.4 Ma at the southeast end of the island chain (Brousse *et al.*, 1990; Desonie *et al.*, 1993). However, the Marquesas Islands are atypical in the following respects: (1) no currently active seamount is found; (2) the volcanic lineation trending N140–150°E is oblique to the direction of plate motion (N110–120°E) (Legendre *et al.*, 2005); and (3) this lineation is broad and apparently defines two subparallel volcanic arrays that are geographically and geochemically distinct, such as the Hawaiian Islands (Desonie *et al.*, 1993; Huang *et al.*, 2011). To account for these facts, several competing models have been proposed that consider regional tectonic controls, such as fractures in the lithosphere that may affect upward magma transport from the mantle plume to the surface and deflection of the mantle plume by westward upper mantle flow beneath Marquesas (McNutt *et al.*, 1989; Desonie *et al.*, 1993; Clouard and Bonneville, 2005).

The volcanism in the Cook–Austral Islands is even more complicated. The classical interpretation suggests

Fig. 2. Map of the Cook–Austral Islands with the ages of islands and seamounts. At least four distinct hotspot tracks are superimposed on each other in the Cook–Austral volcanic chain (gray lines) on the basis of age–location relationships and geochemical compositions. In particular, the trend from Macdonald to Neilson Bank (southern Austral chain; Macdonald trend) and that from Raivavae to Rurutu (northern Austral chain; Raivavae trend) would be formed by distinct mantle plumes, as demonstrated by the geographical offset and different geochemical signatures (Bonneville *et al.*, 2006). It is uncertain whether Mangaia is a part of the volcanic chain of the Raivavae trend with HIMU signature, or it is an extension of the volcanic chain starting from Macdonald Seamount. Older seamounts with ages >20 Ma are not shown in the age–distance diagram. Age data are from Dalrymple *et al.* (1975), Duncan and McDougall (1976), Bellon *et al.* (1980), Turner and Jarrard (1982), Maury *et al.* (1994), and Bonneville *et al.* (2006).

that most of the islands and seamounts were created on the moving plate by a single fixed hotspot, which is currently located at Macdonald Seamount (Fig. 2). However, some lavas on the Cook Islands (the western part of the island chain) exhibit younger ages than those expected on the basis of hypothesis of a single fixed hotspot (Dalrymple *et al.*, 1975; Duncan and McDougall, 1976; Bellon *et al.*, 1980; Turner and Jarrard, 1982; Maury *et al.*, 1994). Moreover, young volcanism on Rurutu is observed to have occurred after a 9-Ma hiatus following the cessation of the old volcanism, the latter of which is plotted on the general age trend in the age–distance diagram (Fig. 2). These facts imply the existence of two additional mantle plumes that formed the Atiu trend and Rarotonga trend, although the activity related to the latter may have commenced very recently (Chauvel *et al.*, 1997). Age determinations had been restricted to subaerial lavas in 1990s (except for Macdonald Seamount), but the presence of Atiu trend was confirmed by the discovery of the Arago Seamount dated at 0.2 Ma; therefore, this seamount

must be near the present position of the hotspot forming the Atiu trend (or Arago trend; Bonneville *et al.*, 2002). Bonneville *et al.* (2006) further suggested that the northern and southern Austral Islands may represent distinct trends because of the geographical offset of the chains between Raivavae and Neilson Bank and the distinct geochemical features of the trends in the northern and southern Austral Islands. Possible interpretations of the complicated emplacement of these island chains will be discussed later.

The South Pacific is characterized by elevated seafloor at two different horizontal scales: the swell in the vicinity of the hotspots (known as hotspot swell, 800–1000 m in the vertical direction, 100–200 km the horizontal direction) and the superswell that is broadly distributed over the South Pacific (700–1100 m in the vertical direction, 3000 km in the horizontal direction). This area subsides less rapidly away from the East Pacific Rise than predicted by any thermal subsidence model of oceanic lithosphere (McNutt and Fischer, 1987). The hotspot swell

can be attributed to the dynamic support of the ascending plume in the upper mantle beneath the hotspots (Adam *et al.*, 2010). The superswell extends between latitudes 10°N and 30°S and longitudes 130°W and 160°W on a seafloor displaying ages between 30 and 115 Ma (Adam and Bonneville, 2005). It has a hemispheric shape composed of two branches. The southern branch corresponds to the location of the South Pacific region defined in the present study, where active volcanism is observed as described above. The northern branch is not clearly correlated with active volcanic features. Moreover, the superswell area is associated with a geoid anomaly. The geoid is an equipotential surface that corresponds to the mean surface of oceans; departures from this reference provide information on mass repartition in the Earth's interior. The correlation between the depth anomaly associated with the superswell and the geoid is still debated upon as it is unclear which long-wavelength gravitational field best represents the superswell (Hager, 1984; Watts *et al.*, 1985; McNutt and Judge, 1990; Adam and Bonneville, 2005). Recently, Cadio *et al.* (2011) applied a wavelet method to analyze the GRACE gravity data. The analysis showed that the northern and southern branches of the superswell are associated at intermediate wavelengths with a geoid high (~10 m) and a geoid low (~5 m), respectively, which can be attributed to thermochemical anomalies in the lower mantle.

SEISMOLOGICAL IMAGING OF THE MANTLE BENEATH THE SOUTH PACIFIC

Seismic images obtained from the global data

Seismic tomography is the most powerful tool available to determine the mantle structure beneath the South Pacific to identify the origin of the hotspots. This technique is based on the same principle as that of a medical CT scan and provides an image of the heterogeneous structure of the Earth's interior using huge amounts of seismic data. Methods using travel times of seismic waves and waveforms themselves are known as travel-time and waveform-inversion techniques, respectively. The higher temperatures of presumed mantle plumes should appear as regions of lower seismic-wave velocity. The spatial resolution of a seismic image, obtained by tomography, depends on the coverage of seismic waves for a target region. Regions with large numbers of earthquakes and many seismograph stations, e.g., the Japanese islands, are well resolved by seismic tomography and those with few earthquakes and stations, e.g., the South Pacific, are poorly resolved. Since the oldest generation of global mantle tomography (e.g., Dziewonski, 1984; Inoue *et al.*, 1990), two large low-seismic-velocity anomalies in the lowermost mantle have commonly been imaged beneath the South Pacific and Africa (Lay, 2007) although the size

and vertical extent of the anomalies, deduced by different tomographic studies, have been inconsistent. Some of the representative tomographic images from previous studies are compared in Fig. 3. Fukao (1992) and Zhao (2004, 2009) resolved a large-scale (1000–1500 km in diameter) anomaly of P-velocities slower by 1–2% throughout the mantle by performing a ray-theoretical travel-time inversion on global P-wave arrival time data from the International Seismological Centre (ISC) (Fig. 3a). Montelli *et al.* (2006) showed conduit-like low-velocity anomalies (~300 km in diameter, slower by ~1% and ~2% for P and S velocities, respectively) throughout the mantle by using finite-frequency, travel-time tomography (Dahlen *et al.*, 2000). The low-velocity conduits ascend from broad low-velocity anomalies immediately above the CMB. Takeuchi (2007, 2009) obtained a similar image of the S-velocity structure from the waveform inversion of long-period transverse records (slower by 1.5–2%). On the other hand, some tomographic studies showed a large (700–1000 km in diameter), dome-like low-S-velocity anomaly (slower by 1.5–2%), occurring only in the lowermost mantle (Masters *et al.*, 2000; Ritsema and van Heijst, 2000) or extending from the CMB to the middle-lower mantle (Figs. 3b and c) (e.g., Mégnin and Romanowicz, 2000; Grand, 2002; Ritsema *et al.*, 2010). Ritsema's *et al.* (2010) model shows low-velocity anomalies in the top half of the lower mantle (Fig. 3c), whereas Grand's (2002) model does not (Fig. 3b). Romanowicz and Gung (2002) also determined a global three-dimensional (3D) attenuation model for the upper mantle, showing a strong attenuation in the upper mantle beneath the South Pacific. In addition, the attenuation was studied by analyzing reverberated ScS waves (traveling through the whole mantle with multiple reflections at the Earth's surface and the CMB) beneath the South Pacific, suggesting that the region of strong attenuation extended into the lower mantle although its vertical extent was not constrained (Suetsugu, 2001). These inconsistencies, particularly for the vertical extent of the slow anomalies, arise mainly due to the scarcity of earthquakes and seismic stations in the region. The lateral resolution of previous tomographic studies was limited to 500–1000 km for P-velocity and 1000–2000 km for S-velocity. This raised a question that the presence of single large slow anomalies might be smearing artifacts due to insufficient resolution beneath the South Pacific. Schubert *et al.* (2004) proposed that the cluster of narrow plumes could be poorly imaged as a single large slow anomaly because of poor seismic resolution. Recently, Bull *et al.* (2009) constructed two synthetic seismic models on the basis of geodynamic models with a cluster of narrow slow anomalies and two broad slow anomalies beneath the South Pacific and Africa; then, they used the models to compute synthetic seismic data for realistic earthquake-station configuration and

Fig. 3. Seismic structure obtained from global tomographic studies. (a) P-velocity structure by P-wave travel-time tomography (Zhao, 2004, 2009); (b) S-velocity structure by S-wave travel-time tomography (Grand, 2002); (c) S-velocity structure by waveform inversion (Ritsema et al., 2010). The mantle cross sections along a line are shown in (d). Green stars in (d) denote active hot spots. Two-letter labels on the stars are the names of the hot spots: SM, Samoa; RT, Rarotonga; SC, Society; AG, Arago; MD, Macdonald; MQ, Marquesas; PT, Pitcairn.

performed seismic tomography. The tomographic images from the model with the broad slow anomalies exhibited a better fit to the actual tomographic model (Ritsema and van Heijst, 2000). Most tomographic studies have relied on seismic waves, generated at remote hypocenters, which travel long distances and are recorded at distant seismic stations. Distortion of the signals due to lateral heterogeneities away from the superswell region has prevented the generation of a definitive image of the low-velocity anomalies.

