Mantle plume or slab window?: Physical and geochemical constraints on the origin of the Caribbean oceanic plateau

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ABSTRACT

The Caribbean oceanic plateau formed in the Pacific realm when it erupted onto the Farallon plate from the Galapagos hotspot at ~90 Ma. The plateau was subsequently transported to the northeast and collided with the Great Arc of the Caribbean thus initiating subduction polarity reversal and the consequent tectonic emplacement of the Caribbean plate between the North and South American continents. The plateau represents a large outpouring of mafic volcanism, which has been interpreted as having formed by melting of a hot mantle plume. Conversely, some have suggested that a slab window could be involved in forming the plateau. However, the source regions of oceanic plateaus are distinct from N-MORB (the likely source composition for slab window mafic rocks). Furthermore, melt modelling using primitive (high MgO) calculations for the source of oceanic plateaus (mantle plumes) and slab windows (ambient upper mantle) cannot produce enough melt to form a plateau. The formation of the Caribbean oceanic plateau by melting of ambient upper mantle in a slab window setting, is therefore, highly improbable.

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1. Introduction

Knowledge of the tectonomagmatic history of the Caribbean plate (Fig. 1) is important in helping us to understand the palaeogeography of the inter-American region from the Jurassic. Of particular importance is the development, and subsequent closure, of the palaeogateway between North and South America and the associated impacts on the global climate (e.g. Droxler et al., 1998; Schneider and Schmittner, 2006).

Most of the Caribbean plate consists of a 8–20 km thick Late Cretaceous oceanic plateau (∼6 × 10^5 km^2) that formed in the Pacific (e.g. Edgar et al., 1971; Mauffret and Leroy, 1997; Kerr et al., 2003) and is possibly derived from the initial plume head phase of the Galapagos hotspot (e.g. Hoernle et al., 2002; Geldmacher et al., 2003; Thompson et al., 2003). Although somewhat controversial (e.g. Pindell et al., 2006), many consider the Caribbean oceanic plateau to have erupted onto the Farallon plate and subsequently transported to the northeast to collide with a large intra-oceanic arc [the Great Arc of the Caribbean, Burke (1988)] in the late Cretaceous (Fig. 2a) (e.g. Duncan and Hargraves, 1984; Burke, 1988; White et al., 1999; Thompson et al., 2003; Kerr et al., 2003; Mann et al., 2007). The Great Arc was located at the western side of the oceanic gap (the

Fig. 1. (a) Map of the Caribbean and Central American region showing the location of the obducted portions of the Caribbean oceanic plateau. Guatemala (Gma), El Salvador (ES), Costa Rica (CR), Panama (Pna), Swan Islands Transform Fault Zone (SITFZ), Oriente Transform Fault Zone (OTFZ), Plantain Garden-Enriquillo Fault Zone (PG-EFZ), Nicoya Complex (NC), Herradura Complex (HC), Quepos Complex (QC), Osa Complex (OC), Sona Complex (SC), Azuero Complex (AC), Serrania de Baudo Complex (SdBC) (from Hastie et al., 2008) and (b) General geology map of Curaçao, Dutch Antilles showing the location of the Curaçao Lava Formation (modified from Kerr et al., 1996c). Locality numbers refer to locations in Table 1.
Proto-Caribbean) which had been opening between the North and South American continents since the Jurassic (Fig. 2a).

In this model, the Caribbean plateau would have been too thick, hot and buoyant to subduct beneath the American or Proto-Caribbean subduction zones (e.g. Saunders et al., 1996). Accordingly, the southern portions of the plateau would have been subducted onto the continental margin of South America forming extensive accreted sequences in Colombia and Ecuador (Fig. 2a) (Kerr et al., 1996b, 2002a,b). In contrast, when the northern portion of the plateau collided with the Great Arc of the Caribbean it clogged the subduction zone and initiated subduction polarity reversal and subduction back-step such that the Proto-Caribbean crust began subducting in a westerly direction beneath the oceanic plateau (Duncan and Hargraves, 1984; Burke, 1988; Kerr et al., 2003). Over the last ~80 Ma the plateau and the segmented Great Arc of the Caribbean were tectonically emplaced between the westward moving North and South American continents to form the Caribbean plate (Duncan and Hargraves, 1984; Burke, 1988; Sinton et al., 1997; Hauff et al., 2000a,b; Kerr et al., 2003; Mann et al., 2007).

