IN-SITU U-PB SECONDARY ION MASS SPECTROMETRY (IN-SIMS)

GEOCHRONOLOGY FROM THE LEEWARD ANTILLES ISLANDS OF ARUBA, CURAÇAO, BONAIRE, AND GRAN ROQUE: IMPLICATIONS FOR THE TEMPORAL EVOLUTION OF THE CARIBBEAN LARGE IGNEOUS PROVINCE (CLIP) AND EARLY ARC MAGMATISM

by

CHRISTOPHER G. HUMPHREY

(Under the Direction of James E. Wright)

ABSTRACT

In-situ U-Pb secondary ion mass spectrometry (IN-SIMS) was performed on both microbaddeleyite and micro-zircon from mafic samples of the Leeward Antilles islands of Aruba, Curacao, Bonaire, and Gran Roque in order to gain a better understanding of the temporal evolution of both the Caribbean large igneous province (CLIP) and early arc magmatism. IN-SIMS analysis of CLIP samples from Aruba, Curacao, and Gran Roque indicate a ~30 m.y. span of CLIP magmatism from the Aptian to Coniacian (ca. 115-87 Ma). The extensive duration of CLIP magmatism argues against the formation of the CLIP within a few million years (e.g. Sinton et al., 1998; Kerr et al., 2003). IN-SIMS geochronological results from Bonaire in addition to U-Pb SHRIMP data from Wright and Wyld (2010) indicate that early arc activity occurred on Bonaire from the Albian to Cenomanian (ca. 112-95 Ma).

INDEX WORDS: Caribbean Large Igneous Province (CLIP), IN-SIMS geochronology, oceanic plateau, large igneous province (LIP), Leeward Antilles, Aruba, Curacao, Bonaire, Gran Roque
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CHAPTER 1

INTRODUCTION

Large igneous provinces (LIP) are enormous crustal emplacements of mafic extrusive and intrusive rocks, which originate by processes other than “normal” seafloor spreading (Coffin and Eldholm, 1994). Based upon both geochemical (Kempton et al., 2000; Herzberg and O’Hara, 2002; Thompson et al., 2003; Fitton and Godard, 2004; Hastie and Kerr, 2010) and physical and computational modeling (Richards et al., 1989; Campbell and Griffiths, 1990; Farnetani and Richards, 1995; Farnetani et al., 2002; Farnetani and Samuel, 2005; Campbell, 2007) LIPs have been proposed to form by decompression melting of a deep-seated mantle plume. These events occur on a global scale and are found throughout the Earth’s history from the Archean to present. Oceanic plateaus are a type of LIP that consist of anomalously thickened oceanic crust on the order of 35 km (e.g. Ontong Java Plateau; Gladczenko et al., 1997; Richardson et al., 2000). These massive outpourings of mafic magmas cover extended areas of the Earth’s crust in excess of $1 \times 10^5$ km$^2$ (Saunders et al., 1996). Oceanic plateaus found in the geologic record range in age from ~230 Ma to ~ 90 Ma (Fig. 1) (Greene et al., 2010 and references therein) and have been hypothesized to form over one or more pulses of magmatism that each occurred over a brief period of time (1-3 m.y.) (Duncan, 2002; Coffin et al. 2002; Kerr, 2003). These short intervals of plateau eruption can exceed the production of midocean ridge volcanism for the same period of time (Duncan and Richards, 1991; Larson, 1991). Oceanic plateau eruptions have been linked to multiple Cretaceous oceanic anoxic events, which occurred around the Cenomanian-Turonian boundary (93.5 Ma) and in the early Aptian (~120 Ma).
These oceanic anoxic events are characterized by widely distributed organic-rich sediments that are believed to form due to the reduction of dissolved O$_2$ in the oceans by hydrothermalism associated with oceanic plateau magmatism (Sinton and Duncan, 1997). However, the timing and duration of oceanic plateau magmatism has been a controversial topic (e.g. Ontong-Java Plateau and Wrangellia Plateau) due to the lack of precise geochronological data.

The Ontong-Java Plateau (OJP), located in the Western Pacific (Fig. 1), is the world’s largest oceanic plateau. The OJP covers an area of 2.0 \times 10^6 \text{ km}^2 with crustal thicknesses in excess of 35 km (Gladczenko et al., 1997; Richardson et al., 2000). Geochronological results from samples recovered from ocean drilling and uplifted sections of OJP on the adjacent Solomon Islands indicate a strongly bimodal distribution of $^{40}\text{Ar}/^{39}\text{Ar}$ ages at ~122 Ma and ~90 Ma (Mahoney et al., 1993; Tejada et al., 1996, 2002). Castillo et al. (2004) also recognized intermediate ages of ~111-115 Ma, which may suggest lower rates of emplacement for OJP magmatism than previously reported. However, the age and duration of OJP magmatism has not been established with any certainty because the basalts of the OJP have been difficult to date with the $^{40}\text{Ar}/^{39}\text{Ar}$ method, due to samples being altered and having very low potassium contents (Fitton et al., 2004). Chambers et al. (2002) even suggests the younger apparent ages (ca. 90 Ma) are the result of argon recoil and represent minimum ages.

The Wrangellia Plateau extends 2500 km within a narrow belt on the western margin of North America from British Columbia to Alaska (Fig. 1) with thicknesses of flood basalts of 6 km (Greene et al., 2010). Samples of Wrangellia flood basalts and plutonic rocks from British Columbia, Yukon, and Alaska have been dated using both U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ techniques to
yield an age range of ~125-130 Ma (Campbell, 1981; Mortensen and Holbert, 1991; Parrish and McNiccol, 1992; Lassiter, 1995; Bittenbender et al., 2003; Sluggett, 2003; Schmidt and Rodgers, 2007). However, recent $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological analyses on eight samples of the Wrangellia flood basalts by Greene et al. (2010) only resulted in one sample that retained a magmatic age (227.5 ± 1.2 Ma) that corresponds to previously reported ages of the Wrangellia Plateau. The remaining seven ages ranged from 191-73 Ma and are indicative of open-system behavior of the $^{40}\text{Ar}/^{39}\text{Ar}$ systematics due to metamorphism.

As seen through previous geochronological studies, difficulties arise when traditional methods are used to date mafic rocks of oceanic plateaus because these rocks are commonly altered and have very low potassium contents. The combination of these factors along with the possible disturbance of $^{40}\text{Ar}/^{39}\text{Ar}$ ratios due to metamorphism may lead to younger ages being reported than the actual emplacement age of the sample. To overcome the problems of the $^{40}\text{Ar}/^{39}\text{Ar}$ method, a recently developed in-situ U-Pb SIMS (IN-SIMS) technique by Schmitt et al. (2010) has been utilized to date micro-baddeleyite and micro-zircon from mafic samples. IN-SIMS bypasses the potential problems encountered when mafic rocks are dated with the $^{40}\text{Ar}/^{39}\text{Ar}$ method by the ability of IN-SIMS to provide precise U-Pb ages. These IN-SIMS ages can then be used to solve fundamental issues that concern oceanic plateaus, such as their age, duration, and environmental implications.

This is a reconnaissance study that tests the application of the IN-SIMS technique on mafic exposures of both oceanic plateau and magmatic arc rocks of the Leeward Antilles within the Caribbean plate. Geochronological age results from this study will provide improved resolution for the temporal evolution of the Caribbean large igneous province (CLIP) and related arcs. If mafic exposures of the Caribbean plate can be successfully dated with the IN-SIMS
technique, then this method can be implemented to precisely date other oceanic plateaus in order
to better understand their development within the world’s oceans.

Throughout the text, this study uses the time scale of Gradstein et al. (2004) for the ages of the Cretaceous and Paleogene stage boundaries.
Figure 1: A map that shows the distribution of Phanerozoic large igneous provinces (LIP). LIPs are indicated in yellow (continental) and orange (oceanic) with peak eruption ages. Red areas are flood basalts and ocean islands (modified from Greene et al., 2010 and references therein)
CHAPTER 2

CARIBBEAN LARGE IGNEOUS PROVINCE (CLIP)

Bathymetry and geophysical studies indicate that the interior of the Caribbean plate is composed of anomalously thick oceanic crust (Edgar et al., 1971; Diebold et al., 1981; Burke, 1988; Mauffret and Leroy, 1997; Driscoll and Diebold, 1998; Case et al., 1990) that is widely interpreted to be representative of an oceanic plateau (Fig. 1; Caribbean large igneous province; CLIP; also referred to by other workers as the Caribbean-Colombian oceanic plateau; CCOP). CLIP rocks found within the Caribbean, South America, and Central America have been shown lithologically, petrologically, and geochemically to indicate an oceanic plateau origin for the CLIP (Fig. 2; Donnelly, 1973; Donnelly et al., 1990; Kerr et al. 1996; Sinton et al., 1998; White et al., 1999; Revillon et al., 2000; Kerr et al., 2009). Geochronological studies of the CLIP (Fig. 3; Alvarado et al., 1997; Sinton and Duncan, 1997; Sinton et al., 1998; Lapierre et al., 1999; Walker et al., 1999; Revillon et al., 2000) suggest that CLIP magmatism occurred over a short period of time from ca. 92-88 Ma, with a minor pulse of volcanism between ca. 78-72 Ma (Fig. 3; Kerr et al., 1997; Sinton et al., 1998; Revillon et al., 2000). The ~90 Ma pulse of CLIP magmatism represents an enormous outpouring of magmatism as recorded extensively throughout the Caribbean plate. It has also been proposed as a major contributor to the ocean anoxic event that occurred at the Cenomanian-Turonian boundary (Sinton and Duncan, 1997; Kerr, 1988; Kerr et al., 2003; Snow et al., 2005).