In addition to the large-scale low-velocity anomalies in the lower mantle discussed above, a thin layer associated with substantial reduction in seismic velocity has been detected at the base of the lower mantle beneath the South Pacific, first observed by Garnero *et al.* (1993). This is known as the ultra-low-velocity zone (ULVZ), which forms a layer 5–40 km thick over the CMB and results in P-wave and S-wave velocity reductions of 10% and 30%, respectively (e.g., Garnero *et al.*, 1998; Lay, 2007), which has been confirmed using various types of seismic waves. However, this is not a global feature. The ULVZ is found more beneath the Pacific Ocean as com-

pared to the circum-Pacific and continental regions. The substantial reduction in seismic velocities suggests the occurrence of partial melting in the ULVZ (Williams and Garnero, 1996) although contributions of chemical anomalies, e.g., enrichment of iron, cannot be precluded.

Mapping topography of mantle discontinuities from global data

The MTZ is bounded by two seismic discontinuities: the 410-km and 660-km discontinuities (hereafter referred to as “410” and “660”). These discontinuities are now broadly accepted as representing the mineral-phase transition from olivine in the upper mantle to wadsleyite at the 410 and from ringwoodite to perovskite and magnesiowüstite at the 660 (e.g., Helffrich, 2000). Because these phase transitions are pressure- and temperature-dependent and the 410 and 660 have opposite Clapeyron slopes, determining the depths of the two discontinuities provides useful information regarding temperature anomalies within the MTZ (e.g., Irifune *et al.*, 1998; Katsura *et al.*, 2004). For instance, if a hot plume ascends from the lower to the upper mantle, the MTZ

Fig. 4. Map of the South Pacific superswell showing the thickness obtained by SS-precursor analysis (Niu *et al.*, 2002). Black “+” signs represent the bounce points of SS-waves at the 410. Hotspots are denoted by stars.

should be thinner than normal because the 410 and 660 are due to exothermic (positive Clapeyron slope) and endothermic (negative Clapeyron slope) phase transformations, respectively, and the depths of phase transformation are controlled by temperature. The depths of 410 and 660 and the thickness of the MTZ can be determined by analyzing weak seismic signals reflected (e.g., Flanagan and Shearer, 1998) or converted from P- to S-waves or S- to P-waves (e.g., Vinnik *et al.*, 1997). We can obtain the depths of reflection/conversion from the measurement of the time interval between directly arriving waves and reflected/converted waves at a seismic station. Niu *et al.* (2002) analyzed differential travel times between an SS-wave and its underside reflections from the 410 and 660 beneath the South Pacific superswell. The average MTZ thickness beneath the superswell region is 237 km, which is near the average value of 240 km found in this study. In addition, Niu *et al.* (2002) found that the MTZ is approximately 30 km thinner locally beneath the Society hotspot than the global average thickness (Fig. 4). The MTZ is not thinned beneath the other hotspots in the superswell area although the coverage of the SS precursors is more limited around the other hotspots than that around the Society hotspot. In addition, Suetsugu *et al.* (2004) analyzed reverberated ScS-waves and converted Ps-waves using broadband oceanic island stations in the South Pacific. They found no broad, hot MTZ beneath the superswell region. Both studies analyzed long-period SS-waves and reverberated ScS-waves with periods longer than 25 s; this long wavelength may smoothen out detailed topography and prevent some features from being identified. Mantle discontinuities have also been mapped in the South Pacific region by using the receiver-function method of converted Ps-waves us-

Fig. 5. BBOBS (large solid triangles) and PLUME (open triangles) stations shown on a bathymetric map in the French Polynesia region. Small solid triangles denote permanent seismic stations. Stars denote active hotspots.

ing global network data (Vinnik *et al.*, 1997; Li *et al.*, 2003; Suetsugu *et al.*, 2004). The Society and Pitcairn hotspots exhibit a thinner MTZ than the average one (Li *et al.*, 2003; Suetsugu *et al.*, 2004). When the above receiver-function studies were conducted, the topography of the discontinuities was not determined for the entire South Pacific region because the area sampled by the Ps-waves was restricted by the small number of seismic stations.

Seismic images from the regional data

To overcome the problem of poor resolution of global seismic data beneath the South Pacific superswell, two regional seismic experiments have been performed within the last decade: the Polynesian Lithosphere and Upper Mantle Experiment (PLUME) project (Barruol *et al.*, 2002) and the Polynesia Broadband Ocean Bottom Seismograph (BBOBS) array (Suetsugu *et al.*, 2005) (Fig. 5).

Under the PLUME project, 10 broadband stations were temporarily installed on the oceanic islands of French Polynesia in 2001 and operated between 2001 and 2005. Maggi *et al.* (2006a, b) obtained a Pacific-scale S-velocity model from surface-wave tomography using land-based data, including PLUME data (Fig. 6). Their model showed deep-rooted low-velocity zones (~2% reduction) beneath the Society and Macdonald hotspots but only superficially rooted (<150 km) low-velocity zones

Fig. 6. Seismic structure obtained from regional S-wave tomography by using land-based data, including those from the PLUME network (Maggi *et al.*, 2006a). The positions of the profiles are provided in the bottom right panel. The symbol convention is same as in Fig. 3.

beneath the Marquesas. The low-velocity anomalies, shown by Maggi *et al.* (2006a), are 700–1000 km in diameter. However, Fig. 6 clearly demonstrates that these anomalies in the superswell region are confined to localized structures and are not pervasive throughout the area. The resolution achieved by PLUME stations is approximately 800 km.

Maggi *et al.* (2006b) and Fontaine *et al.* (2007) investigated the anisotropy in the Pacific upper mantle and its relation to mantle plumes. Fontaine *et al.* (2007) analyzed SKS splitting from the island stations in the superswell region. SKS splitting is a phenomena of separated arrivals of two orthogonally-polarized SKS wave (a type of S-waves propagated in the mantle and outer core). This splitting represents the anisotropy in the mantle beneath the stations. The results of Fontaine *et al.* (2007) indicated that this anisotropy in the asthenosphere can be explained by the mantle flow due to Pacific plate motion in most of the superswell region, with the flow locally dis-

turbed by a mantle plume near the Society hotspot. This conclusion was confirmed later by additional data from BBOBS array (Barruol *et al.*, 2009). Maggi *et al.* (2006b) obtained the azimuthal anisotropy of long-period Rayleigh waves in the Pacific. Moreover, they showed that the general pattern of anisotropy is well correlated with the mantle flow due to Pacific plate motion in the asthenosphere; however, the correlation is broken at hotspots, including those in French Polynesia, supporting the SKS-splitting results of Fontaine *et al.* (2007). The interaction of plate-motion-driven horizontal flow and mantle-plume vertical flow generated a parabolic flow pattern in the asthenosphere (Sleep, 1990; Ribe and Christensen, 1994), which deviated from the plate-motion-driven flow pattern and was consistent with the anomalous anisotropy beneath the hotspots. The PLUME network was the first passive seismic experiment to investigate the South Pacific mantle plumes and improved seismic coverage in the region.

To further improve the seismic coverage provided by the permanent and temporary land-based stations, 10 seismometers were deployed on the French Polynesian seafloor as part of the Japan–France cooperative Polynesia BBOBS array (Fig. 5) (Suetsugu *et al.*, 2005). In contrast to land-based stations, which are restricted to oceanic islands and clustered in a nonuniform manner, the BBOBS observations had the advantage of flexibility in selecting station sites; thus, the BBOBS stations were located to supplement the existing stations on oceanic islands. The observation period of the BBOBS project overlapped with that of the PLUME project for approximately two years. The BBOBS system was developed by the Earthquake Research Institute of the University of Tokyo, starting in the 1990s (Kanazawa *et al.*, 2001). Isse *et al.* (2006) and Suetsugu *et al.* (2009) determined the 3D S-velocity variations by analyzing surface waves. They identified slow anomalies of approximately 2–3% near the Society, Macdonald, Pitcairn, and Marquesas hotspots at depths of 200 km. The lateral resolution is about 500 km.