Conversely, others have proposed a significantly earlier Aptian/Albian (125–99.6 Ma) subduction polarity reversal (Fig. 2b) (e.g. Lebron and Perfit, 1993, 1994; Kesler et al., 2005; Pindell et al., 2005; Escuder Viruete et al., 2007; Marchesi et al., 2007). This has led Pindell et al. (2006) to challenge the mantle plume model by proposing that the Caribbean oceanic plateau may have been largely formed by magmatism associated with a slab window. Pindell and Kennan (2001) and Pindell et al. (2006) argue from plate reconstructions that a slab window existed in the Caribbean region from the Aptian/Albian to the early Campanian (83.5–70.6 Ma) when an active Proto-Caribbean seafloor spreading centre subducted beneath the Great Arc of the Caribbean (Fig. 2b). Pindell et al. (2006) proposed that partial asthenospheric melting facilitated by the slab window would form the thickened crust of the Caribbean oceanic plateau.

The aims of this paper are to compare oceanic plateau rocks to volcanoanl associated with slab windows and to determine if an oceanic plateau can be formed by magmatism resulting from a slab window. Furthermore, primitive lavas from the island of Curacao, Dutch Antilles (Fig. 1) will be assessed using the PRIMELT2 software of Herzberg and Asimow (2008) to determine the composition, degree of partial melting, and the potential temperature of their primary magmas in order to resolve the likely affinity of the mantle source region of the Caribbean oceanic plateau. This information will used to place constraints on the tectonomagmatic evolution of the Caribbean plate in the Late Cretaceous.

2. Geological background

2.1. Mantle plumes and the formation of LIPs

In the oceanic environment large igneous provinces (LIPs) are represented by oceanic plateaus, oceanic basins and aseismic ridges (Coffin and Eldholm, 1994). It has been demonstrated both geochemically (e.g. Kempton et al., 2000; Herzberg and O'Hara, 2002; Thompson et al., 2003; Fitton and Godard, 2004) and by physical computational modelling (Richards et al., 1989; Campbell and Griffiths, 1990; Farnetani and Richards, 1995; Farnetani et al., 2002; Farnetani and Samuel, 2005; Campbell, 2007) that voluminous oceanic plateaus can be formed from the partial adiabatic decompression melting of a hot, ascending, deep-mantle-derived, compositionally heterogeneous mantle plume head as it collides with the base of the lithosphere.

In characterising the thermal properties of a given mantle source region McKenzie and Bickle (1988) defined the term mantle potential temperature ($T_p$), which represents the temperature of a mass of convecting mantle on the Earth’s surface if it was to ascend along an adiabat (i.e. it neither loses or gains heat to its surroundings during ascent) and did not melt. Calculation of the $T_p$ enables us to determine if a primary magma is derived from a mantle source region with excess heat relative to ambient upper mantle (e.g. Bown and White, 1995).

Decompression partial melting, combined with higher calculated $T_p$ values within a mantle plume head ($\Delta T = 100–400 \, ^\circ C$) relative to the ambient upper mantle ($T_p \sim 1280–1475 \, ^\circ C$) (e.g. McKenzie and Bickle, 1988; Kinzler and Grove, 1992; Herzberg and O’Hara, 2002; Putirka, 2005; Courtier et al., 2007; Herzberg et al., 2007; Putirka et al., 2007), will rapidly produce large amounts of partial melt [15–30% e.g. Kerr et al. (2002b); Chazey and Neal (2004), Fitton and Godard (2004); Herzberg (2004)] that will eventually form a LIP, such as an oceanic plateau.

2.2. Slab windows

A slab window is formed at a ridge–trench–trench triple boundary when an active ocean spreading centre subducts beneath an overriding plate (Dickinson and Snyder, 1979). As the ridge subducts the trailing edges of either one or both of the plates will continue to diverge to form an ever-widening, slabless gap between the two subducting plates.
Q5

The third type of magmatism consists of enriched basalts (e.g. Aguillón-Robles et al., 2001; Calmus et al., 2003; Bellon et al., 2006) and (3) highly complex decompression partial melting and mixing of sub-slab and/or supra-slab asthenosphere to form alkaline and tholeiitic mafic volcanism (e.g. Benoit et al., 2002; Goring et al., 2003; Bellon et al., 2006; Pallares et al., 2007). These upwelling asthenospheric melts can subsequently interact with overlying enriched lithospheric and crustal sources resulting in anatectic mixing to form more enriched mafic magmas (e.g. Cole and Basu, 1992, 1995; Gorring et al., 1997, 2003) and depleted alkaline and tholeiitic ocean island basalt (OIB)-type (Cole and Basu, 1992, 1995; Benoit et al., 2002; Bellon et al., 2006) mantle components and lavas. Additionally, “normal” arc-derived volcanism can sporadically occur in the slab-free region (e.g. Goring et al., 1997; Pallares et al., 2007) suggesting that asthenospheric mantle which has been metasomatised by slab-related fluids exists, and can be melted, in a slab window environment.