In addition to oceanic plateau rocks, fragments of island arc-related sequences are found in the Greater Antilles, Aves Ridge, and Leeward Antilles along the complex accretionary
regions associated with the tectonic margins of the Caribbean plate (Pindell and Dewey, 1982; Pindell and Barrett, 1990; Pindell et al., 2005, 2006). These rocks are thought to be part of a Pacific-derived island arc that formed by subduction of oceanic crust beneath the “Great Arc of the Caribbean” (Beets et al., 1984; Duncan and Hargraves, 1984; Bouysse, 1988; Burke, 1988; White et al., 1999; Kerr et al., 2003; Thompson et al., 2004; Jolly et al., 2006; Pindell et al., 2006). Geochronological results on the island arc rocks of the “Great Arc of the Caribbean” range in age from late Early Cretaceous to Late Cretaceous (Maurasse, 1990; Stockhert et al., 1995; Stanek et al., 2000; Maresch et al., 2000; Pindell et al., 2005).

PREVIOUS STUDIES OF KNOWN OCCURRENCES OF THE CLIP

Previous studies of known occurrences of the CLIP indicate that the main characteristics of the CLIP include high-MgO lavas, chemically homogeneous basalts with flat chondrite-normalized REE patterns, lack of a subduction-related trace element signature, pillow lavas that indicate submarine volcanism, low abundance of volcaniclastic deposits, lack of sheeted dyke complexes, and a relatively thick (~5 km) extrusive section (Kerr et al., 2000). Based upon these characteristics, known occurrences of the CLIP have been identified and studied through oceanic drilling, submersible collections, and on-land exposures throughout the Caribbean, South America, and Central America. Known occurrences of the CLIP include DSDP and ODP sites, the Beata Ridge, the Dumisseau Formation and Duarte Complex in Hispaniola, the Bath-Dunrobin Formation in Jamaica, the Leeward Antilles, the Cordillera of western Colombia, the Western Cordillera of Ecuador, the Nicoya Complex of Costa Rica, and the Azuero-Sona Complex in Panama (Fig. 2). In the sections below, key data from these different areas of the CLIP are summarized, except for the Leeward Antilles, which forms the focus of a more detailed later section.
**DSDP Leg 15 & ODP Leg 165**

DSDP (Deep Sea Drilling Project) Leg 15 at Sites 146, 150, 151, 152, and 153 (Fig. 2; Donnelly, 1973; Donnelly et al., 1973) and ODP (Ocean Drilling Project) Leg 165 at Site 1001 (Fig. 2; Sigurdson et al., 1997; Sinton et al., 2000) sampled the upper ~400-1000 m of the CLIP. These drilled sections consist of thick, coarse-grained, basaltic sills or flows that are overlain by or intrude into limestone (Donnelly et al., 1973). Geochemical analyses of rocks collected by these drilling studies indicate an oceanic plateau affinity (Donnelly et al., 1973; Bence et al., 1975; Sinton et al., 1998; Hauff et al., 2000; Kerr et al., 2002a, 2002b).

Sinton et al. (1998) obtained \(^{40}\text{Ar}/^{39}\text{Ar}\) ages of 90.6 ± 3.2 Ma and 92.1 ± 4.7 Ma for DSDP Leg 15 Site 146 and 94.3 ± 2.8 Ma for DSDP Leg 15 Site 150. These ages are both in agreement with Coniacian sediments that overlie DSDP basalts (Edgar and Saunders, 1973). However, Sinton et al. (1998) consider the age results from DSDP Leg 15 to be imprecise due to low K\(_2\)O values and alteration of collected samples. Sinton et al. (2000) reported three \(^{40}\text{Ar}/^{39}\text{Ar}\) ages of basalts that range in age from 80.8-81.3 Ma for ODP Leg 156 Site 1001. This is in agreement with Campanian sediments intercalated within the basalts (Edgar and Saunders, 1973).

**Beata Ridge**

The Beata Ridge is a topographic structure that trends SSW from Cape Beata in Hispaniola over a distance of 450 km and divides the Caribbean into the Colombia and Venezuela basins (Fig. 2; Fox et al., 1970; Mauffret and Leroy, 1997; Mauffret et al., 2001). Samples of the Beata Ridge were collected by dredging (Fox et al., 1970), drilling during the DSDP Leg 15 (Donnelly et al., 1973), and from the submersible Nautil (Mauffret et al., 2001). The samples consist mainly of gabbros and dolerites, along with rare pillow basalts (Mauffret et
Geochemical analysis of gabbros and dolerites from the Beata Ridge revealed that these samples are similar to basalts collected from DSDP and ODP Sites, as well as other known samples of the CLIP (Revillion et al., 2000).

Revillion et al. (2000) reported \(^{40}\)Ar/\(^{39}\)Ar ages that indicate that the predominate age range of both intrusive and volcanic rocks are between 80-75 Ma with a subordinate intrusive phase of 56-55 Ma that has been related to a localized lithospheric thinning event. However, these radiometric ages are inconsistent with the age of intercalated Turonian-Cenomanian sediments that occur within basalts recovered during Nautile studies (Mauffrêt et al., 2001) and intercalated Turonian-Santonian sediments recovered at DSDP sites of the southern Beata Ridge (Edgar and Saunders, 1973).

**The Dumisseau Formation and Duarte Complex**

The Dumisseau Formation and the Duarte Complex represent two fragments of oceanic plateau exposed on the island of Hispaniola (Fig. 2; Lewis et al., 1983, 2002; Sen et al., 1988; Draper and Lewis, 1989, 1991; Lewis and Jimenez, 1991; Lapierre et al., 1997, 2000). Many workers have assigned these rocks to the CLIP, based upon both geochemical and petrological analysis (e.g. Sen et al., 1988; Lapierre et al., 1997, 2000; Sinton et al., 1998; Kerr et al., 2003).

The Dumisseau Formation, located in southwestern Haiti (Fig. 2), is divided into an upper and lower unit. Both units consist of pillowed and massive basalts and minor picrites that are interlayered with minor stratified sedimentary layers. The upper and lower units are also locally cut by gabbroic intrusions and dolerite dykes (Maurasse et al., 1979). Sinton et al. (1998) reported five \(^{40}\)Ar/\(^{39}\)Ar ages from basalts of the lower unit that range between 88.7 ± 1.5 Ma and 92.0 ± 4.8 Ma. However, paleontological evidence from sediments of the lower unit are possibly early Cretaceous-Cenomanian in age (Maurasse et al., 1979), which does not agree with
radiometric ages obtained by Sinton et al. (1998). There are no direct radiometric age data for basalts of the upper unit; however, intercalated sediments of Late Campanian age (Maurasse et al., 1979) and a K-Ar date of a late stage sill that intrudes the upper basalts with an age of 75 ± 1.5 Ma (Sen et al., 1988) suggests that the upper unit is younger than the lower unit. These upper and lower units are overlain by late Campanian to Maastrichtian limestones (Maurasse et al., 1979).

The Duarte Complex, located in the Central Cordillera of the Dominican Republic (Fig. 2), is divided into an upper and lower unit (Palmer et al., 1979). The lower unit consists of massive and banded picrites and basalts that are locally capped by breccias and intruded by synvolcanic dikes and sills of basalt and dolerite (Escouder-Viruete et al., 2007). The upper unit consists of a massive and homogeneous pile of basaltic submarine flows with sparse interlayered carbonate pelagic sediments, and rare intrusions of synvolcanic mafic dikes (Escouder-Viruete, 2007). The picrites and high-Mg basalts are chemically related to plume magmas (Draper and Lewis, 1991; Lewis and Jimenez, 1991). A geochronological study by Lapierre et al. (1999) obtained ⁴⁰Ar/³⁹Ar amphibole ages of 86.1 ± 1.3 Ma for a picrite and 86.7 ± 1.6 Ma for an amphibolite. However, Lewis et al. (1999) argue that these ages are due to thermal effects associated with metamorphism and do not represent the primary crystallization age of the Duarte Complex. In an effort to further constrain the timing of tectonometamorphism of the Duarte Complex, Escouder-Viruete et al. (2007) reported two ⁴⁰Ar/³⁹Ar ages from amphibolites of 93.9 ± 1.4 Ma and 95.8 ± 1.9 Ma. Escouder-Viruete et al. (2007) propose that the protoliths of the amphibolites must be older than the reported Cenomanian ⁴⁰Ar/³⁹Ar ages due to metamorphism. This hypothesis is consistent with a Sm/Nd isochron age of 115 ± 20 Ma obtained from eight samples of picrites and high-Mg basalts of the Duarte Complex (Escuder-Viruete et al., 2004).
and Albian to Upper Cenomanian fossils found within the overlying Tiero Formation (Montgomery and Pessagno, 1999).