Suetsugu *et al.* (2007, 2009) mapped the topography of the MTZ by using the receiver-function method. The depths of the 410 and 660 beneath the stations were estimated and used to determine the MTZ thickness. The average MTZ thickness beneath the South Pacific (240 km) is approximately equal to the global average (241–243 km) reported in other studies (Flanagan and Shearer, 1998; Gu *et al.*, 1998), suggesting that the average temperature in the MTZ beneath the South Pacific superswell is normal. Notable exceptions are the locally thinned MTZ beneath the Society hotspot (216–219 km), which suggests that hot plumes may ascend to this hotspot from the lower mantle. Temperature anomalies are estimated to be 150–200 K, assuming the experimentally determined Clapeyron slope of the phase transformations (e.g., Irifune *et al.*, 1998; Katsura *et al.*, 2004). The stations near the Macdonald and Pitcairn hotspots may also have a moderately thinned MTZ beneath them (227 km and 225 km, respectively), but the other hotspots are not underlain by thin MTZs.

Body-wave tomography is a useful technique for studying mantle plumes in the lower mantle because body waves from remote earthquakes, recorded at the PLUME and the Polynesia BBOBS stations, traveled in the lower mantle beneath the superswell. We used the travel times measured at PLUME and BBOBS stations together with the ISC P-wave arrival time data for global mantle ray-theoretical tomography (e.g., Fukao *et al.*, 2003). The spatial resolution achieved by our calculations in the lower mantle is estimated to be ~500 km. The resolution in the MTZ and upper mantle is much lower, which prevents us from obtaining well-resolved images above the 660. Low-velocity anomalies are found throughout the lower man-

Fig. 7. Maps showing lateral variations in P-wave velocity in the lower mantle at depths of 629–712 km, 800–893 km, 1095–1203 km, and 1317–1435 km (Suetsugu *et al.*, 2009). Deviations from the average one-dimensional structure beneath the Pacific Ocean are shown. Green stars denote hotspots.

tle beneath the superswell, but the lateral extent of the anomalies changes drastically below a depth of 1000 km (Fig. 7). Below 1000 km, slow anomalies of 1% are as large as 3000 km × 3000 km, an area that previous studies referred to as the South Pacific superplume (Fukao, 1992; Mégnin and Romanowicz, 2000; Zhao, 2004, 2009). These anomalies appear to continue down to CMB. On the other hand, above 1000 km, the slow anomalies split into narrower and more localized anomalies a few hundred kilometers in diameter. In addition, the large-scale, low-velocity anomaly was found by Tanaka *et al.* (2009a, b), who used the BBOBS and PLUME data for regional P- and S-wave travel-time tomography.

The vertical profiles of the P-wave seismic images in the lower mantle are presented together with those of the upper mantle S-wave seismic structure and the MTZ thicknesses in Fig. 8. We plot the S-velocity model from the global model S20RTS (Ritsema and van Heijst, 2000) in the MTZ since no tomography model based on the PLUME and BBOBS data is available. The most remarkable feature of Fig. 8 is the abrupt change in the scale of the slow anomalies within the lower mantle: they are 1000–2000 km wide at depths greater than 1000 km and only a few hundred kilometers wide above that depth. This observation suggests that the narrow, slow anomalies above 1000 km may originate at the top surface of a broad superplume. Moreover, the distribution of the MTZ thickness supports the presence of these narrow anomalies: two

Fig. 8. Cross sections of the seismic structure for the entire mantle (Suetsugu et al., 2009). S-wave velocities are shown for the upper mantle (0–410 km). P-wave velocities are shown for the lower mantle (660–2900 km). The model by Ritsema and van Heijst (2000) is shown for the MTZ structure. Velocity scales are $\pm 3\%$ in the upper mantle and $\pm 0.75\%$ in the lower mantle. The positions of the profiles are provided in the bottom right panel. Circles plotted in the MTZ (410–660 km) are the locations of the MTZ thickness estimates near the profiles within 2.5° , where red and blue represent the data thinner and thicker than the global average, respectively, and the size is proportional to the deviation from the average. The symbol convention is same as in Fig. 3.

significantly thin MTZ areas beneath the Society hotspots are located near the narrow anomalies at the top of the lower mantle. We ascribe these anomalies to two continuous plumes that ascend from the top surface of the superplume to the Society hotspot (Fig. 8a). This geometry also occurs beneath the Macdonald hotspot (Fig. 8c). In contrast, the Pitcairn hotspot is characterized by a slow upper mantle and thin MTZ but without slow anomalies at the top of the lower mantle (Fig. 8a). Therefore, now, the Pitcairn hotspot might not be deep rooted in the lower mantle. The slow anomalies beneath the Marquesas hotspot appear to be restricted to the upper mantle, and

there is no evidence of slow anomalies at the top of the lower mantle beneath this region (Fig. 8b). Therefore, we consider that there are no plumes rising continuously from the lower mantle to the Marquesas hotspot, which is consistent with a previous geodetic study (McNutt and Bonneville, 2000). In summary, the South Pacific region has mantle plumes with different maximum depths: the Society and Macdonald hotspots are fed by mantle plumes originating at the top of the superplume, whereas the other hotspots are not. In the next section, we discuss the geochemical signatures of the volcanic rocks at the hotspots and their depths of origin. By combining two

independent approaches, seismology and geochemistry, should help us to construct a complete picture of the South Pacific mantle plumes and the origin of the chemical heterogeneities in the mantle.

GEOCHEMICAL SIGNATURE OF OCEAN ISLAND BASALTS IN THE SOUTH PACIFIC ISLANDS AND THE ORIGIN OF THE MANTLE RESERVOIRS

Petrological features

The volcanic rocks in the Society, Pitcairn–Gambier, and Cook–Austral Islands are dominated by alkalic lavas, including alkali basalt, mugearite, hawaiite, trachyte nephelinite and phonolite. Transitional basalt is observed occasionally, but tholeiitic basalt is very rare (except for drilled lavas from Fangataufa atoll in Gambier Islands; Dupuy *et al.*, 1993). On the other hand, rocks from Marquesas include tholeiitic basalt together with transitional basalt and alkalic lavas (Desonie *et al.*, 1993; Legendre *et al.*, 2005). The rarity of tholeiitic basalt in the Society, Pitcairn–Gambier, and Cook–Austral Islands contrasts with the composition of large-scale hotspot volcanoes in other regions, such as Hawaii, Reunion, and Iceland, where tholeiitic basalt is the main rock type during the shield-building stage with subordinate pre-shield and post-shield alkalic lavas. It must be noted that the studied rocks in the South Pacific are mostly subaerial rocks that would have erupted in later stages of volcano growth, which may correspond to the post-shield stage or the later part of the shield-building stage. The currently active and growing seamounts at Mehetia (Society), Pitcairn, and Macdonald (Austral) could be the only manifestation of the inception stage of the volcano growth. Nonetheless, the dredged rocks from these seamounts are predominantly alkalic lavas and their differentiates (Devey *et al.*, 1990, 2003; Hémond *et al.*, 1994; Hekinian *et al.*, 2003). Moreover, lavas collected from the submarine ridges of Rurutu, Tubuai, and Raivavae (Austral Islands) are all alkalic and transitional lavas (Hanyu *et al.*, 2009). So far, we have no evidence for tholeiitic shield-stage lavas in the Society, Pitcairn–Gambier, or Cook–Austral Islands, and further drilling and submarine surveys are required to elucidate the magmatic evolution of the ocean islands. Post-shield stage lavas have been known to occur after a hiatus of thousands–millions of years following the main shield stage of many volcanic islands (e.g., Pitcairn, Tahiti, Marquesas Islands; Duncan and McDougall, 1974; Duncan *et al.*, 1994; Legendre *et al.*, 2005). Such rocks are generally highly alkalic and differentiated.

Alkalic lavas are the product of low-degree partial melting under high-pressure conditions with garnet residue. This is consistent with the trace element features such as the enriched large-ion lithophile elements and elevated

LREE/HREE ratios (light/heavy rare earth elements), common in lavas in the South Pacific. Furthermore, these rocks exhibit major element compositions different from MORB. For example, the lavas referred to as HIMU (high- μ ; see below) from the Austral Islands are enriched in FeO, TiO₂, and CaO and depleted in SiO₂, K₂O, and Al₂O₃ relative to MORB (Kogiso *et al.*, 1997a; Jackson and Dasgupta, 2008). Such melts could not have been produced by partial melting of volatile-free peridotite such as MORB source mantle (e.g., Hirose and Kushiro, 1993), even by assuming an incipient melt produced from garnet peridotite (Davis *et al.*, 2011). The source of OIB must be more fertile than MORB mantle, as discussed later.