In contrast, oceanic plateaus such as the Ontong Java (OJP) and the Caribbean are composed of voluminous, predominantly tholeiitic basalts with minor picritic and komatiitic successions (e.g. Kerr et al., 1996a,b,c; Neal et al., 1997; Arndt et al., 1997, 1998; Révillon et al., 1999; Hauff et al., 2000a,b; Tejada et al., 2002; Fitton and Godard, 2004; Hastie et al., 2008). Oceanic plateau rocks are mostly tholeiitic with predominantly flat, primitive mantle-normalised multi-element patterns (Fig. 3) and these characteristics contrast markedly with the enriched mafic alkaline, and slab-melt related rocks formed in slab window environments (Fig. 3).

Furthermore, oceanic plateau rocks lack the negative Nb and Ta anomalies that are present in slab window related adakites and high-magnesian andesites (e.g. Pallares et al., 2007) as well as other geochemical characteristics of these igneous rocks e.g. SiO₂ > 56%, Al₂O₃ > 15%, MgO generally < 3%, low Y (∼18 ppm) and HREE (Yb < 1.9 ppm) for an adakite or Sr up to 3000 ppm, Ba > 1000 ppm and high Na/K ratios for magnesian andesites (e.g. Saunders et al., 1987; Defant et al., 1992; Yogodzinski et al., 1995). Trace element and radiogenic isotope compositions (Figs. 3 and 4) also reveal that oceanic plateau basalts are derived from mantle source regions distinct from depleted N-MORB-source upper mantle, with predominantly higher LREE/HREE ratios and lower (more enriched) εNd(t) values than present-day N-MORB (e.g. Kerr et al., 1996a; Kempton et al., 2000; Tejada et al., 2004; Hastie et al., 2007, 2008).

Consequently, the mafic alkaline and tholeiitic N-MORB and OIB-type lavas and slab-related adakitic volcanism formed in slab window environments are compositionally very different to the mantle plume-derived tholeiitic basalts, picrites and komatiites of oceanic plateaus.

3.2. Volume of volcanic products from a slab window and oceanic plateau

Oceanic plateaus are voluminous LIPs formed by massive outpourings of mafic extrusive lavas on “normal” ocean crust and are often associated with extensive intrusive activity (e.g. Coffin and
4. Physical and geochemical influences on mantle melting during oceanic plateau formation

When comparing the partial melting of a mantle plume and a slab window source region a number of basic questions need to be considered: (a) is the source hydrous or anhydrous? (b) are the melts derived from mantle source regions with similar Tp values? (c) are the melts derived from similar depths? (d) are the melts formed by comparable degrees of partial melting? and (e) are the mantle source regions compositionally similar?

4.1. Anhydrous nature of plateau lavas

Kerr and Mahoney (2007) have shown that, apart from two localities in the Caribbean (gabbros and pegmatites in Bolivar, Colombia and one komatiite flow on Gorgona), the source regions responsible for forming the Caribbean oceanic plateau magmas, are anhydrous mantle peridotite. Thus, if the oceanic plateau was derived from decompression melting in a slab window, a hydrated, arc-derived supra-slab mantle source region cannot be considered as a viable source region. This point is further demonstrated by the fact that the lavas of the Caribbean oceanic plateau (even those with a hydrated source) lack any subduction-related geochemical signal (e.g. La/Nb > 1), which also argue against the involvement of a supra-slab source region (Fig. 3) (Kerr et al., 2004). Therefore, any proposed slab window source region for Caribbean oceanic plateau has to be completely devoid of any geochemical signature of mantle that has been compositionally modified by subduction-related processes.

4.2. Previous Tp calculations for the source of oceanic plateaus (mantle plumes) and slab windows (ambient upper mantle)

Herzberg (2004) calculates that the primary magmas of the OJP had a Tp of 1500–1560 °C. Based on Caribbean oceanic plateau lavas from Gorgona Island (located ~60 km off the Colombian Pacific coast (Fig. 1a), Herzberg and O’Hara (2002), and Herzberg et al. (2007) calculate that the Tp of high-MgO komatiites is 1520–1570 °C while associated picrites have higher Tp values of 1600–1700 °C.