**Bath-Dunrobin Formation**

The Bath-Dunrobin Formation is located within Jamaica (Fig. 2) and consists of a thick sequence of late Cretaceous extrusive, tholeiitic, massive basalts that occur with intercalated island arc tuffs (Hastie et al., 2008). The Bath-Dunrobin lavas are exposed as two fault bounded blocks that cover an area of 40 km$^2$ (Wadge et al., 1982). Based on major element and limited trace elemental data, Jackson et al. (1980) and Jackson (1987) concluded that the Bath-Dunrobin lavas have a MORB affinity. However, other studies have shown that the Bath-Dunrobin lavas are similar to samples of DSDP Leg 15 (e.g. Wadge and Draper, 1978; Wadge et al., 1982) and are more likely representative of the CLIP (Donnelly et al., 1990; Kerr et al., 2003; Hastie et al., 2008). There are no available radiometric ages for the Bath-Dunrobin Formation; however, Hastie et al. (2008) concluded that the Bath-Dunrobin Formation is likely Turonian-Coniacian in age based upon the ages of radiolarian found within interbedded red cherts and mudstones (Montgomery and Pessagno, 1999). The Bath-Dunrobin Formation is overlain by a thick sequence of Maastrichtian volcaniclastic rocks (Wadge et al., 1982).

**The Central Cordillera, Western Cordillera, and Serriano de Baudo of Colombia**

In western Colombia, mafic rocks outcrop into three north-south trending, fault-bounded belts that are named (from east to west) the Central Cordillera, Western Cordillera, and Serriano de Baudo. These three terranes in western Colombia have been shown to be geochemically consistent with derivation from an oceanic plateau and linked to the CLIP (Marriner and Millward, 1984; Millward et al., 1984; Nivia, 1987; Storey et al., 1991; Kerr et al., 1996, 1997, 2002, 2003). Most authors agree that these allochthonous belts were accreted to the South
American margin in the Late Cretaceous or early Tertiary (e.g. McCourt et al., 1984; Millward et al., 1984; Bourgois et al., 1987).

The Central Cordillera consists of Cretaceous massive and pillowed picritic to tholeiitic basalts with some ultramafic cumulates (McCourt et al., 1984; Spadea et al., 1989). These mafic rocks occur in several discontinuous lenses and are bounded to the east by the Romeral fault, which separates continental crust from oceanic crust (Case et al., 1973). $^{40}\text{Ar}/^{39}\text{Ar}$ ages were obtained from picrites of the Central Cordillera and these yield ages of $93.21 \pm 3.60 \text{ Ma}$ and $88.95 \pm 3.27 \text{ Ma}$ (Kerr et al., 2002b). However, these ages are not in agreement with a Rb-Sr isochron age of $99 \pm 4 \text{ Ma}$ obtained from a cross-cutting arc-related intrusion of the Buga batholith (McCourt et al., 1984).

The Western Cordillera is separated from the Central Cordillera by the Cauca-Patia graban and separated from the Serriano de Baudo by the San Juan Atrato trough (Kerr et al., 1997). The Western Cordillera consists of tectonic slices of pillowed and massive basalts, dolerites, local gabbros, and rare tuffs that are separated by steeply dipping fault-bounded lenses of metasediments (Barrero, 1979; Aspden, 1984). Basalts of the Western Cordillera were dated by $^{40}\text{Ar}/^{39}\text{Ar}$ methods and yield two distinct ages of $91.7 \pm 2.7 \text{ Ma}$ and $76.5 \pm 1.6 \text{ Ma}$ (Kerr et al., 1997; Sinton et al., 1998). The ~92 Ma radiometric ages are consistent with Cenomanian-Turonian and Turonian-Coniacian ages from intercalated sediments (Bourgois et al., 1987; Barrerro, 1979). Early Tertiary subduction related volcanics occur on the western periphery of the Western Cordillera (Tistl and Slazar, 1994) and have $^{40}\text{Ar}/^{39}\text{Ar}$ ages of $43.1 \pm 0.4 \text{ Ma}$ (Sinton, 1996).

The Serrania de Baudo is the westernmost mafic belt and crops out along the northwest Pacific coast of Colombia (Goossens et al., 1977; Macia, 1985; Kerr et al., 2002b). It consists of
pillowed and massive basalts with some basaltic breccias, dolerites and gabbros (Goossens et al., 1977; Macia, 1985). $^{40}\text{Ar}/^{39}\text{Ar}$ analyses from Serrania de Baudo basalts yield ages of 72.3 ± 0.4 Ma and 77.9 ± 1.0 Ma (Kerr et al., 1997), which are consistent with Upper Cretaceous bivalves found within the basalts by Gansser (1973). This unit is overlain by a poorly exposed sequence of subduction related basalts that are intercalated with Eocene limestones (Gansser, 1973).

**Western Cordillera of Ecuador**

The Western Cordillera of Ecuador consists of multiple allochthonous oceanic blocks that were accreted to the South American margin from the Late Cretaceous to Eocene (Goosens and Rose, 1973; Feininger and Bristow, 1980; Kerr et al., 2002; Jaillard et al., 2004; Pratt et al., 2005; Spikings et al., 2005; Luzieux et al., 2006; Vallejo et al., 2006). The San Juan, Guaranda, and Pinon terranes (from east to west) are located in the Western Cordillera of Ecuador. These terranes have been shown to be both petrologically and geochemically consistent with an oceanic plateau origin, and are interpreted to be parts of the CLIP (Kerr et al., 1996, 2002; Reynaud et al., 1999; Lapierre et al., 2000; Pourtier, 2001; Mamberti et al., 2003, 2004; Vallejo et al., 2009).

The San Juan terrane (eastern part of the Pallatanga terrane of McCourt et al., 1998; Kerr et al., 2002) outcrops in a narrow belt at the eastern border of the Ecuadorian Western Cordillera and consists mostly of ultramafic rocks that include peridotites, layered cumulates, and gabbros (Mamberti et al., 2004). Lapierre et al. (2000) obtained a Sm-Nd isochron of 123 ± 13 Ma from a gabbro thought to be within the San Juan terrane. However, Vallejo et al. (2009) suggest that the age obtained from Lapierre et al. (2000) may be due to erroneous mapping and speculated that the dated sample is actually part of the Peltec unit, which is part of an arc related sequence that was accreted to the South American margin in the Aptian (Litherland et al., 1994). Mamberti et al. (2004) also obtained a poor $^{40}\text{Ar}/^{39}\text{Ar}$ integrated age from a gabbro within the
San Juan Terrane of 105 Ma. A more recent U-Pb SHRIMP age of 87.10 ± 1.66 Ma was determined from extracted zircons within a layered gabbro of the San Juan terrane (Vallejo et al., 2006; Vallejo, 2007). The San Juan Terrane is overlain by a turbiditic sequence, known as the Yunguilla Formation, of Early Maastrichtian age and these strata are interpreted to postdate the accretion of the San Juan terrane to the South American margin (Bristow and Hoffstetter, 1977; Jaillard et al., 2004). A Late Campanian age of accretion of the San Juan terrane is therefore established due to the stratigraphic relationship of the San Juan terrane with the Early Maastrichtian Yunguillia Formation (Hughes and Pilatasig, 2002; Kerr et al., 2002; Jaillard et al., 2004, 2008, 2009).

The Guaranda terrane (western part of the Pallatanga terrane of McCourt et al., 1998; Kerr et al., 2002) is located to the west of the San Juan terrane and these two terranes are separated by a major fault (Hughes and Pilatasig, 2002; Jaillard et al., 2009). The Guaranda terrane consists of hyaloclastics, pillow basalts, dolerites, gabbros, high-Mg basalts, ankaramites, and picrites (Mamberti et al., 2003). Although there are no direct geochronological data for the Guaranda terrane, it is overlain by either pelagic cherts that yield Santonian-Maastrichtian radiolaria (Boland et al., 2000) or the 85-72 Ma lavas and volcaniclastic products of the Rio Cala island arc (Luzieux et al., 2006; Vallejo et al., 2006, 2009). These relations suggest a Late Cretaceous age for the Guaranda terrane. The latest Cretaceous oceanic cherts are unconformably overlain by micaceous quartz-sandstones of Early and Middle Paleocene age (Hughes et al., 1998). The abrupt arrival of the quartz-sandstones are interpreted to be the result of the accretion of the Guaranda terrane to the South American margin, which occurred in the Middle to Late Maastrichtian (Jaillard et al., 2004).
The Pinon terrane is located in the coastal area of the Western Cordillera and consists of basalts, pillow basalts, dolerites, and small gabbroic intrusions (Jaillard et al., 2009). $^{40}\text{Ar}/^{39}\text{Ar}$ analyses from Pinon basalts yielded an age of $88 \pm 1.6$ Ma (Luzieux et al., 2006; Luzieux, 2007). The Pinon basement is overlain by a thick series of andesitic breccias, tuffs, and basaltic lavas that are then overlain by black siliceous limestones of Coniacian age (Reynaud et al., 1999). Jaillard et al. (2009) proposed a Late Paleocene age for the accretion of the Pinon terrane to the South American margin based upon the unconformable deposition of quartz-rich deposits of latest Paleocene age within the Pinon terrane. These quartz-rich deposits are interpreted to represent the arrival of continent-derived sediments from South America (Jaillard et al., 1995; Boland et al., 2000; Deniaud, 2000)