Geochemical heterogeneity in the South Pacific

OIB in the South Pacific are geochemically unique because they exhibit wide variations in isotopic and trace element compositions (e.g., Vidal *et al.*, 1984; Nakamura and Tatsumoto, 1988; Weaver, 1991; Hémond *et al.*, 1994) (Fig. 9). The isotopic variation of OIB documents the existence of several enriched mantle reservoirs, referred to as HIMU, EM1, and EM2 (EM; enriched mantle) together with DM (depleted mantle) and PM (primitive mantle), in the source region of mantle plumes (e.g., Zindler and Hart, 1986; Hofmann, 1997) (Fig. 9). HIMU is defined by very radiogenic Pb isotopic compositions, reflecting high time-integrated μ values ($\mu = {}^{238}\text{U}/{}^{204}\text{Pb}$). The main geochemical feature of HIMU is low ${}^{208}\text{Pb}/{}^{204}\text{Pb}$ for a given ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ and unradiogenic ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ similar to MORB. Both EM1 and EM2 show enriched isotopic characteristics with low ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ and variously radiogenic ${}^{87}\text{Sr}/{}^{86}\text{Sr}$, but EM2 exhibits higher ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ than EM1. A significant difference between these two endmembers is highlighted by their Pb isotopes. EM1 has the least radiogenic Pb isotopic compositions among OIB, whereas EM2 has moderately high ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ and remarkably elevated ${}^{208}\text{Pb}/{}^{204}\text{Pb}$ for a given ${}^{206}\text{Pb}/{}^{204}\text{Pb}$.

Many OIB are the products of mixing of components from several mantle reservoirs; therefore, they are plotted between the mantle reservoirs in multi-isotope spaces (Fig. 9). However, some lavas in the South Pacific show extreme isotopic compositions close to the enriched mantle reservoirs, such as HIMU, EM1, and EM2 reservoirs. The Pitcairn–Gambier Islands is the type-locality for the EM1 reservoir, together with Walvis Ridge and the Tristan and Gough in the South Atlantic. Along the island chain, the influence of EM1 reservoir appears to increase temporarily from the Gambier, Mururoa and Fangataufa atolls toward Pitcairn Island and Pitcairn Seamounts (Dupuy *et al.*, 1993). On the Pitcairn Island, lavas with a strong EM1 flavor have been found in the Tedsite Formation underlying the post-shield lavas (Eisele *et al.*, 2002). The lavas from Pitcairn Seamounts located 70–100 km to the southeast of Pitcairn Island show isotopic compositions

Fig. 9. Sr, Nd and Pb isotopic compositions for OIB in the South Pacific. The Cook–Austral Islands exhibit both HIMU and EM1 isotopic signatures. The Pitcairn–Gambier and Society Islands exhibit isotopic trends toward EM1 and EM2, respectively. Lavas from the Marquesas show diverse isotopic compositions between EM2 and DM. The extent of isotopic variation in this region is as large as that defined by OIB worldwide. Data are compiled from the GEOROC database (<http://georoc.mpch-mainz.gwdg.de/georoc/>; compiled in 2012).

partly overlapping those from Pitcairn Island, but the most extreme EM1 signature is inherited in the differentiated lavas from small volcanic edifices among the seamount group (Woodhead and Devey, 1993; Devey *et al.*, 2003).

The Society and Marquesas lavas, together with Samoa in the southwest Pacific, exhibit broad isotopic trends that point to the EM2 reservoir (Duncan *et al.*, 1994; White and Duncan, 1996). Neither the Society nor Marquesas Islands show systematic geochemical variation with ages. However, their geochemical compositions change during the growth of a single volcano. At Tahiti

in the Society Islands, the EM2 signature is generally observed in the shield stage lavas, whereas the post-shield stage lavas are more alkalic and isotopically depleted; this is similar to the temporal geochemical variations observed in the Hawaiian volcanoes (Duncan *et al.*, 1994). In contrast, the EM2 signature is more robust in the post-shield lavas than in the shield stage lavas at Marquesas Islands (Castillo *et al.*, 2007).

Although the component with radiogenic Pb isotopes is apparently involved in many OIB, the occurrence of “pure” HIMU lavas (e.g., $^{206}\text{Pb}/^{204}\text{Pb} > 20.5$) is restricted to the Cook–Austral Islands and the island of St. Helena in the Atlantic Ocean. In the Cook–Austral Islands, lavas with HIMU signature only occur before 6 Ma on Mangaia, Tubuai, Rurutu, and Raivavae (Nakamura and Tatsumoto, 1988; Chauvel *et al.*, 1992; Hémond *et al.*, 1994; Woodhead, 1996; Kogiso *et al.*, 1997a; Schiano *et al.*, 2001; Lassiter *et al.*, 2003; Hanyu *et al.*, 2011a). Temporal isotopic variation within individual volcanoes has not been identified in these islands. Pb, Nd, Sr, and Hf isotopic variations of the lavas with robust HIMU signature (Mangaia, Rurutu (older lavas), Tubuai) define broad mixing trends (Fig. 10). Isotopic analyses of clinopyroxene phenocrysts further document the well-defined mixing trends between the HIMU reservoir and the depleted component by taking advantage of the lower susceptibility to alteration and late-stage assimilation for clinopyroxenes compared to whole-rock samples (Hanyu and Nakamura, 2000; Jackson *et al.*, 2009; Hanyu *et al.*, 2011a). This indicates that the HIMU lavas were formed by melts derived from the HIMU reservoir and the depleted component mixed in various proportions. Moreover, they were free from melt from the EM1 reservoir despite EM1 being common in lavas younger than 6 Ma in the Cook–Austral Islands (Figs. 2 and 10).

Inter- and intra-island geochemical variations

Erupted OIB do not always reflect the exact geochemical compositions of the mantle reservoirs. The geochemical variation along a hotspot track and within a single volcano must reflect source heterogeneity, melting conditions of the source materials, and interaction between the mantle plume and shallow lithospheric and asthenospheric materials. Consequently, understanding the inter- and intra-island geochemical variation is essential to explore the true composition of the mantle reservoirs, which are estimated by the extrapolation of geochemical trends defined by the lavas.

Overall isotopic trends observed in lavas from the Pitcairn–Gambier and Society Islands demonstrate mixing of either EM1 or EM2 components with other components. Devey *et al.* (2003) studied dredged submarine lavas from the Mehetia and Pitcairn seamounts and found that the most extreme isotopic compositions appear in

Fig. 10. Multi-isotope diagrams for the HIMU lavas from Mangaia, Rurutu, Tubuai, and Raivavae. Open and solid symbols denote the isotope data for whole-rock samples and clinopyroxene separates, respectively. Note that clinopyroxene data display well-defined isotopic trajectories, which are interpreted to be two-component mixing trends between the HIMU reservoir and the depleted component, because clinopyroxenes are less susceptible to crustal contamination and post-eruption alteration than whole rocks (Hanyu *et al.*, 2011a). Light gray fields indicate global MORB–OIB arrays. Dark gray fields indicate East Pacific Rise (EPR) MORB that are a proxy for the local lithosphere and asthenosphere beneath the studied region in terms of the isotopic composition. The northern hemisphere reference line (NHRL) in panels (a) and (b) is from Hart (1984). The compiled data are from the GEOROC database.

low-degree melts from small volcanic edifices, whereas the lavas from larger seamounts have more variable but less extreme isotopic compositions. Therefore, Devey *et al.* (2003) suggested that EM1 and EM2 mantle reservoirs have the lowest melting temperatures of the components involved, and the first products of melting from the mantle plume exhibit the most enriched isotopic compositions. On the other hand, the late stage lavas tend to have depleted isotopic compositions relative to the shield stage lavas as observed at Tahiti, even if they were produced by low-degree melting (Duncan *et al.*, 1994). This indicates the absence of EM2 in the source region of the late stage lavas. In comparison with the Hawaiian case, this may be explained by a radially zoned mantle plume with a central zone consisting of material from the EM2 reservoir in the deep mantle and an outer rim containing entrained depleted asthenospheric material. The early and main shield stage lavas were derived from the hot central zone, whereas the late stage lavas were derived from the

outer rim of the plume. However, the Marquesan volcanoes show the opposite geochemical variation from the shield stage to the late stage. The late stage lavas exhibit isotopic compositions more like EM2 than those of the shield stage lavas; therefore, the late stage lavas could represent the primary composition of the mantle source (Castillo *et al.*, 2007). It is suggested that the mantle plume beneath Marquesas is much weaker than that beneath Hawaii (and possibly Society), and the shield stage lavas are substantially affected by the assimilation of lithospheric material with depleted isotopic compositions (Caroff *et al.*, 1999; Legendre *et al.*, 2005; Castillo *et al.*, 2007).