In contrast, the most recent studies of Herzberg et al. (2007) and Putirka et al. (2007) based on phase equilibria and olivine thermometry respectively, calculate the average Tp of upper mantle sources beneath mid-ocean ridge systems to be ~1300–1454 °C, similar to previous studies (e.g. ~1280–1475 °C McKenzie and Bickle, 1988; Kinzler and Grove, 1992; Asimov et al., 2001; Herzberg and O’Hara, 2002; Wang et al., 2002; Putirka et al., 2005). It should be noted, that although this review uses the full range of calculated ambient mantle temperatures, many recent studies favour a Tp range of 1300–1400 °C for the upper mantle (e.g. McKenzie et al., 2005; Courtier et al., 2007; Herzberg et al., 2007). Nevertheless, although the Tp of the upper mantle is variable, the highest estimated values are still not as hot as the estimated Tp of the Caribbean and OJP oceanic plateau source regions. Consequently, mantle plume source regions are clearly associated with excess Tp relative to the mantle sources of N-MORB.

4.3. Depth of melting

Herzberg and O’Hara (2002) calculated using forward and inverse phase equilibria modelling that the Gorgona komatiites and picrites were derived from mantle sources at depths of 115–140 km and 240 km respectively. Herzberg (2004) also concluded that similar melts in the OJP were derived from a mantle source at a depth of 108–132 km. These high pressures (~2.5 GPa) indicate that the primary magmas are derived from a garnet peridotite mantle source region. This further explains the high-MgO contents of the mantle plume primary magmas because not only do high degrees of partial melting produce high-MgO melts, but melts in equilibrium with garnet peridotite (high pressures) become similar to those of the upper mantle (e.g. McKenzie et al., 2005; Courtier et al., 2007, Herzberg et al., 2007). Nevertheless, although the Tp of the upper mantle is variable, the highest estimated values are still not as hot as the estimated Tp of the Caribbean and OJP oceanic plateau source regions. Consequently, mantle plume source regions are clearly associated with excess Tp relative to the mantle sources of N-MORB.

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Consequently, in contrast to mantle plumes, slab window, anhydrous sub-lab melt, with lower mantle \( T_p \) values, would have to rise to shallower levels in the upper mantle in order to intersect the mantle dry solidus and partially melt. Hotter mantle plumes, on the other hand, would intersect the mantle dry solidus at deeper levels in the upper mantle because of their excess \( T_p \) thus promoting deeper and more extensive partial melting (Fig. 5). This is supported by Aguillon-Robles et al., 2001; Van Wijk et al., 2001; McCrory et al., in press who demonstrated that melts formed in slab window environments are derived from passive upwelling asthenosphere that undergoes decompression melting below \(< 100 \text{ km depth.}

4.4. Oceanic plateau primary magma compositions and degrees of partial melting

4.4.1. Previous work

Modelling demonstrates that the primary magmas of the OJP, are derived from 25 to 30% partial melting of a peridotite source region (Chazey and Neal, 2004; Fitton and Godard, 2004). Previous work on the primary magmas of the Caribbean oceanic plateau indicates that they formed by \(\sim 20–30\% \) partial melting of a depleted and/or enriched spinel-peridotite source (e.g. Kerr et al., 1996a,b, 2002a,b; Hauff et al., 1997; Révillon et al., 2000).

Fitton and Godard (2004) and Tejada et al. (2004), have shown that the trace element compositions and relatively depleted Sr–Nd–Pb–HF radiogenic isotope ratios of primary magmas from the OJP can be attributed to \( \sim 30\% \) batch partial melting of a primitive mantle source (bulk silicate Earth composition) which has undergone 1% continental crust extraction 3–4 Ga. The composition of this theoretical enriched peridotite source region is similar to the fertile Kettle River peridotite source (KR-4003) used in the forward phase equilibria modelling of Herzberg (2004) which determined the major element composition, \( T_p \) and degrees of partial melting of the OJP primary magmas.

Fig. 4 shows that the primary magmas from the OJP have \( e_{\text{Nd}}(t) \) and \( e_{\text{Hf}}(t) \) ratios that fall within the Caribbean oceanic plateau field, close to lavas from the island of Curaçao (Fig. 1a and b). This shows that at least some of the Caribbean oceanic plateau source is isotopically similar to the OJP source. Consequently, although the mantle source region for the Caribbean oceanic plateau is more heterogeneous than the source for the OJP lavas (Hastie et al., 2008), the theoretical mantle source from Fitton and Godard (2004) and Tejada et al. (2004) represents a possible mantle composition from which the Caribbean oceanic plateau lavas could be derived.