**The Nicoya Complex**

The Nicoya Complex is a tectonic block located within the Nicoya Peninsula of Costa Rica at the southwestern edge of the Chortis block (Fig. 2; Hauff et al., 2000). It consists of massive and pillowed basalts, volcanic breccias, dolerites, gabbros, plagiogranite intrusives, hyaloclastites, cherts, and limestones (Dengo, 1962; Kuijpers, 1980) and is widely believed to be part of the CLIP based upon its geochemical similarity to other known rocks of the CLIP (Donnelly, 1994; Sinton et al., 1997; Hauff et al., 2000; Hoernle et al., 2004; Denyer et al., 2006; Denyer and Gazel, 2009). Multiple $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological studies have been conducted on the Nicoya Complex and these data suggest an age range of ~50 m.y. Sinton et al. (1997) and Hauff et al. (2000) obtained ages that ranged from 84-83 Ma for intrusives and 95-88 Ma for basalts of the Nicoya Complex. However, Hoernle et al. (2004) reported two separate age ranges for basalts of 119-111 Ma and 139-133 Ma, which are significantly older than previous geochronological studies. Two facies of ribbon radiolarian cherts are intercalated within the
Nicoya Complex and yield paleontological ages: Fe-radiolarites with a Coniacian-Santonian age and Mn-radiolarites with a Bajocian to Albian age (Kuijpers, 1980; Baumgartner, 1984; Denyer and Baumgartner, 2006). The age of the Fe-radiolarites (Coniacian-Santonian) are largely synchronous with the earlier radiometric ages reported by Sinton et al. (1997) and Hauff et al. (2000). However, the age of the Mn-radiolarites (Bajocian-Albian) are inconsistent with reported $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Nicoya Complex because there has not been any Middle Jurassic basement found in the Nicoya Complex. The Nicoya Complex is overlain by Upper Cretaceous sediments (~75 Ma) from the Central American arc, which provide a minimum age for this complex (Hauff et al., 1997).

**Azuero-Sona Complex**

The Azuero-Sona Complex is located within the Azuero and Sona Peninsulas of the Pacific margin of southwestern Panama (Fig. 2; Goosens et al., 1977). The basement of the Azuero-Sona complex consists of rocks that range from highly tectonized and metamorphosed basalts to relatively undeformed pillow basalts (Del Giudice and Recchi, 1969; Metti and Recchi, 1972; Krawinkel et al., 1999). Based upon geochemical and petrological characteristics, the basement rocks of the Azuero-Sona Complex are similar to that of the Nicoya Complex and are inferred to represent part of the CLIP (Wildberg et al., 1984; Hauff et al., 1997, 2000; Sinton et al., 1997, 1998; Hoernle et al., 2002, 2004; Kerr et al., 2003, Worner et al., 2009). The Azuero Peninsula yields $^{40}\text{Ar}/^{39}\text{Ar}$ ages of basalts between 66 to 50 Ma (Hoernle et al., 2002). The Sona Peninsula basalts are slightly older based upon $^{40}\text{Ar}/^{39}\text{Ar}$ geochronological studies by Hoernle et al. (2002) and Hoernle and Hauff (2007) indicating an age of 71 ± 2 Ma for one analyzed grain. These older ages are consistent with the paleontological age of the Ocu Formation (Upper Cretaceous; Del Giudice and Recchi, 1969), which unconformably overlies the Azuero-Sona
Complex. However, the younger ages from the Azuero Peninsula are inconsistent with the fossil ages of the Ocu Formation.

**Summary of known CLIP occurrences**

Geochronological studies of known occurrences of the CLIP from oceanic drilling, submersible collections, and on-land exposures result in an age range of ~90 m.y. for the CLIP (Fig. 3; Table 1; 139-50 Ma; Berriasian-YPresian). The majority of the ages obtained within these studies do occur within the ~88-92 Ma interval that is seen widespread throughout the CLIP by previous geochronological studies (Fig. 3; Table 1; e.g. Alvarado et al., 1997; Sinton and Duncan, 1997; Sinton et al., 1998; Lapierre et al., 1999; Walker et al., 1999; Revillon et al., 2000). However, due to the extensive duration of CLIP magmatism (~90 m.y.), previous interpretations for the formation of the CLIP within a short interval of time (1-3 m.y.) should be reevaluated (e.g. Sinton et al., 1998; Kerr et al., 2003).

The accuracy of ages obtained within these studies should also be reexamined. The majority of the geochronological data from the CLIP were analyzed with the $^{40}\text{Ar}/^{39}\text{Ar}$ method (Fig. 3; Table 1). As discussed earlier, the $^{40}\text{Ar}/^{39}\text{Ar}$ method can result in imprecise ages when geochronological data is obtained for oceanic plateau rocks because these rocks are commonly altered and contain low potassium contents. Another important concern is the disagreement between $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric data from the CLIP with paleontological evidence that is overlain/intercalated with CLIP occurrences (Table 1). These discrepancies provide doubt in the validity of ages reported when the $^{40}\text{Ar}/^{39}\text{Ar}$ technique is used.
Table 1: Data from Known Occurrences of the CLIP. See figure 2 for CLIP locations. All ages shown are from $^{40}$Ar/$^{39}$Ar analyses, unless noted by the symbol # representing U-Pb zircon analysis and * representing IN-SIMS analysis. Data that is red in color indicates geochronological data that is inconsistent with either intercalated or overlying sediments.

Geologic time scale from Gradstein et al., 2004.

<table>
<thead>
<tr>
<th>Location</th>
<th>Age (Ma)</th>
<th>Intercalated Sediments</th>
<th>Overlying Sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td>DSDP Leg 15 Site 146</td>
<td>90.6 - 92.1</td>
<td>N/A</td>
<td>Coniacian (89.3 - 85.8)</td>
</tr>
<tr>
<td>DSDP Leg 15 Site 150</td>
<td>94.3 ± 2.8</td>
<td>N/A</td>
<td>Coniacian (89.3 - 85.8)</td>
</tr>
<tr>
<td>ODP Leg 156 Site 1001</td>
<td>80.8 - 81.3</td>
<td>N/A</td>
<td>Campanian (83.5 - 70.6)</td>
</tr>
<tr>
<td>Beata Ridge</td>
<td>56 - 85</td>
<td>Cenomanian - Turonian (99.6 - 89.3) Turonian - Sanionian (93.5 - 83.5)</td>
<td>Mid Eocene (48.6 - 40.4)</td>
</tr>
<tr>
<td>Dumousseau Formation</td>
<td>92 - 88</td>
<td>Early Cretaceous - Cenomanian (112.0 - 93.5)</td>
<td>Campanian-Maastrictian (83.5 - 65.5)</td>
</tr>
<tr>
<td>Upper</td>
<td>N/A</td>
<td>Late Campanian (75.0 - 70.6)</td>
<td>Campanian-Maastrictian (83.5 - 65.5)</td>
</tr>
<tr>
<td>Duarte Complex</td>
<td>96 - 86</td>
<td>N/A</td>
<td>Albion - Upper Cenomanian (112 - 96)</td>
</tr>
<tr>
<td>Bath-Dunrobin Fm</td>
<td>N/A</td>
<td>Turonian - Coniacian (93.5 - 85.8)</td>
<td>Maastrictian (70.6 - 65.5)</td>
</tr>
<tr>
<td>Colombia</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Central</td>
<td>93.2 - 89.0</td>
<td>N/A</td>
<td>Paleogene (65.5 - 23.03)</td>
</tr>
<tr>
<td>Western</td>
<td>91.7 ± 2.7</td>
<td>Cenomanian - Turonian (99.6 - 89.3) Turonian - Coniacian (93.5 - 85.8)</td>
<td>Paleogene (65.5 - 23.03)</td>
</tr>
<tr>
<td>Serriano de Baudo</td>
<td>77.9 - 72.3</td>
<td>Upper Cretaceous</td>
<td>Eocene (55.8 - 33.9)</td>
</tr>
<tr>
<td>Ecuador</td>
<td>105 87.0 ± 1.6</td>
<td>N/A</td>
<td>Early Maastrictian (70.6 - 65.5)</td>
</tr>
<tr>
<td>Guaranda</td>
<td>N/A</td>
<td>N/A</td>
<td>Santonian - Maastrictian (85.8 - 65.5)</td>
</tr>
<tr>
<td>Pinon</td>
<td>88.6 ± 1.6</td>
<td>N/A</td>
<td>Coniacian (89.3 - 85.8)</td>
</tr>
<tr>
<td>Nicoya Complex</td>
<td>95 - 88</td>
<td>Bajocian - Albion (171.6 - 99.6) Coniacian - Santionian (89.3 - 83.5)</td>
<td>Upper Cretaceous</td>
</tr>
<tr>
<td>Azuero-Sona Complex</td>
<td>66 - 50</td>
<td>N/A</td>
<td>Upper Cretaceous</td>
</tr>
<tr>
<td>Aruba Lava Formation</td>
<td>99.4 ± 4.5*</td>
<td>Turonian (93.5 - 89.3)</td>
<td>Eocene (55.8 - 33.9)</td>
</tr>
<tr>
<td>Curacao Lava Formation</td>
<td>89.5 ± 1.0</td>
<td>Albion (112 - 99.6)</td>
<td>Santonian (85.8 - 83.5)</td>
</tr>
<tr>
<td>Gran Roque Gabro</td>
<td>87.0 ± 4.1*</td>
<td>N/A</td>
<td>Neogene (23.03 - 0)</td>
</tr>
</tbody>
</table>
Figure 2: A map that shows the extent of the Caribbean plate and locations of on land CLIP fragments within the Caribbean, Central America, and South America. Abbreviations are as follows: A (Aruba), C (Curaçao), B (Bonaire), GR (Gran Roque) (modified from Kerr et al., 2003).
Figure 3: A graphical display of radiometric ages of known CLIP occurrences from the Caribbean, Central America and South America. The black stars represent IN-SIMS ages, the white stars represent U-Pb SHRIMP ages, the open circles represent $^{40}\text{Ar}/^{39}\text{Ar}$ ages, and the brackets represent an age range for multiple analyses. Geologic time scale from Gradstein et al., 2004.
CHAPTER 3