The effect of lithospheric assimilation on lava chemistry has been also advocated in other OIB in the South Pacific, suggested by the heterogeneous Pb isotopic composition recorded in melt inclusions. For example, olivine-hosted melt inclusions in Mangaia lavas show a very heterogeneous Pb isotopic composition with variations far

beyond the Pb isotopic range of the host lavas, extending from HIMU-like composition toward a composition more similar to a depleted mantle source (Saal *et al.*, 1998, 2005; Yurimoto *et al.*, 2004). Saal *et al.* (2005) preferred a model in which the depleted melt is derived from oceanic lithosphere. The lithospheric melt could have been incorporated into the HIMU melt during its upward migration from the mantle plume through the overlying lithosphere, and then incompletely mixed melts could have been conceivably trapped in crystallizing olivine phenocrysts. However, it must be noted that there are other possible interpretations. Yurimoto *et al.* (2004) proposed an alternative scenario based on their findings that sulfide-rich inclusions always exhibit the most radiogenic Pb isotopes, whereas silicate inclusions show wide isotopic variation. Because sulfide fractionation must occur continuously and simultaneously with olivine fractionation in HIMU magmas, lack of sulfide-rich inclusions with less-radiogenic Pb isotopes is not indicative of the incorporation of lithospheric melt into the magmas. Yurimoto *et al.* (2004) suggested that the depleted melt trapped in melt inclusions can be derived from the mantle plume along with the HIMU melt. In fact, such Pb isotopic heterogeneity recorded in melt inclusions has been recently contested by Paul *et al.* (2011); thus, further studies are required to assess possible small-scale geochemical heterogeneity in OIB.

Origin of the EM1 and EM2 mantle reservoirs

The formation of chemically distinct mantle reservoirs is a source of longstanding debate in mantle geochemistry (e.g., Zindler and Hart, 1986; Hofmann, 1997). Although this issue has been investigated using global OIB geochemical data, studies related to hotspot lavas in the South Pacific have made significant contributions. Several models have been presented for the origin of HIMU, EM1, and EM2 reservoirs: recycling of surface materials (oceanic crust, continental crust, and sediment) with or without chemical modification by alteration and subduction processes, remelting of recycled metasomatized lithospheric mantle, and intra-mantle differentiation of these subducted materials.

Several lines of evidence suggest that some portion of the mantle is fused to form fractionated melt, which subsequently migrates to metasomatize other mantle portions (e.g., Pearson *et al.*, 1991; Pilet *et al.*, 2005). This process could viably create enriched mantle components, because low-degree metasomatic melts transport highly incompatible elements more effectively than moderately incompatible to compatible elements. Such melts could have wide compositional variation depending on source composition, source mineralogy, degree of melting, and presence of volatiles. Niu and O'Hara (2003) argued that the enrichment in incompatible elements in conjunction

with radiogenic Sr and unradiogenic Nd isotopes in enriched OIB (EM1 and EM2) can be explained by a lithospheric source refertilized by low-degree metasomatic melts.

However, the isotopic compositions of some mantle reservoirs resemble those of oceanic crust, continental crust, and sediment; therefore, some geochemists favored a model that considers the subduction of these surface materials (e.g., Zindler and Hart, 1986; Weaver, 1991; Hauri and Hart, 1993; Woodhead and Devey, 1993; Eisele *et al.*, 2002; Stracke *et al.*, 2003; Jackson *et al.*, 2007). For instance, isotopic compositions of EM2 overlap those of terrigenous sediments, derived primarily from the continental crust. Sr–Nd–Hf isotopic characteristics of EM1 appear to be similar to those of pelagic sediments, and unradiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ in the EM1 reservoir can be ascribed to low U/Pb ratios in the sediments (Woodhead and McCulloch, 1989; Woodhead and Devey, 1993; Dostal *et al.*, 1998; Eisele *et al.*, 2002). An alternative model for EM1 origin is the delamination of subcontinental lithospheric mantle, which also possesses isotopic characteristics similar to the EM1 reservoir (McKenzie and O'Nions, 1983; Tatsumi, 2000; Gibson *et al.*, 2005).

Discrimination between the proposed models may require an additional geochemical evidence coupled with constraints from Sr–Nd–Pb–Hf isotopes and trace element compositions. Oxygen is a good indicator of recycled materials, including fresh and altered oceanic crust, and sediment, because oxygen isotope ratios are susceptible to weathering, alteration, and precipitation that occur near the Earth's surface (Eiler, 2001; Day *et al.*, 2009). Relative enrichment in ^{18}O is clearly observed in EM2 samples from Aitutaki and Tahaa in the Cook–Austral Islands, Moorea in the Society Islands and Savaii in the Samoan Islands (Fig. 11; Eiler *et al.*, 1997). Elevated $^{18}\text{O}/^{16}\text{O}$ in Samoan lavas was further confirmed by Workman *et al.* (2008), who suggested that sediment recycling is the only process that can account for enriched ^{18}O in OIB. In contrast, EM1 lavas from Pitcairn exhibit $^{18}\text{O}/^{16}\text{O}$ overlapping with the range of MORB values; therefore there is no indication of sediments with $^{18}\text{O}/^{16}\text{O}$ or altered oceanic crust with low $^{18}\text{O}/^{16}\text{O}$ in the EM1 reservoir (Eiler *et al.*, 1997).

Oceanic crust and sediment are highly enriched in Re but depleted in Os; therefore, recycled crust and sediment would attain high $^{187}\text{Os}/^{188}\text{Os}$ after long-term isolation (Roy-Barman and Allègre, 1995; Becker, 2000; Dale *et al.*, 2007). EM2 lavas exhibit relatively low $^{187}\text{Os}/^{188}\text{Os}$ compared to other OIB (Hauri and Hart, 1993; Reisberg *et al.*, 1993; Workman *et al.*, 2004). Indeed, Os isotopic composition is susceptible to crustal assimilation, but the relatively low $^{187}\text{Os}/^{188}\text{Os}$ characteristics of EM2 are confirmed by Os isotope measurements of olivine phenocrysts in Samoan lavas (Jackson and Shirey, 2011). This fact is

Fig. 11. $^{206}\text{Pb}/^{204}\text{Pb}$ versus (a) helium isotopes ($^3\text{He}/^4\text{He}$), (b) oxygen isotopes ($^{18}\text{O}/^{16}\text{O}$; $\delta^{18}\text{O}$), (c) Li isotopes ($^7\text{Li}/^6\text{Li}$; $\delta^7\text{Li}$), and (d) Os isotopes ($^{187}\text{Os}/^{188}\text{Os}$). (a) The HIMU lavas, except for Raivavae, define a clear two-component mixing trend between the HIMU reservoir and the depleted lithospheric or asthenospheric component. Error bars denote 1σ uncertainties. Data sources are Parai et al. (2009) and Hanyu et al. (2011b). (b) $\delta^{18}\text{O}$ values of olivine separates of HIMU (Austral; solid symbols), EM1 (Pitcairn), and EM2 (Society) lavas are shown. The gray line indicates the range of $\delta^{18}\text{O}$ of N-MORB. The ranges of $\delta^{18}\text{O}$ values of hydrothermally altered MORB at low and high temperatures are shown by arrows. $\delta^{18}\text{O}$ values of sediments are between 15 and 25. Data sources are Eiler et al. (1995, 1997). (c) Open and solid symbols for Mangaia, Tubuai, and Raivavae lavas denote plots for whole rock samples and olivine separates, respectively. The ranges of $\delta^7\text{Li}$ values of hydrothermally altered MORB at low and high temperatures are shown by arrows. The gray line indicates the range of $\delta^7\text{Li}$ of N-MORB. Data sources are Nishio et al. (2005), Vlastélic et al. (2009), and Chan et al. (2009). (d) Open and solid symbols for Mangaia, Rurutu and Tubuai lavas denote plots for whole rock and olivine separates, respectively. Note that Os isotope analyses with olivine separates yield more reliable data than those with whole rock analyses (Hanyu et al., 2011a; Jackson and Shirey, 2011). The data for HIMU lavas do not fit the mixing lines between the depleted mantle and the ancient subducted oceanic crust (solid lines), the latter of which has extremely high $^{187}\text{Os}/^{188}\text{Os}$ (>1) and low Os/Pb relative to the former. Consequently, the HIMU reservoir cannot be the oceanic crust itself. An alternative model is that the HIMU reservoir formed by the hybridization of subducted oceanic crust with an ambient mantle. The hybrid material is assumed to be a mixture of 40% oceanic crust and 60% mantle material in this model (large solid circles). The hybrid material is plotted on the mantle–oceanic crust mixing line (solid lines), because we assumed a large degree of melting of the subducted oceanic crust prior to hybridization with the mantle material. Mixing between the melt from such a hybrid material and the melt from a depleted mantle (either lithosphere or asthenosphere) results in concave-down mixing trends (dashed lines). The mixing calculations are after Hanyu et al. (2011a). Data sources are Hauri and Hart (1993), Reisberg et al. (1993), Roy-Barman and Allègre (1995), Schiano et al. (2001), Lassiter et al. (2003), and Hanyu et al. (2011a).

not obviously concordant with a model that considers terrigenous sediment contribution to the EM2 reservoir based on substantial enrichment in Sr–Nd–Pb isotopes and elevated $^{18}\text{O}/^{16}\text{O}$. Jackson and Shirey (2011) suggested that the EM2 reservoir was formed by the mixing of subducted terrigenous sediment with the peridotitic mantle.