4.4.2. Calculated primary magma compositions of primitive lavas from Curaçao, Dutch Antilles

Although primary magma compositions have been determined from primitive lavas on Gorgona and the Galapagos, to date no primary magma compositions have been calculated from the main Caribbean oceanic plateau. The island of Curaçao in the southern Caribbean Sea (Fig. 1a and b) is predominantly composed of the \( \sim 5 \text{ km thick Curaçao Lava Formation (CLF)} \) with picrites at the base that evolve into tholeiitic pillow basalts and dolerites nearer the top (Fig. 1b) (Kerr et al., 1996a; Révillon et al., 1999). The presence of the primary lavas in the CLF can be used to study the composition of the primary melts involved in the formation of the Caribbean oceanic plateau.

PRIMELT2 software (Herzberg and Asimow, 2008) can determine the major element composition of primary melts from mass balance solutions derived from a parameterization of peridotite partial melt compositions in Herzberg and O’Hara (2002). The software is sensitive to mantle source heterogeneities and clinopyroxene/plagioclase fractionation and accumulation. Consequently, only primary lavas that have solely fractionated olivine can be used in the calculations, which resolve the primary melt composition by determining all the potential primary magma compositions by successive 1% additions and subtractions of equilibrium olivine (e.g. Herzberg and Asimow, 2008). The composition of the primary melt is determined by identifying the composition of the potential primary melt that has the same melt fraction in both FeO–MgO and projection space [see Herzberg et al. (2007) and Herzberg and Asimow (2008) for further details].

Many of the lavas analysed for major elements from the CLF by Klaver (1987) and Kerr et al. (1996a) are relatively evolved and have fractionated clinopyroxene and/or plagioclase in addition to olivine. However, four picrites and basalts reported by Klaver (1987) have only fractionated olivine and so are suitable for modelling using PRIMELT2. The results of the primary magma calculations are shown in Table 1. All four samples are derived from a heterogeneous mantle source composed of garnet peridotite and hib 습자. The interpretation of the remaining results depends on whether the Curaçao lavas were formed by non-modal batch partial melting or accumulated fractional partial melting. At similar degrees of partial melting the two different melting regimes produce melts, and leave mantle residues, with differing compositions that require assessed fertile mantle sources with different \( T_p \) values (Herzberg and O’Hara, 2002; Herzberg et al., 2007; Herzberg and Asimow, 2008). This is seen in Table 1 where, for example, sample 79Be073 can be formed by fractionation of a primary magma formed by (a) 31.7% batch partial melting of a fertile mantle source region with a \( T_p \) of 1585 °C or (b) 29.8% accumulated fractional partial melting of a fertile mantle source region with a \( T_p \) of 1536 °C.

Fitton and Godard (2004) suggest that because of large degrees of partial melting (~30%) in the source region for the OJP lavas it is likely that batch partial melting may become the dominant process. Therefore, if the same is assumed for the Caribbean plateau samples, the batch melting results show that the Curaçao primary magmas were formed by 30–32% partial melting (similar to the OJP primary magmas) of a fertile mantle peridotite source region with a \( T_p \) of 1564–1614 °C. The primary melts contained 19.6–22.2 wt.% MgO that were in equilibriump with olivines with Fo22.5–33.3.

A 30% batch partial melt of the theoretical mantle plume source region of Fitton and Godard (2004) produces similar trace element contents to the OJP primary magmas (Fig. 6). Trace element compositions...
Table 1
Major element composition of the Curaçao lavas and their respective primary magma compositions based on batch and accumulated fractional partial melting using PRIMELT2 software (Herzberg and Asimow, 2008). Major element XRF analyses are from Klaver (1987) and have an accuracy and precision of ±1%.

<table>
<thead>
<tr>
<th>Sample number</th>
<th>79Be073</th>
<th>79Kv019</th>
<th>79Kv615</th>
<th>79Kv619</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Rock type</strong></td>
<td>Pillow basalt</td>
<td>Pillow basalt</td>
<td>Pillow basalt</td>
<td>Pillow basalt</td>
</tr>
<tr>
<td><strong>Rock series</strong></td>
<td>Tholeiite</td>
<td>Tholeiite</td>
<td>Tholeiite</td>
<td>Tholeiite</td>
</tr>
<tr>
<td><strong>Source type</strong></td>
<td>Fertile mantle</td>
<td>Fertile mantle</td>
<td>Fertile mantle</td>
<td>Fertile mantle</td>
</tr>
<tr>
<td><strong>FeO/Fe2O3</strong></td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
</tr>
</tbody>
</table>

Curaçao primary magmas are very similar to (although they are slightly more enriched in the most incompatible elements than) the OJP primary magmas and can be modelled by a 30% batch partial melt of the theoretical mantle plume source region.