NEW IN-SIMS STUDIES IN THE LEEWARD ANTILLES

The Leeward Antilles represent isolated exposures of a submarine ridge that extends from Aruba on the west to La Blanquía on the east that trends parallel to the transpressional Caribbean-South American plate boundary zone in a WNW-ESE direction (Fig. 2; Beardsley and Ave Lallemant, 2007). The Leeward Antilles islands of Aruba, Curacao, Bonaire, and Gran Roque were chosen to test the application of the IN-SIMS technique. These islands are ideal for geochronological studies within the Caribbean plate because they contain on-land exposures of both the CLIP and Cretaceous magmatic arc assemblages that have been tectonically uplifted along the South American continental margin (Beets et al., 1984; White et al., 1999; Kerr et al., 2003; Thompson et al., 2004). Exposures on these islands allow for lower sections of the magmatic sequence to be sampled in order to gain a full perspective on magmatic processes that occur within the Caribbean plate. These islands have also been recently remapped in detail to provide accurate field relations for each island (e.g. Wright and Wyld, 2004, 2010). Another benefit for new geochronological studies is that various age dating techniques have been utilized in prior studies that allow for comparison between these previous methods and new IN-SIMS age results. These previous studies also indicate discrepancies between previous age determinations and paleontological evidence (Table 1) that suggests that there may be problems with the accuracy of previous geochronological data from the Leeward Antilles.

This study utilizes the IN-SIMS technique on both oceanic plateau and arc magmatic rocks of the Leeward Antilles islands of Aruba, Curacao, Bonaire, and Gran Roque (Fig. 2). The
IN-SIMS technique will provide U-Pb ages that bypass the potential problems encountered when using the $^{40}\text{Ar}/^{39}\text{Ar}$ method to date mafic rocks. This improved resolution in geochronological data will provide insights into the temporal evolution of both the CLIP and related arcs of the Leeward Antilles.

In the following sections, an overview of the regional geology for Aruba, Curaçao, Bonaire, and Gran Roque will be presented, with an in-depth description of each unit in which samples were collected for IN-SIMS geochronology. Discussion of previous geochronological studies and paleontological evidence will also be addressed to compare with new IN-SIMS results from this study.

**ANALYTICAL METHODS**

Hand-sample sized samples were collected from the Leeward Antilles islands of Aruba, Curacao, Bonaire, and Gran Roque (Fig. 2). These samples were collected at localities with minimal degrees of weathering and where clear contact relations with other units are present. Samples were stored in sealed plastic and fabric bags in order to prevent any contamination by mixing with other samples, dirt, or dust. Polished thin sections were then prepared from each of the samples. One polished thin section was used for IN-SIMS analysis from each of the Leeward Antilles islands of Aruba, Curacao, Bonaire, and Gran Roque.

Polished thin sections of collected samples were analyzed by a microprobe at the University of Wyoming, which used x-ray mapping of zirconium and back-scattered electron (BSE) imaging. Potential zirconium targets were then differentiated by electron dispersive spectrometry (EDS) to determine if the grains were zircon, baddeleyite, or another mineral. Images of potential targets were taken under varying magnifications (40x, 300x, and 2000x) with BSE to locate these grains under reflected light (Fig. 4A). These targets were then found in
reflected light and images were taken under various magnifications (10x, 20x, and 40x objective lenses) (Fig. 4B). The 40x images in reflected light are comparable in scale to the 2000x BSE images.

Potential target regions of the polished thin sections were cut out and mounted on an epoxy disk along with previously polished zircon and baddeleyite standards (Fig. 4C). These mounts were then gold-coated for conductivity purposes. Images were taken of the gold-coated mounts under varying magnifications (10x, 20x, 40x objective lenses) for identification while using the SIMS (Fig. 4D).

The CAMECA ims1270 SIMS at UCLA was used to conduct the in-situ U-Pb SIMS (IN-SIMS) technique. Due to the small size (<10 um) of target grains, a field aperture was utilized in order to block secondary ions emitted from host phases. Analyses consisted of seven cycles for standards and 15 to 30 cycles for zircon and baddeleyite targets. Each cycle measured counts of $^{203.5}$ (background), $^{94}$ZrO$^+$, $^{204}$Pb$^+$, $^{206}$Pb$^+$, $^{207}$Pb$^+$, $^{208}$Pb$^+$, $^{232}$Th$^+$, $^{238}$U$^+$, $^{238}$UO$^+$, and $^{238}$UO$_2$$^+$. Pb/U instrumental fractionation of the unknowns were corrected by $^{206}$Pb/$^{238}$U relative sensitivity vs. UO$_2$$^+$/U$^+$ calibrations. Data reduction and error calculations used UCLA in-house software ZIPS version 3.1.4. A detailed explanation of operating conditions for IN-SIMS U-Pb analysis is described in Schmitt et al. (2010).

**ARUBA**

The basement of Aruba consists of the Late Cretaceous Aruba Lava Formation (ALF) and cross-cutting 89 ± 1 Ma Aruba batholith, which are unconformably overlain by Eocene limestones and younger strata (Fig. 5; Westermann, 1932; Beets and MacGillavry, 1977; Helmers and Beets, 1977; Monen, 1977; Beets et al., 1984, 1996; Jackson and Robinson, 1994; White et al., 1999; Wright and Wyld, 2004, 2010).
The Aruba Lava Formation (ALF)

The ALF is located within the central part of Aruba and covers an area of approximately 20 km$^2$, truncated at the northeastern side by the Caribbean Sea and elsewhere by the composite Aruba batholith (Fig. 5; Beets et al., 1984, 1996). Beets et al. (1984) described the ALF as a sequence of weakly metamorphosed mafic lavas, diabase, and pyroclastic and volcaniclastic sediments. Snoke et al. (2001) performed geochemical analysis on samples from the ALF and concluded that the lavas represent an island arc succession. However, based upon the presence of high-MgO lavas, the absence of a subduction related trace-element signature, and the chemical similarity to other analyzed samples of the CLIP, most authors now agree that the ALF represents a fragment that belongs to the CLIP (e.g. Kerr et al., 1997, 2003; Sinton et al., 1998; White et al., 1999; Wright and Wyld, 2004, 2010).

The ALF has been interpreted by previous mapping studies to be an unbroken stratigraphic succession intruded by diabase sills at all levels (Beets et al., 1984, 1996). However, recent mapping by Wright and Wyld (2010) identified an angular unconformity that separates the ALF into two separate stratigraphic packages, and these are referred to as the upper and lower ALF (Fig. 5). The following summary of geological relations on Aruba is taken from Wright and Wyld (2010), except where noted otherwise.

The lower ALF is divided into three mappable units: the basalt, argillite, and diabase units (Fig. 5). The basalt unit consists mostly of pillowed and massive basalt flows with rare layers of associated volcaniclastic strata. The overlying argillite unit is composed mostly of thinly bedded argillite with less common volcaniclastic strata. Ammonites have been found in strata of the argillite unit that indicate a Turonian age (Fig. 5; MacDonald, 1968). The basalt and
argillite units are both intruded by a complex assemblage of texturally variable dikes, sills, and small intrusions that make up the diabase unit.

The upper ALF consists of a locally preserved basal unit of basaltic tuffs and an overlying polymictic conglomerate that overlie the lower ALF along an angular unconformity (Fig. 5). The tuff unit consists of well-bedded accretionary lapilli tuff that is locally dominated by tuff breccias. The conglomerate unit consists of a well-rounded pebble to cobble polymictic conglomerate, in which clasts are derived entirely from lower units. Wright and Wyld (2010) conclude that the unconformity between the upper and lower ALF represents a period of uplift and erosion, and marks a distinct transition from marine deposition to subaerial conditions.

White et al. (1999) attempted to obtain $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the ALF, however the Ar systematics of the samples were too disturbed for reliable age results to be determined. However, the age of the ALF is constrained by the presence of a Turonian fossil found within sedimentary intercalations of the lower ALF (MacDonald, 1968) and the cross-cutting 89 ± 1 Ma Aruba batholith (Wright and Wyld, 2010).