Conversely, EM1 lavas from Pitcairn exhibit elevated $^{187}\text{Os}/^{188}\text{Os}$. Eisele et al. (2002) ruled out the recycling of subcontinental lithosphere as an EM1 source, because it has $^{187}\text{Os}/^{188}\text{Os}$ equivalent to or even lower than MORB and bulk silicate Earth values. They preferred the incorporation of pelagic sediment in the EM1 reservoir because of high $^{187}\text{Os}/^{188}\text{Os}$ associated with unradiogenic Pb iso-

Fig. 12. Trace element abundances normalized to primitive upper mantle (McDonough and Sun, 1995). Typical trace element abundances for HIMU lavas (Mangaia), EM1 lavas (Pitcairn), and EM2 lavas (Tahaa) are after White and Duncan (1996), Eisele *et al.* (2002), and Hanyu *et al.* (2011a). The compositions of N-MORB (Hofmann, 1988), bulk oceanic crust (Stracke *et al.*, 2003), and typical sediment (GLOSS; Plank and Langmuir, 1998) are shown for comparison. Trace element abundances of dehydrated bulk oceanic crust are calculated using mobility coefficients for dehydration (Kogiso *et al.*, 1997b).

topes. However, this hypothesis contradicts oxygen isotope evidence that is not indicative of sediment in the EM1 reservoir. No model proposed so far seems to successfully account for all the geochemical features of EM1; thus, this issue remains to be resolved in future studies.

Origin of the HIMU reservoir

Two major models have been proposed for the origin of HIMU reservoir: recycled metasomatized lithosphere and subducted oceanic crust. HIMU lavas are discriminated from EM1 and EM2 lavas on the basis of very radiogenic Pb and less-radiogenic Sr isotopes, suggesting long-term storage of a HIMU reservoir with high (U+Th)/Pb and low Rb/Sr. Moreover, the relative depletion in strongly incompatible elements, in the order Rb, Ba, Th, U, and Nb, is an indigenous feature of HIMU lavas (Fig. 12) (Weaver, 1991; Kogiso *et al.*, 1997a; Willbold and Stracke, 2006). Such a geochemical signature can be reproduced by remelting of amphibole veins in metasomatized lithosphere (Pilet *et al.*, 2005, 2008). Although this model is viable, the compositions of low-degree metasomatic melts would be heterogeneous and the scale of such processes would be variable. HIMU lavas from different domains of the mantle (the Cook–Austral Islands in the South Pacific and St. Helena in the Atlantic) have very similar isotopic and elemental composi-

tions; this suggests that the HIMU reservoir must be geochemically very uniform (Stracke *et al.*, 2005).

Alternative models for the origin of HIMU reservoir consider chemical modification of the oceanic crust by hydrothermal alteration and subsequent dehydration during subduction (Hofmann and White, 1982; Weaver, 1991; Chauvel *et al.*, 1992; Woodhead, 1996; Kogiso *et al.*, 1997a; Tatsumi and Kogiso, 2003; Stracke *et al.*, 2005; Hanyu *et al.*, 2011a; Kawabata *et al.*, 2011). High-pressure experiments have demonstrated that K, Pb, Rb, and Ba are highly fluid-mobile elements, in contrast to moderately mobile LREE and Sr and less-mobile Nb, U, and Th (Keppler, 1996; Kogiso *et al.*, 1997b). Consequently, dehydrated altered oceanic crust is expected to display trace element characteristics indigenous to HIMU lavas together with high (U+Th)/Pb and low Rb/Sr.

The HIMU lavas in the Cook–Austral Islands show higher $^{187}\text{Os}/^{188}\text{Os}$ than bulk silicate Earth and MORB (Hauri and Hart, 1993; Reisberg *et al.*, 1993; Schiano *et al.*, 2001; Lassiter *et al.*, 2003; Hanyu *et al.*, 2011a), consistent with the involvement of the recycled oceanic crust in the HIMU reservoir (Fig. 11). Hauri and Hart (1993) further argued against shallow metasomatic process for the generation of HIMU because metasomatized xenoliths generally have low Re/Os and $^{187}\text{Os}/^{188}\text{Os}$. Stable isotopes of oxygen and lithium can be fractionated by hydrothermal processes. Slightly lower $^{18}\text{O}/^{16}\text{O}$ and higher $^7\text{Li}/^6\text{Li}$ than MORB for HIMU lavas indicate the involvement of hydrothermally altered oceanic crust in the HIMU reservoir (Eiler *et al.*, 1997; Nishio *et al.*, 2005; Chan *et al.*, 2009) (Fig. 11).

Evidence of crust recycling may be provided by noble gases. Helium isotopes could potentially distinguish components involved in the mantle plume, such as depleted upper mantle ($^3\text{He}/^4\text{He} \sim 8 \text{ Ra}$; Ra is atmospheric $^3\text{He}/^4\text{He}$), primitive deep mantle ($>35 \text{ Ra}$), and recycled crustal material ($<1 \text{ Ra}$). The HIMU lavas from the Cook–Austral Islands (and those from St. Helena) display distinctly lower $^3\text{He}/^4\text{He}$ than MORB and many OIB with elevated $^3\text{He}/^4\text{He}$ (Graham *et al.*, 1992; Hanyu and Kaneoka, 1997; Parai *et al.*, 2009) (Fig. 11). Indeed, shallow crustal contamination may be responsible for some OIB with low $^3\text{He}/^4\text{He}$ (Hilton *et al.*, 1995). However, a clear correlation of $^3\text{He}/^4\text{He}$ with $^{206}\text{Pb}/^{204}\text{Pb}$ for most of the HIMU lavas precludes the low- $^3\text{He}/^4\text{He}$ signature being the result of shallow crustal contamination (Hanyu *et al.*, 2011b). Because the HIMU reservoir is plotted at the radiogenic Pb isotopic end of the mixing trend, $^3\text{He}/^4\text{He}$ must be less than 6 Ra, consistent with the oceanic crust with low (U+Th)/He being the precursor of the HIMU reservoir.

However, pure recycled oceanic crust without any modification in its composition may not be the source of HIMU, because partial melting of basaltic oceanic crust

(eclogite) produces Si-rich melt, whereas HIMU lavas are undersaturated in SiO_2 . As an alternative model, high-pressure experiments demonstrate that partial melting of silica-deficient pyroxenite produces melts with compositions similar to those of primary alkalic HIMU lavas in terms of major elements (Hirschmann *et al.*, 2003; Kogiso *et al.*, 2003; Kogiso and Hirschmann, 2006). Such pyroxenite can be produced from eclogite by the extraction of Si-rich fluid/melt during subduction or more possibly be formed by the hybridization of subducted eclogite with ambient peridotitic (or pyrolitic) mantle. Mismatch with respect to Al_2O_3 , TiO_2 , and K_2O between the primary lavas and experimental melts from silica-deficient pyroxenite cannot be disregarded (Pilet *et al.*, 2008; Gerbode and Dasgupta, 2010). This problem could be reconciled by considering the presence of carbon dioxide in the peridotite or pyroxenite source material (Dasgupta *et al.*, 2007; Gerbode and Dasgupta, 2010).

Whether the HIMU reservoir was formed by the hybridization of subducted oceanic crust with ambient mantle rather than simple accumulation and isolation of oceanic crust may be constrained by Os isotopic composition. Contrary to Pb, Nd, Sr, and Hf isotopic variations, which reflect two-component mixing of the HIMU reservoir and a lithospheric or asthenospheric component, the HIMU lavas exhibit uniform and moderately radiogenic $^{187}\text{Os}/^{188}\text{Os}$, as demonstrated by isotopic analyses with whole rocks (Schiano *et al.*, 2001) and olivine phenocrysts (Hanyu *et al.*, 2011a) (Fig. 11). This contradicts a model in which the HIMU reservoir consists of pure recycled oceanic crust without modification, because aged oceanic crust (>0.3 Ga) has very radiogenic $^{187}\text{Os}/^{188}\text{Os}$ (>1) owing to very high Re/Os (Becker, 2000). Moreover, because the oceanic crust (and its melt) is much more depleted in Os (20–50 ppm) than peridotitic mantle (2000–3000 ppm), mixing between the oceanic crust (or its melt) and peridotitic mantle does not reproduce uniform and moderately radiogenic $^{187}\text{Os}/^{188}\text{Os}$ together with variably enriched Pb and Hf isotopes (solid lines in Fig. 11(d)). Alternatively, Hanyu *et al.* (2011a) proposed a two-step process in which (i) the HIMU reservoir was formed by the hybridization of subducted oceanic crust and mantle material several billion years ago, and (ii) such a hybrid material was transported by an upwelling mantle plume toward the base of the lithosphere, where it melted and mixed with another melt derived from the depleted mantle component. In this process, Re/Os of hybrid material was moderately increased by the addition of melt from the oceanic crust to the mantle material, and therefore it acquires moderately high $^{187}\text{Os}/^{188}\text{Os}$ by radiogenic ingrowth over several billion years. In addition, because such a hybrid material can accommodate high Os concentrations, the recent mixing between the hybrid material and the depleted source (or their melts) results in the

production of HIMU lavas with uniform $^{187}\text{Os}/^{188}\text{Os}$ and variable Pb and Hf isotope ratios observed in HIMU lavas.