4.4.3. Calculated primary magmas from a slab window
Most slab window related mafic lavas are derived from sub-slab mantle sources that have depleted MORB mantle (DMM)-like compositions (e.g. Cole and Basu, 1992, 1995; Bellon et al., 2006; Pallares et al., 2007) and any OIB-like enrichment of the sub-slab derived lava is a result of (1) a fortuitously located mantle plume and/or (2) contamination of the sub-slab melts with continental lithospheric mantle/crust.

It can be seen in Fig. 6 that 30% batch partial melting of a sub-slab source region with an average DMM composition (Workman and Hart, 2005) cannot form a primary magma with the incompatible trace element systematics of the calculated OJP or Curaçao primary magmas. A 30% partial melt of DMM is more depleted in incompatible elements relative to the primary magmas of oceanic plateaus. Furthermore, the radiogenic isotope ratios of oceanic plateaus are distinct from N-MORB (Fig. 4) (e.g. Hastie et al., 2008).

5. Discussion

5.1. Melt generation in mantle plumes: a synthesis

As noted in Section 4.2, estimates of upper mantle \( T_p \) range from 1280 °C to 1475 °C and using Fig. 7a (modified from McKenzie and Bickle, 1988) it can be seen that these temperatures can generate melt thicknesses ranging from ~7 to 25 km (assuming that all of the melt is extracted and decompression melting ceases at the surface of the crust).

Although a melt thickness equivalent to the thickness of oceanic plateau crust can be generated at a \( T_p \sim 1475 \degree C \) mid-ocean ridge basaltic sequences only attain crustal thicknesses of ~7 km (Fig. 7a) (White et al., 1992) and have mantle sources which are thermally and compositionally distinct from those of oceanic plateaus (Section 3.1). Also, it has already been discussed in Section 4.2 that ambient mantle temperatures are now considered to range from 1300 to 1400 °C and are not as high as 1475 °C.

If the Galapagos plume head ascended beneath thin oceanic lithosphere, or a spreading centre, the minimum lithospheric thickness would be represented by “normal” ~7 km thick oceanic crust up to the 20 km thick estimate for the Caribbean plateau crust (e.g. Mauffret and Leroy, 1997). These thicknesses require $T_0$ of $>1400 °C$ and $\geq 1500 °C$ respectively to produce the 30–32% partial melting needed to form the Caribbean primary magmas (Fig. 7b). Additionally, if the Curaçao primary magmas formed by accumulated fractional partial melting the PRIMELT2 software calculates that they are derived from 1 to 2 GPa (40–60 km), whereas if they were formed by batch melting they are derived from $\sim 3$ GPa (85–95 km), which requires $T_0$ values of $>1500 °C$ and $>1600 °C$ respectively (Fig. 7b). Therefore, the Galapagos plume at 90 Ma is likely to have been located beneath a thicker lithospheric lid ($\sim 20$ km), and thus, the generation of Curaçao primary magmas by 30% partial melting would require much higher $T_0$ values ($>1500 °C$) to produce the Caribbean oceanic plateau (Fig. 7b). This supports the calculated $T_0$ values of 1564–1614 °C for the Curaçao primary magmas in Section 4.4.2.

Therefore, in contrast to ambient upper mantle, a mantle plume source region must have a high enough $T_0$ to produce large degree partial decomposition melting below thinned and thickened oceanic lithosphere without the need for lithospheric extension. Fig. 5 shows that mantle material with a $T_0$ of $>1500 °C$ (similar to that which melted to form the Caribbean and OJP lavas) will intersect the dry solidus at $>100$ km depth. Conversely, upper mantle material at “normal” $T_0$ values ($\sim 1300–1475 °C$; Kizner and Grove (1992) and Herzberg et al. (2008)) will only commence decomposition melting at depths less than 100 km (Section 4.2). Thus, considering that the average thickness of the oceanic lithosphere is 100 km (e.g. Conrad and Lithgow-Bertelloni, 2006) the majority of upper mantle source regions require lithospheric extension to allow decomposition melting to occur.