**Deformation, Metamorphism, and Plutonism**

Previous work attributed deformation and metamorphism of the ALF to the intrusion of the Aruba batholith (Westermann, 1932; Monen, 1977; Beets et al., 1984, 1996; White et al., 1999), which has been dated at 89 ± 1 Ma (U-Pb zircon; Wright and Wyld, 2010). Recent mapping and structural studies, however, indicate that deformation (folds and foliation) and metamorphism (subgreenschist grade) is regional in the ALF and not spatially associated with the Aruba batholith (Wright and Wyld, 2010). In addition, deformation structures are cut discordantly by dikes associated with the Aruba batholith and are overprinted by contact metamorphism in a narrow zone around the batholith (Wright and Wyld, 2010). These relations
led Wright and Wyld (2010) to conclude that the Aruba batholith emplacement postdates the regional deformation of the ALF.

**IN-SIMS Results from the ALF**

Samples were collected from the diabase unit of the lower ALF (see figure 5 for sample location). The mineralogy of the diabase consists of plagioclase + actinolite + clinopyroxene + epidote + chlorite + quartz + opaques (Figs. 6A & 6B). Ages of nine baddeleyite analyses yielded a mean age of 101.8 ± 8.5 Ma (95% conf.) (Fig. 7A). Six zircon analyses yielded a mean age of 97.3 ± 5.2 Ma (95% conf.) (Fig. 7B). A mean age of both the baddeleyite and zircon analyses resulted in an age of 99.4 ± 4.5 Ma (95% conf.). The mean age of 99.4 ± 4.5 Ma from both baddeleyite and zircon analyses is considered to be representative of the age of the diabase unit of the lower ALF because the resultant ages of the baddeleyite and zircon analyses overlap within error.

**Implications for IN-SIMS Results from the ALF**

An IN-SIMS age of 99.4 ± 4.5 Ma was determined for the diabase unit of the ALF from both nine baddeleyite and six zircon analyses. This age for the diabase unit of the ALF is slightly older than a Turonian fossil found within the lower ALF (Fig. 5; MacDonald, 1968) and indicates that the ALF is at least Cenomanian in age.

**CURAÇAO**

The basement of Curaçao consists of the Cretaceous Curaçao Lava Formation (CLF) that is overlain unconformably by the Late Cretaceous to Early Paleocene Knip Group and Midden-Curaçao Formation (Fig. 8; Beets, 1972; Klaver, 1976; Beets et al., 1977; Wright and Wyld, 2004, 2010). These rocks are then unconformably overlain by Eocene limestone and younger strata (Fig. 8).
The Curaçao Lava Formation (CLF)

The CLF is a 5 km thick succession that covers most of Curaçao, and consists of predominantly picritic to basaltic pillow lava flows with minor reworked hyaloclastites and some diabasic sills and dikes (Fig. 8; Beets et al., 1984; Klaver, 1987). Kerr et al. (1996) showed that the picrites of the CLF are related to the basalts by simple fractional crystallization of olivine, clinopyroxene, and plagioclase. As shown by multiple authors, CLF lavas show a similarity in major and trace element chemistry to other samples of the CLIP, especially the ALF (e.g. Donnelly and Rogers, 1978; Beets et al., 1982, 1984; Donnelly et al., 1990; Kerr et al., 1996; Sinton et al., 1998; White et al., 1999).

Early geochronological studies by Santamaria et al. (1974) reported two K-Ar analyses of whole rock dolerites collected within the CLF with ages of 118 ± 10 Ma and 126 ± 12 Ma. A more recent investigation by Sinton et al. (1998) obtained two whole rock $^{40}\text{Ar}/^{39}\text{Ar}$ analyses that yielded plateau ages of 89.5 ± 1.0 Ma and 88.0 ± 1.2 Ma for CLF basalts collected at the base and top of the CLF, respectively. However, these $^{40}\text{Ar}/^{39}\text{Ar}$ ages are not consistent with an Albian ammonite found in a pelagic interval of the CLF (Fig. 8; Weidman, 1978). Kerr et al. (2003) argue that the ammonites may be reworked or misidentified due to the condition of the fossils being poorly preserved, incomplete, and distorted. However, Snoke et al. (2001) maintain the ammonites accurately reflect the age of the CLF and volcanism began in the Albian. A minimum age can be concluded for the CLF lavas based upon the 86.2 ± 0.8 Ma age determined for a diorite dike that intrudes the CLF (U-Pb Zircon; Wright and Wyld, 2004).

Deformation and Metamorphism

Curaçao was affected by a single phase of regional deformation and metamorphism that affected all units older than the Eocene, including the Midden-Curaçao Formation (Beets, 1972).
Beets (1972) and Klaver (1987) described this Paleocene event to be characterized by NW-trending folds on both the outcrop and map-scale where they control the map pattern of bedrock units. Wright and Wyld (2010) reported that mineral assemblages indicate that metamorphism occurred at a phrenite-pumpellyite to zeolite grade.

**IN-SIMS Results from the CLF**

Samples were collected from diabase within the CLF (see figure 8 for sample location). The mineralogy of the diabase consists of plagioclase + actinolite + clinopyroxene + opaques (Figs. 9A & 9B). Ages of eight baddeleyite analyses yielded a mean age of $114.7 \pm 7.4$ Ma (95% conf.). One zircon analysis yielded an age of $102 \pm 6$ Ma (95% conf.). A mean age of both baddeleyite and zircon analyses yield an age of $112.7 \pm 7.3$ Ma (95% conf.) (Fig. 10). An age of $114.7 \pm 7.4$ Ma from baddeleyite analyses is considered to be representative of the age of the diabase within the CLF because only one zircon grain was analyzed and the age was considerably younger than the eight analyzed baddeleyite grains.

**Implications for IN-SIMS Results from the CLF**

IN-SIMS analyses of diabase within the CLF yielded an age of $114.7 \pm 7.4$ Ma from eight baddeleyite analyses. This Aptian age is significantly older than previously reported Turonian-Coniacian $^{40}$Ar/$^{39}$Ar ages (ca. 89-90 Ma) of basalts sampled from the base and top of the CLF (Sinton et al., 1998). However, the IN-SIMS age of the CLF does agree within error to an Albian ammonite found within a pelagic interval of the CLF (Fig. 8; Wiedman, 1978), which would indicate that the Turonian-Coniacian $^{40}$Ar/$^{39}$Ar ages obtained by Sinton et al. (1998) may not truly represent the age of the CLF.
The bedrock geology of Bonaire has been defined as a continuous succession of volcaniclastic, sedimentary, and intrusive rocks known as the Late Cretaceous “Washikemba Formation” (Pipjers, 1933; Klaver, 1976; Beets et al., 1977, 1984; Thompson, 2002; Thompson et al., 2004; Wright and Wyld, 2004). The “Washikemba Formation” is unconformably overlain by Maastrichtian and younger strata (Fig. 11). Prior studies have utilized both petrological and geochemical analyses to demonstrate that the “Washikemba Formation” does not contain exposures of the CLIP, but records a succession of volcanic rocks with a magmatic arc signature (e.g. Klaver, 1976; Donnelly and Rogers, 1980; Beets et al., 1984; Thompson, 2002; Thompson et al., 2004). These rocks of Bonaire are viewed to be part of the southern edge of the Cretaceous “Great Arc of the Caribbean” (Beets et al., 1984; Thompson et al., 2004).

Recent mapping by Wright and Wyld (2010) indicates that the “Washikemba Formation” of northwest Bonaire cannot be considered part of a single stratigraphic succession, but actually consists of two separate units that are in fault contact (Fig. 11). Wright and Wyld (2010) therefore renamed these rocks as the Washikemba Group and the Matijs Group. The Matijs Group refers to the Salina Matijs assemblage of prior workers (Klaver, 1976; Beets et al., 1977; Thompson, 2002), and is dominated by sedimentary rocks. The Washikemba Group consists almost entirely of igneous rocks, and refers to the remainder of the “Washikemba Formation” as previously defined in northwest Bonaire. The following summary of geologic relations on Bonaire is taken from Wright and Wyld (2010), except where noted otherwise.

**Washikemba Group (WG)**

The WG comprises the Wecua, Slagbaai and Branderis assemblages of prior workers (Klaver, 1976; Thompson, 2002) and consists mostly of felsic volcanics and shallow-level
intermediate and felsic intrusions (Fig. 11; Klaver, 1976; Thompson, 2002; Wright and Wyld, 2010). Two different suites of hypabyssal rocks intrude the WG. The lower part of the WG are intruded by diorite dikes and small sills, while the upper part are intruded by rhyodacite sills. At the top of the WG, Wright and Wyld (2010) located a distinctive conglomerate unit that was derived entirely from the underlying WG.

The age of the WG is constrained by the presence of mid to Late Albian ammonites from the lower part of the WG (Beets et al., 1977) and \(^{40}\text{Ar} / ^{39}\text{Ar}\) ages of 95 ± 2 Ma (Thompson, 2002) and 96 ± 4 Ma (Thompson et al., 2004) from the upper part of the WG. Wright and Wyld (2010) recently obtained similar U-Pb zircon ages of 98.2 ± 0.6 Ma and 94.6 ± 1.4 Ma from the upper part of the WG. Available data thus indicates that the WG accumulated over the interval from Albian to Cenomanian.