Where do the mantle reservoirs exist?

It has been a fundamental question; how was the heterogeneous mantle, consisting of chemically distinct mantle reservoirs, formed and how have these mantle reservoirs been maintained in the convecting mantle (e.g., Zindler and Hart, 1986; Hofmann, 1997). The classical two-layer mantle model may no longer be acceptable, because the upper–lower mantle boundary cannot be a barrier to material transport in a convecting mantle. If subduction and delamination of slabs and continental materials play a major role in creating mantle heterogeneity, such materials may be trapped in the mid-mantle (e.g., the MTZ) (Hanan and Graham, 1996) or dispersed throughout the mantle (Christensen and Hofmann, 1994; Phipps Morgan and Morgan, 1999). However, some part of the subducted slab could sink into the lower mantle because its density is greater than the surrounding pyrolitic material in most part of the mantle (Ono *et al.*, 2001; Hirose *et al.*, 2005), eventually accumulating and forming a sizable reservoir above CMB. The materials in such a layer may be isolated for geologically long periods in the convecting mantle (Kellogg *et al.*, 1999; Tackley, 2000; McNamara *et al.*, 2010).

From a geochemical point of view, attempts have been made to pin the location of mantle reservoirs down to the lower mantle above the core. Potential tracers for core–mantle interaction include $^{182}\text{W}/^{184}\text{W}$ and $^{186}\text{Os}/^{188}\text{Os}$, because these isotope ratios should differ between the silicate mantle and the metallic core (Brandon *et al.*, 1998; Kleine *et al.*, 2002). Although $^{186}\text{Os}/^{188}\text{Os}$ have not been determined for OIB in the South Pacific, coupled enrichment in $^{186}\text{Os}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ for Hawaiian and Icelandic lavas suggests that the associated mantle plumes are derived from the bottom of the lower mantle (Brandon *et al.*, 1998, 2007). However, the Hawaiian and Icelandic lavas lack a $^{182}\text{W}/^{184}\text{W}$ signature that would imply contributions from the core (Scherstén *et al.*, 2004). Apparent decoupling between Os and W may be explained by assuming that either W is a less-sensitive tracer than Os for detecting core material or that elevated $^{186}\text{Os}/^{188}\text{Os}$ in the Hawaiian and Icelandic mantle plumes cannot be ascribed to a core component but comes from recycled Mn-rich sediments (Scherstén *et al.*, 2004; Arevalo and McDonough, 2008). W isotopic data of HIMU lavas from the Cook–Austral Islands have been presented by Takamasa *et al.* (2009). They report $^{182}\text{W}/^{184}\text{W}$ for HIMU lavas as being indistinguishable from mantle values; therefore, clear evidence of interactions between mantle reservoirs and the core has not yet been provided from the study of OIB in the South Pacific.

Considering Nf–Hf–Pb isotope systematics, Hanyu *et*

al. (2011a) suggested a possible lower mantle origin for the HIMU reservoir on the basis of the hybridization model. Hybridization would most likely be triggered by partial melting of the oceanic crust and subsequent mixing with the ambient mantle in a metasomatic manner. Such melting could potentially occur in the upper mantle, MTZ, or lower mantle. It is expected that Lu/Hf, Sm/Nd, and U/Pb in the partial melt change depending on the mineral assemblages in the subducted oceanic crust that vary according to pressure. Therefore, the depth of partial melting of subducted oceanic crust might be constrained by assessing whether such melt reproduces the present-day Hf, Nd, and Pb isotopic compositions of the HIMU reservoir (Hanyu *et al.*, 2011a). Under upper mantle conditions, partial melting of the oceanic crust in eclogite or pyroxenite facies drastically reduces Lu/Hf in the melt because of the presence of garnet, thereby predicting too low $^{176}\text{Hf}/^{177}\text{Hf}$ in the melt (Fig. 13). In the MTZ, Lu/Hf fractionation in majoritic garnet is moderate relative to that in low-pressure garnet, but such melts apparently do not reproduce Hf–Nd–Pb isotopic compositions similar to those of the HIMU reservoir (Fig. 13).

In fact, it is difficult to initiate the partial melting of dehydrated oceanic crust, because the dry solidus of the oceanic crust is much higher than the geotherm in most parts of the mantle (Zerr *et al.*, 1998; Ono, 2008). An exception to this may be in the lowermost mantle, where a sharp temperature increase is expected above the core (Ono, 2008; Fiquet *et al.*, 2010). Partition coefficients for perovskite determined by recent experiments allow us to infer element fractionations associated with possible partial melting in the lower mantle (Hirose *et al.*, 2004; Walter *et al.*, 2004; Corgne *et al.*, 2005). Collerson *et al.* (2010) suggested that the melting residue from the primitive lower mantle material with Ca–perovskite could have parent/daughter ratios relevant to the HIMU reservoir (note that they did not consider the recycled material). Alternatively, Hanyu *et al.* (2011a) proposed that the partial melting of oceanic crust with residual Mg–perovskite and subsequent metasomatism with the ambient mantle could be responsible for the fractionated parent/daughter ratios required for the HIMU reservoir (Fig. 13). These models imply that both slab subduction and deep melting processes are important for generating mantle heterogeneity.

THE ORIGIN OF SOUTH PACIFIC HOTSPOTS: THERMOCHEMICAL MANTLE PLUMES AND SUPERPLUME

In the preceding sections, we discussed how seismological and geochemical characteristics are related to the origin of the South Pacific hotspots. We have presented recently obtained seismic images showing the low-

Fig. 13. Calculated present-day isotopic compositions of recycled materials: N-MORB, altered MORB, bulk oceanic crust, partial melts of bulk oceanic crust with eclogite facies (in the upper mantle), garnetite facies (in the MTZ), and perovskite facies (in the lower mantle). The model bulk oceanic crust is assumed to be a subducted slab package, consisting of 25% N-MORB, 25% altered MORB, and 50% gabbro after Stracke *et al.* (2003). Each recycled material was partially melted with a non-modal fractional melting scheme and was formed at 2 Ga. For simplicity, the residual phases of modeled eclogite, garnetite, and perovskite are assumed to consist of 50% clinopyroxene + 50% garnet, 100% majoritic garnet, and 100% Mg-perovskite, respectively. Numbers on the curves denote degrees of melting in percentage. The HIMU reservoir should be plotted on the extension of the trends defined by the HIMU lavas; therefore, partial melting with residual perovskite is the process that most likely causes parent/daughter fractionation responsible for the isotopic compositions of the HIMU reservoir. Bulk silicate Earth (BSE) isotopic compositions are denoted by thin crossed lines.

velocity superplume extending from the CMB to 1000 km depth and the narrow low-velocity plumes above this depth. It has to be analyzed whether the seismic anomalies can have a thermal origin. Laboratory and numerical experiments have shown that a purely thermal plume is characterized by a mushroom-shaped model, with a broad plume head at the top and narrow plume stem below (e.g., Turcotte and Schubert, 2002). This is in contrast to our seismic images in which the narrow plumes ascend from the top of the superplume. Conversely, as shown by the geochemical studies reviewed in the preceding geochemistry section, basalts in the superswell region have highly heterogeneous geochemical signatures. The HIMU rocks found in the Cook–Austral hotspot chain may have their origin in the lowermost mantle. Therefore, it is reasonable to assume that the superplume and narrow plumes are thermochemical.

Davaille (1999), Davaille *et al.* (2002), and Kumagai *et al.* (2007) performed laboratory experiments in tanks using two chemically distinct layers, with the lower layer being slightly denser and more viscous. The lower layer developed a large dome that protruded into the upper layer as the bottom of the lower layer was heated. Narrow plumes were occasionally generated and ascended from the roof of the dome in a manner similar to the superplume and narrow plumes in recently obtained seismic images (Fig. 8). Moreover, a dome structure has been produced by 2D and 3D thermochemical numerical simulations (McNamara and Zhong, 2004; Farnetani and Samuel, 2005; Nakagawa and Tackley, 2004, 2005; Ogawa, 2007; Tan and Gurnis, 2007). Nakagawa and Tackley (2005) suggested that the exothermic perovskite to post-perovskite transformation (Murakami *et al.*, 2004; Oganov and Ono, 2004) destabilizes the thermochemical boundary layer above the CMB and enhances the development of the thermochemical dome. The secondary narrow plumes are generated in sporadically and are not temporally persistent (Davaille, 1999; Ogawa, 2007), which could explain why the age progressions of the South Pacific hotspots cover relatively short periods (Duncan and McDougall, 1976; Clouard and Bonneville, 2005).