5.2. Melt generation in a slab window environments: a synthesis

To model the effect of a lithospheric lid, McKenzie and Bickle (1988) describe the $\beta$ value (stretching factor), which is the ratio of the surface area before and after an extensional event has occurred in oceanic or continental lithosphere (Fig. 8). The $\beta$ value can control the depth and volume of partial melting beneath a “lithospheric lid.” The volume of melt is therefore dependant on the thickness of the lithospheric lid, the degree of lithospheric thinning, the duration of extension and on the $T_0$ of the upper mantle source region (Figs. 5, 7 and 8) (Bown and White, 1995).

Fig. 7. (a) Depth vs. melt thickness diagram modified from McKenzie and Bickle (1988). Temperatures are mantle $T_0$ values and thickness of Caribbean and OJP oceanic crust are illustrated (e.g. Gladczenko et al., 1997; Mauffret and Leroy, 1997). (b) Partial melting vs. depth diagram modified from Fitton and Godard (2004). Temperatures are $T_0$ values. Thicknesses of “normal” oceanic lithosphere and crust (e.g. White et al., 1992; Conrad and Lithgow-Bertelloni, 2006), the OJP and Caribbean oceanic plateau crust (e.g. Gladczenko et al., 1997; Mauffret and Leroy, 1997) are shown as is the batch partial melting field for the Curaçao and OJP primary magmas.

Fitton and Godard (2004) recognise that plume magmas would not be generated by adiabatic decompression partial melting up to the surface of the crust and that in the case of the OJP a lithospheric lid would inhibit decomposition partial melting. Fitton and Godard (2004) suggest that the primary magmas of the OJP would have been generated beneath at least a ~30 km thick crust, and this value is likely to represent a minimum thickness for the lithospheric lid during the formation of the plateau. Given that the OJP primary magmas are derived by ~30% partial melting (Fitton and Godard, 2004; Herzberg, 2004) and a mantle plume head would have underlain at least a ~30 km thick OJP crust (Gladczenko et al., 1997) when the primary magmas were generated, the $T_0$ of the OJP primary magmas must have been ~1500 °C (Fig. 7b). This supports the calculated 1500–1560 °C $T_0$ of OJP primary magmas based on phase equilibria (Herzberg, 2004).

If the Caribbean oceanic plateau represents the plume head phase of the Galapagos plume did it form above “normal,” thickened or thinned oceanic lithosphere? Pinell et al. (2005) have suggested that the hotspot was below a spreading centre at ~90 Ma. Conversely, it has been proposed that any Galapagos plume head at ~90 Ma would have underlain Farallon oceanic lithosphere well away from any spreading ridge (e.g. Duncan and Hargraves, 1984; Hoernle et al., 2002; Mann et al., 2007; Liu et al., 2008).

Fig. 8. Modified melt thickness vs. $\beta$ value diagram from McKenzie and Bickle (1988). Temperatures are $T_0$ values and curves for each temperature are for initial lithospheric thicknesses of 70 and 100 km.
As noted in Section 2.2, the lithosphere above a slab window represents a passive extensional regime (Thorkelson, 1996). The extensional Basin and Range province in the western U.S.A and Mexico is a tectonically and magmatically complex area (e.g. Wang et al., 2002) and is located above several slab windows (e.g. Dickinson, 1997; Pallares et al., 2002; McCrory et al., in press). The extension has been linked to the development of the slab windows (e.g. Dickinson, 1997); consequently, the area can be used as an analogue for studying the amount of extension in a slab-free region.

It has been calculated that the Basin and Range province has a β elongation stretching factor of 2–3 (Gans, 1987; McKenzie and Bickle, 1988), which indicates that even with a relatively thin initial lithosphere thickness of 70 km, and the highest estimated upper mantle Tp of ~1475 °C (Kinzler and Grove, 1992), decompression melting due to extension can only form melt thicknesses from ~11.5 km (β = 2) to ~17.5 km (β = 3) (Fig. 8). Moreover, if 1400 °C is considered as the more likely higher Tp estimate for the upper mantle (e.g. Courtier et al., 2007) decompression melting due to extension can only form melt thicknesses ~3.75 km (β = 2) to ~7.5 km (β = 3) (Fig. 8). These values are less than the ~20–30 km melt thicknesses of the Caribbean and Ontong Java oceanic plateaus (e.g. Gladczenko et al., 1997; Mauflret and Leroy, 1997).

The models of McKenzie and Bickle (1988) assume an instantaneous rifting event; however, in reality lithospheric extension is more likely to occur over a protracted amount of time (e.g. Bown and White, 1995). The effect of protracted rifting means that as the upper part of the asthenosphere upwells it will cool due to conductive heat loss to the lithosphere. Therefore, finite (protracted) extension is even less likely to produce the melt thicknesses observed in oceanic plateaus from a slab window, than instantaneous extension.