**Matijs Group (MG)**

The MG is divided into three separate sedimentary units: an argillite unit that is locally cross-cut by diabase stocks, a conglomerate unit, and a chert unit (Fig. 11).

The Matjis argillite occurs at the base of the unit and consists of pelagic and hemipelagic strata. The argillite unit is intruded by mafic stocks that are found only in the argillite unit (Fig. 11), and exhibit arc related trace element characteristics. The diabasic intrusions found in the MG are different than those in the WG due to the diabase being much less weathered, coarser grained, non-vesicular, and forming discrete bodies that have narrow contact metamorphic aureoles (Klaver, 1976; Thompson, 2002). The overlying conglomerate, referred to by prior workers as ‘boulder beds’ (Klaver, 1976; Thompson, 2002), consists of a distinctive polymictic conglomerate and breccia beds. The conglomerate unit contains paleontological evidence that
indicates a Coniacian age (Beets et al., 1977). The chert unit consists of pelagic strata that include thinly bedded chert, radiolarian chert, and argillaceous chert.

**IN-SIMS Results from the Matijs diabase of Bonaire**

Samples were collected from a diabase intrusion within the argillite unit of the Matijs Group (see figure 11 for sample location). The mineralogy of the diabase consists of plagioclase + epidote + clinopyroxene + opaques (Figs. 12A & 12B). Ages of seven baddeleyite analyses yielded a mean age of $111.6 \pm 5.1$ Ma (95% Conf.) (Fig. 13A). Five zircon analyses yielded a mean age of $93 \pm 10$ Ma (95% Conf.) (Fig. 13B). A mean age of both the baddeleyite and zircon analyses resulted in an age of $102 \pm 7.7$ Ma (95% Conf.). An age of $111.6 \pm 5.1$ Ma from seven baddeleyite analyses is considered to be representative of the age of the Matijs diabase because the zircon analyses resulted in discordant ages (Fig. 13B).

**Implications for IN-SIMS Results from the Matijs diabase of Bonaire**

IN-SIMS geochronological results from the diabase unit within the MG yielded an age of $111.6 \pm 5.1$ Ma from seven baddeleyite analyses. This IN-SIMS age in combination with Cenomanian U-Pb SHRIMP ages ($98.2 \pm 0.6$ Ma and $94.6 \pm 1.4$ Ma; Wright and Wyld, 2010) from the WG indicates an overall span of arc activity that occurred on Bonaire from the Albian to Cenomanian (ca. 112-95 Ma). The $111.6 \pm 5.1$ Ma age of the Matijs diabase also provides geochronological data to support mapping evidence of Wright and Wyld (2010) that a fault separates the MG and WG. This fault relationship is clearly evident based upon the overlying MG (Albian) is older in age than the underlying WG (Cenomanian) (Fig. 11).

**GRAN ROQUE**

Gran Roque is located within the Los Roques Archipelago and is the only island that contains exposures of igneous and metamorphic basement beneath Neogene cover. The
basement of Gran Roque consists of a mafic complex of amphibolites intruded by altered gabbro. The mafic complex is intruded by tonalitic and quartz-dioritic intrusions with an age of 65.6 ± 1.4 Ma (U-Pb Zircon; Wright and Wyld, 2004), that occur as small bodies and dikes. All of these units are then intruded by numerous aplite and pegmatite dikes, with no available age information (Fig. 14; Aguerrevere and Lopez, 1938; Rost, 1938; Schubert and Moticska, 1972, 1973; Wright and Wyld, 2004)

**Mafic Complex (Amphibolites + Gabbro)**

The mafic complex of Gran Roque consists of amphibolites and altered cross-cutting, coarse-grained gabbro (Fig. 14; Aguerrevere and Lopez, 1938; Schubert and Moticska, 1972, 1973). Mineralogically the amphibolites consist of hornblende + plagioclase + epidote ± quartz ± opaques and the gabbro consists of plagioclase + clinopyroxene + actinolite. Early geochemical analysis of both the amphibolites and gabbro indicate a tholeiitic affinity in which the rocks may have formed in either a MORB or island arc setting (Santamaria and Schubert, 1974; Ostos, 1990; Ostos and Sisson, 2005). However, further geochemical analysis by Giunta et al. (2002) indicates that the amphibolites and gabbros have an oceanic plateau affinity. New geochemical data from this study is consistent with Giunta et al.’s (2002) hypothesis that the mafic complex of Gran Roque is representative of an oceanic plateau and new data from this study indicate that both the amphibolites and gabbros of Gran Roque lack a subduction related trace-element signature, have flat chondrite normalized REE patterns, and correlate well with geochemical data of known CLIP rocks of Aruba and Curacao (Figs. 15A & 15B). This new geochemical data suggests a CLIP origin for the mafic complex of Gran Roque. The rocks of the mafic complex locally display spheroidal weathering features in which the weathering rinds are replaced by phosphate minerals (Marcano, 1873; Aguerrevere and Lopez, 1938; Rost, 1938).
Wright and Wyld (2004) postulate that the origin of these phosphate minerals is due to the reaction of guano from Cretaceous sea birds with the mafic complex. There is no evidence of spheroidal weathering on any of the younger arc intrusions and in places these arc rocks actually intrude into the phosphatized mafic complex.

The only direct age information available for the mafic complex are two K-Ar ages of $127 \pm 15$ Ma and $130 \pm 14$ Ma from hornblende separates of the amphibolites (Santamaria and Schubert, 1974). However, the age of the cross-cutting $65.6 \pm 1.4$ Ma quartz-diorite provides an upper age limit for the mafic complex of Gran Roque (Wright and Wyld, 2004).

**IN-SIMS Results from the Gabbro of the Mafic Complex of Gran Roque**

Samples were collected from the gabbro unit within the mafic complex (see figure 14 for sample location). The mineralogy of the gabbro unit consists of plagioclase + actinolite + clinopyroxene + opaques (Figs. 16A & 16B). Ages of 15 zircon analyses yielded a mean age of $87.0 \pm 4.1$ Ma (95% Conf.) (Fig. 17).

**Implications for IN-SIMS Results from the Gabbro of the Mafic Complex of Gran Roque**

An IN-SIMS age of $87.0 \pm 4.1$ Ma was determined for the cross-cutting gabbro of the mafic complex of Gran Roque from 15 zircon analyses. This Coniacian age provides an upper limit for the age of the mafic complex of Gran Roque.

**SUMMARY OF NEW IN-SIMS DATA OF THE LEEWARD ANTILLES**

IN-SIMS age results from this study of CLIP exposures from Aruba, Curaçao, and Gran Roque indicate a ~30 m.y. span of CLIP magmatism that occurred within the Leeward Antilles from the Aptian to Coniacian (ca. 115-87 Ma). An IN-SIMS age of $99.4 \pm 4.5$ Ma was determined for the diabase unit of the ALF from both nine baddeleyite and six zircon analyses. This age for the diabase unit of the ALF is slightly older than a Turonian fossil found within the
lower ALF (Fig. 5; MacDonald, 1968) and indicates that the ALF is at least Cenomanian in age. IN-SIMS analyses of diabase within the CLF yielded an age of 114.7 ± 7.4 Ma from eight baddeleyite analyses. This Aptian age is significantly older than previously reported Turonian $^{40}\text{Ar}/^{39}\text{Ar}$ ages (ca. 89-90 Ma) of basalts sampled from the base and top of the CLF (Sinton et al., 1998). However, the IN-SIMS age of the CLF does agree within error to an Albian ammonite found within a pelagic interval of the CLF (Fig. 8; Wiedman, 1978), which would indicate that the Turonian $^{40}\text{Ar}/^{39}\text{Ar}$ ages obtained by Sinton et al. (1998) may not truly represent the age of the CLF. An IN-SIMS age of 87.0 ± 4.1 Ma was determined for the cross-cutting gabbro of the mafic complex of Gran Roque from 15 zircon analyses. This Coniacian age provides an upper limit for the age of the mafic complex of Gran Roque.