There has been much debate regarding the generation process of the secondary narrow plumes from the superplume. As mentioned above, Davaille (1999) and Kumagai *et al.* (2007) argued that these secondary plumes are generated by thermal instability on the roof of the hot superplume. Interestingly, Kumagai *et al.* (2007) showed that the secondary narrow plumes were generated from the rim of the roof of the superplume (referred to as the lower dome in their study), thereby forming a ring-shaped plume, which is similar to the narrow low-velocity anomalies at and immediately above the superplume (Fig. 7). An alternative view of the relative locations of the plume and superplume has been proposed. Torsvik *et al.* (2006,

2008, 2010) paleomagnetically reconstructed the sites of the large igneous provinces (LIPs) and major hotspots that have erupted over the last 200–300 Myr and compared these reconstructions to the seismic images of broad low-velocity anomalies in the lowermost mantle. They found that most LIPs and many of the hotspots were located near the margins of the broad slow anomalies beneath Africa and the South Pacific and not in the center of the regions. Their correlation was good, particularly for the African slow anomaly, but it was less for the South Pacific. Many South Pacific hotspots, discussed in the present study, are not necessarily located away from the margins of the South Pacific slow anomaly. In particular, the seismically imaged secondary plumes to the Society and Macdonald hotspots are located above the superplume (Fig. 8), which is consistent with the hypothesis that the secondary plumes are generated from the top of the superplume. The spatial relationship between the superplumes and narrow plumes appears to be different beneath the South Pacific compared to that beneath Africa, although the reasons for this occurrence are yet to be found.

Nakagawa and Tackley (2005) and Ogawa (2007) included harzburgite and basalt components in their simulation. Dense basaltic components sank to CMB as a subducted crust, isolated for a long time, and were then swept toward the upwelling material and brought up in the upwelling hot dome. Their simulation results were consistent with the conclusion that the HIMU reservoir is most possibly a hybrid material, produced by the mantle metasomatism of a partial melt from the subducted oceanic crust. At the bottom of the mantle beneath the South Pacific, simultaneous analysis of the geoid and very long-period seismic waves detected a high-density anomaly, which was possibly of chemical origin (Ishii and Tromp, 1999). McNamara *et al.* (2010) performed a high-resolution simulation of the thermochemical mantle convection and stated that the ultra-low-velocity and high-density zones beneath the South Pacific and Africa (the ULVZ) can survive in the convecting mantle for several hundred million years. The ULVZ may represent the HIMU reservoir, isolated from the ambient mantle for geologically long periods of time.

Figure 14 illustrates a schematic of the subducted slabs, superplume, and narrow plumes beneath the South Pacific. The slab with altered and dehydrated oceanic crust is subducted into the MTZ; it stagnates for a while, and then, penetrates into the lower mantle because of its low temperature (e.g., Fukao *et al.*, 2009). The altered and dehydrated subducted oceanic crust sinks down to the CMB because of its negative buoyancy (Hirose *et al.*, 2005; Ono *et al.*, 2005). A part of the oceanic crust may be detached from the main body of the subducted slab, as suggested by seismic scatterers commonly found beneath

Fig. 14. Schematic model illustrating the origin of the South Pacific hotspots. Dehydrated subducted oceanic crust sinks to CMB because of its negative buoyancy; some of the oceanic crust may be detached from the subducted slab and detected as scatterers beneath the circum-Pacific subduction zones. The oceanic crust melts in the lowermost mantle, which metasomatizes the ambient mantle to form the HIMU reservoir. The HIMU reservoir is then swept into the large thermochemical superplume developed from the CMB to a depth of ~1000 km. From the top of the superplume, secondary narrow plumes ascend to some of the Polynesian hotspots (e.g., Society, Macdonald). The secondary plume that has formed the Cook–Austral hotspot chain entrained the HIMU pieces in the superplume, which may be an origin of the HIMU lava in the Cook–Austral chain. Each narrow plume may sample a different volume with different geochemical characteristics and exhibit different geochemical signatures at different hotspots in the South Pacific. Small symbols in the superplume and narrow plumes are the distinct geochemical components (HIMU, EM1, and EM2). The slow-seismic-wave anomalies beneath the hotspots other than the Society and Macdonald hotspots may represent remnant secondary plumes, which have detached from the superplume.

circum-Pacific subduction zones (Kaneshima and Helffrich, 2010). The oceanic crust melts in the lowermost mantle, and this melt metasomatizes the ambient mantle to form the HIMU reservoirs. Such a metasomatized mantle would become denser than the surrounding unmetasomatized mantle because of iron enrichment in the oceanic crust-derived melt (Knittle, 1998). Because of their dense nature, the HIMU reservoirs could be stabilized and isolated at the bottom of the mantle for a geologically long time (e.g., McNamara *et al.*, 2010). The potential HIMU reservoir could then be swept into the thermochemical superplume, extending from the CMB to a depth of ~1000 km. From the top of the superplume, secondary narrow plumes ascend to some of the South Pacific hotspots (e.g., Society, Macdonald). Once the secondary plumes impinge upon the shallow upper mantle, the plumes can interact with the lithosphere/asthenosphere, which may explain the complicated emplacement of several island chains in the Cook–Austral islands. McNutt *et al.* (1997) emphasized the importance of structural weakness (e.g., fracture) or local stress fields in the lithosphere for the origin of the volcanism in the Cook–Austral chains. Turner and Jarrard (1982) preferred the “hot-line” hypothesis rather than the hotspot hypothesis. This concept has recently been developed by Ballmer *et al.* (2010); the dense distribution and apparent long life of volcanoes may be explained by the elongated hot

region in the direction of the plate motion, induced by small-scale sublithospheric convection triggered by a hot plume or upwelling. Much higher resolution seismic imaging is required to understand the interaction of the secondary plume and lithosphere/asthenosphere.

The secondary plumes that formed the Cook–Austral hotspot chains entrained the HIMU components involved in the superplume, which may be the origin of the HIMU lavas in the Cook–Austral chain. Each narrow plume may sample volumes with different geochemical characteristics, thereby resulting in different geochemical signatures (e.g., trace element and isotopic composition) at different hotspots in the South Pacific (Zindler and Hart, 1986; Bonneville *et al.*, 2006). We speculate that all distinct geochemical components, including EM1 and EM2, may be present in the superplume although the depth of origin and formation processes for the EM1 and EM2 lavas are yet to be determined. The narrow plumes that formed the Cook–Austral hotspot chain may have entrained the pieces of HIMU and EM1 components, explaining the two types of lavas found there. The slow anomalies beneath the hotspots other than the Society and Macdonald hotspots are not connected to the lower mantle slow anomalies but may represent remnant secondary plumes, which have detached from the superplume. Alternatively, they may be the result of upwelling of shallow origin that is unrelated to the secondary plume because the direction of the

Marquesas island chain differs from that of the Pacific plate motion by 20–30°. A discontinuity in the structure of the lithosphere, e.g., a fracture zone, may induce local upwelling to form the volcanism of the Marquesas Islands (McNutt and Bonneville, 2000; King and Ritsema, 2000).

The model presented in Fig. 14 is similar to a hotspot type proposed by Courtillot *et al.* (2003) as one of three hotspot types (the other two being hotspots associated with plume ascent directly from CMB and hotspot with surficial origin, e.g., lithosphere breakup). In their proposed model, the thermochemical dome is developed from the CMB to the 660 and gets stuck there because of negative buoyancy acting on the superplume due to an endothermic phase transformation (Vinnik *et al.*, 1997). Their superplume could be stagnant and could spread laterally immediately beneath the 660, and the laterally spreading superplume could generate secondary narrow plumes because of thermal instability. The difference between our hypothetical model and that of Courtillot *et al.* (2003) is the vertical extent of the superplume (1000 km versus 660 km). We prefer the top of the superplume to be at a depth of 1000 km on the basis of the regional seismic image (Fig. 8). In our context, the superplume does not interact with the post-spinel phase boundary at the 660.

Davaille (1999) showed that the superplume (or dome in their study) oscillates vertically in their laboratory experiment when the density contrast (of chemical origin) between the superplume and ambient mantle is marginally balanced by the thermal density difference. If this is the case for the South Pacific superplume, the superplume imaged seismologically is a transient feature and could move vertically, which may cause a pulse of volcanic activity in the South Pacific. Cadio *et al.* (2011) suggested that the superplume is ascending (descending) beneath the southern (northern) half of the superswell where the geoid is low (high). If the vertical oscillation actually occurs, the superplume can rise across the 660 to the upper mantle. It was previously assumed that this could not occur, but recent high-pressure and high-temperature experiments have shown that phase transformation at the 660 may not be a barrier in a high-temperature environment because the 660 should represent the majorite–perovskite phase transition with a positive Clapeyron slope at temperatures higher than 1800°C (Hirose, 2002). At present, the South Pacific hotspots display relatively little volcanic activity possibly because the superplume is located in the lower mantle. The vertical oscillation may have caused an elevation of the superplume in the past, which may be responsible for the production of large oceanic plateaus in the Cretaceous period (Larson, 1991; Utsunomiya *et al.*, 2007).

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