The main magmatic event which formed the Caribbean oceanic plateau only lasted ~3–4 m.y. (e.g. Kerr et al., 2003). Bown and White (1995) have calculated melt thickness as a function of rift duration, β factor, upper mantle Tp and initial lithosphere thickness. Using their models and assuming an upper mantle Tp of 1400 °C along with a thick initial lithosphere representing Farallon or Great Arc crust and a rifting event lasting 3–4 m.y. a β = 2 produces little or no melt, while β = 3 only forms a melt thickness of ~3 km. These values represent maximum melt thicknesses for an initial lithospheric thickness comparable to oceanic/arc lithosphere as heat loss is assumed to be in a vertical direction only. If horizontal heat loss into adjacent thickened lithosphere is also taken into account, the melt thicknesses are reduced still further (see Fig. 10 in Bown and White, 1995).

5.3. Could a mantle plume pass through a slab window?

The estimated large size 1000–1200 km of a mantle plume head (Campbell, 2007) means that if one fortuitously ascended below a slab window, the plume would dominate the thermal regime resulting in decompression melting to form a LIP. However, the plume would also be likely to cause the overlying lithosphere to extend (e.g. Campbell, 2007) and would conductively transfer heat into the hydrous supra-slab asthenosphere promoting partial melting.

Accordingly, such a LIP would contain significant successions of geochemically distinctly island arc lavas, or oceanic plateau lavas with a subduction-related signature, caused by thermally induced partial melting in the supra-slab asthenosphere. However, to date no such lavas have been found interlayered within the Caribbean oceanic plateau (e.g. Kerr et al., 2003). It must therefore be concluded, from an igneous perspective, that a mantle plume could not have been associated with a slab window when the Caribbean oceanic plateau formed.

6. Conclusions

The main Caribbean oceanic plateau represents a large outpouring of mafic volcanism 93–89 Ma. Calculated Curacao primary magmas contained 19.6–22.2 wt.% MgO and were derived from 30 to 32% batch partial melting of a fertile peridotite source region which had a Tp of 1564–1614 °C. Conversely, the primary magmas can also be derived from 29 to 30% accumulated fractional partial melting of a fertile peridotite source with a Tp of 1517–1564 °C.

This, combined with the thicker than normal oceanic crust in the Caribbean region (up to 20 km) demonstrate that the Caribbean oceanic plateau lavas are derived from the decompression melting of a hot, heterogeneous, upwelling mantle plume consisting of relatively fertile peridotite ~90 Ma. The arguments against the Caribbean oceanic plateau lavas being derived from a slab window are:

1. Caribbean oceanic plateau igneous rocks are predominantly composed of large volumes of tholeiitic basalts with minor sequences of picrites and komatitites. In contrast, slab window magmatism is dominated by relatively small volumes of very diverse, but compositionally unique, volcanic rocks such as intraplate tholeiitic and alkaline mafic rocks, “normal” island arc volcanism and slab-melt related adakites, magnesian andesites (bajaites) and Nb-enriched basalts.

2. Major and trace element modelling and radiogenic isotope systematics clearly demonstrate that the mantle sources of lavas in oceanic plateaus are geochemically distinct from N-MORB-type (the likely source composition for slab window mafic rocks) and OIB-type mantle source regions.

3. The Tp of the Curacao and other Caribbean oceanic plateau lavas are higher than any estimated Tp from a mid-ocean ridge, which imply a mantle plume origin.

4. Extensional decompression partial melting of sub-slab asthenosphere with an ambient mantle Tp beneath oceanic lithosphere (Farallon plate/Great Arc) of normal thickness e.g. 70–100 km cannot produce enough melt to form a LIP.

Therefore, a slab window cannot be responsible for forming the Caribbean oceanic plateau even in conjunction with a mantle plume because of the lack of island arc volcanism within the Caribbean oceanic plateau successions. Accordingly, given the constraints presented in this paper, we continue to advocate that the thickened crust of the Caribbean oceanic plateau was formed 89–93 Ma by melting of a hot mantle plume in the Pacific realm, and that subsequent collision of this plateau with the Great Arc of the Caribbean in the late Cretaceous caused subduction polarity reversal. This late Cretaceous polarity reversal enabled the Caribbean plate (plateau) to be tectonically emplaced between the American continents.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.earscirev.2009.11.001.


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