IN-SIMS geochronological results from the diabase unit within the MG yielded an age of 111.6 ± 5.1 Ma from seven baddeleyite analyses. This IN-SIMS age in combination with Cenomanian U-Pb SHRIMP ages (98.2 ± 0.6 Ma and 94.6 ± 1.4 Ma; Wright and Wyld, 2010) from the WG indicates an overall span of arc activity that occurred on Bonaire from the Albian to Cenomanian (ca. 112-95 Ma). The 111.6 ± 5.1 Ma age of the Matijs diabase also provides geochronological data to support mapping evidence of Wright and Wyld (2010) that a fault separates the MG and WG (Fig. 11).
Figure 4: Photos of sample preparation for IN-SIMS analyses. (A) BSE images of potential targets (40x, 300x, 2000x). (B) Reflected light image of potential targets (40x objective). (C) Reflected light image of gold-coated target sections that are mounted on an epoxy disk along with baddeleyite and zircon standards (4x objective). (D) Reflected light image of gold-coated potential targets (40x objective).
Figure 5: Geology of the island of Aruba. Inset shows a simplified map of the island (from Beets et al., 1996). Detailed geologic map of the central part of the island (see location in inset) and a generalized time-stratigraphic column from the Cretaceous to Eocene (modified from Wright and Wyld, 2010). The location of geochronological samples from this study are shown.
Figure 6: Photomicrographs of the diabase unit of the ALF (4x objective). (A) Plane polarized light. (B) Cross polarized light.
Figure 7: Concordia plots for IN-SIMS analyses of the diabase unit within the lower ALF. See figure 5 for sample location. (A) Nine baddeleyite analyses. (B) Six zircon analyses.
Figure 8: Geology of the island of Curaçao. Inset shows a simplified map of the island (from Beets, 1972). Detailed geologic map of the northwest part of the island (see location in inset) and a generalized time-stratigraphic column of Curaçao geology from the Cretaceous to Eocene (modified from Wright and Wyld, 2010). The location of geochronological samples from this study are shown.
Figure 9: Photomicrographs of diabase from the CLF (4x objective). (A) Plane polarized light. (B) Cross polarized light.
Figure 10: Concordia plot of IN-SIMS analyses of eight baddeleyite and one zircon from diabase within the CLF. See figure 8 for sample location.
Figure 11: Geology of the island of Bonaire. Inset shows a simplified map of the island (from Klaver, 1976 and Thompson, 2002). Detailed geologic map of northwest Bonaire based on new mapping from Wright and Wyld (2010) and previous mapping by Beets (1972), Beets et al. (1977), and Klaver (1976), which shows the fault contact between the Washikemba Group and Matijs Group. The location of geochronological samples from this study are shown. A revised generalized time-stratigraphic column of Bonaire geology from the Cretaceous to Eocene is also shown (modified from Wright and Wyld, 2010).
Figure 12: Photomicrographs of the diabase within the argillite unit of the Matijs Group (4x objective). (A) Plane polarized light. (B) Cross polarized light.
Figure 13: Concordia plots for IN-SIMS analyses of diabase within the argillite unit of the Matijs Group. See figure 11 for location. (A) Seven baddeleyite analyses. (B) Five zircon analyses.
Figure 14: A simplified map of the island of Gran Roque, located in the Los Roques Archipelago (modified from Wright and Wyld, 2004). The location of geochronological samples from this study are shown.
Figure 15: Geochemical plots comparing the amphibolites and gabbro (mafic complex) of Gran Roque to the geochemical range of the ALF and CLF. In all figures, the “diamond shape” represents the amphibolites of Gran Roque, the “X shape” represents the gabbros of Gran Roque, and the grey shading represents the range of geochemical data from the ALF.
and CLF. (A) Multi-element plot normalized to primitive mantle (Sun and McDonough, 1989). (B) Chondrite-normalized REE diagram (Sun and McDonough, 1989).

Figure 16: Photomicrographs of the gabbro unit of the mafic complex of Gran Roque (4x objective). (A) Plane polarized light. (B) Cross polarized light.
Figure 17: Concordia plot of IN-SIMS analyses of 15 zircon from the gabbro unit of the mafic complex of Gran Roque. See figure 14 for sample location.
CHAPTER 4

IMPLICATIONS FOR THE TEMPORAL EVOLUTION OF THE CLIP

The extensive duration of CLIP magmatism (~ 30 m.y.) within the Leeward Antilles islands of Aruba, Curaçao, and Gran Roque suggests that the simple plume head model for the formation of the CLIP within a short period of time (ca. 92-88 Ma) must be reevaluated (e.g. Sinton et al., 1998; Kerr et al., 2003). To explain the protracted history of the CLIP, two different models are presented: the Leeward Antilles islands of Aruba, Curaçao, and Gran Roque may be representative of (1) multiple pulses of oceanic plateau magmatism or (2) a continuous period of magmatism that lasted from the Aptian to Coniacian.

Due to the modality of ages presented with IN-SIMS analyses, the Leeward Antilles islands of Aruba, Curaçao, and Gran Roque could represent multiple pulses of oceanic plateau magmatism (Fig. 18). This age distribution can be accomplished by one hot spot with multiple pulses (Mauffret and Leroy, 1997) or by multiple hotspots derived from a similar source (Hauff et al., 2000). The similarity of IN-SIMS ages within error from Gran Roque and Aruba (Fig. 18), in conjunction with a Turonian fossil found intercalated in the ALF (Fig. 5) could indicate that these two exposures of the CLIP formed during the same pulse of plateau magmatism. However, CLIP exposures from Aruba and Curaçao could have also formed through entirely separate events. An IN-SIMS age of 114.7 ± 7.4 Ma from the CLF is significantly older and does not overlap with IN-SIMS ages from either Aruba or Gran Roque (Fig. 18). An Albian fossil found intercalated within the CLF (Fig. 8) also supports the IN-SIMS age of the CLF. This would suggest that the CLF may have formed by an entirely separate pulse of plateau magmatism that
was significantly older than the earlier pulse or pulses that created the CLIP exposures on Aruba and Gran Roque.

Another possible explanation for the IN-SIMS age distribution is that the CLIP exposures of the Leeward Antilles represent a continuous period of CLIP magmatism that occurred from the Aptian to Coniacian (ca. 115-87 Ma). However, if the CLIP was formed through an extended period of magmatism, the CLIP could not be considered a LIP (Coffin and Eldholm, 1994) and previous interpretations for the rate of emplacement of the CLIP would be drastically reduced.
Figure. 18: An X-Y plot that graphically shows the IN-SIMS ages, with error bars, of CLIP exposures from Aruba, Curaçao, and Gran Roque. Geologic time scale from Gradstein et al., 2004
CHAPTER 5

FURTHER RESEARCH WITH THE IN-SIMS TECHNIQUE

The IN-SIMS U-Pb geochronological technique effectively obtained precise age data from mafic exposures of both oceanic plateau and arc magmatic rocks of the Caribbean. Due to the fact that IN-SIMS age determinations were conducted only within the Leeward Antilles islands of Aruba, Curaçao, Bonaire, and Gran Roque, definitive conclusions for the overall timing and duration of the CLIP and early arc magmatism cannot be made. However, important contributions are presented to guide further research projects that involve the CLIP and early arc magmatism.

Based upon the extended age range of IN-SIMS data, the hypothesis that the CLIP formed over a short period of time through a single pulse of magmatism (e.g. Sinton et al., 1998; Kerr et al., 2003) should be reevaluated. This hypothesis was mainly based upon $^{40}$Ar/$^{39}$Ar radiometric ages from previous studies of the CLIP (Fig. 3). However, as observed through this study, there were significant differences between radiometric ages obtained with the $^{40}$Ar/$^{39}$Ar method versus the IN-SIMS technique (Fig. 3). The validity of the IN-SIMS ages were upheld because the IN-SIMS ages also correlate with paleontological evidence found on Leeward Antilles islands (Table 1). The discrepancy between $^{40}$Ar/$^{39}$Ar and IN-SIMS ages would imply that other $^{40}$Ar/$^{39}$Ar ages of the CLIP could also be incorrect. To correct for these discrepancies, CLIP samples dated previously by the $^{40}$Ar/$^{39}$Ar method should be dated again with the IN-SIMS technique to provide more precise and accurate ages. Reevaluation of these samples with the IN-SIMS method may present different ages than previous $^{40}$Ar/$^{39}$Ar studies that could change
interpretations for the temporal evolution of the CLIP. In addition to the reevaluation of $^{40}\text{Ar}/^{39}\text{Ar}$ studies of the CLIP, more samples from ocean drilling, dredging, and on-land exposures of the CLIP also need to be collected and dated with the IN-SIMS technique to provide a more robust data set for conclusions on the timing and duration of CLIP magmatism to be determined.

Use of the IN-SIMS technique would provide improved resolution compared to traditional dating methods because oceanic plateau samples may be altered or have low potassium contents. The increase in accuracy of the IN-SIMS technique is important because this would provide a more reliable geochronological method to determine the temporal evolution of other oceanic plateaus throughout the world. This information could then be used to determine trends in oceanic plateau magmatism, such as their timing and duration.

Based upon the protracted history of IN-SIMS ages from CLIP exposures of the Leeward Antilles (Fig. 18), this could imply that other oceanic plateaus may form in a similar manner. This would suggest that oceanic plateaus may not form by a single, short-lived pulse of magmatism, but would occur over much longer periods of time than previously believed. The IN-SIMS technique can now provide valuable insight into the true formational history of oceanic plateaus from around the world.
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### APPENDIX A

REDUCED DATA OF IN-SIMS ANALYSES FROM THE LEEWARD ANTILLES

<table>
<thead>
<tr>
<th>Name</th>
<th>Age (Ma)</th>
<th>Correlation of Concordia Ellipses</th>
<th>% Radiogenic U/238U</th>
<th>% Radiogenic U/207Pb</th>
<th>Th/U</th>
<th>U/238U</th>
<th>U/235U</th>
<th>204Pb/206Pb</th>
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<td>207Pb/206Pb</td>
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ZIRCON (Bad = Rejected)

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<th>Name</th>
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<th>% Radiogenic U/238U</th>
<th>% Radiogenic U/207Pb</th>
<th>Th/U</th>
<th>U/238U</th>
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Figure A1: Reduced data from both baddeleyite and zircon analyses of the diabase unit of the Aruba Lava Formation (ALF).
Figure A2: Reduced data from both baddeleyite and zircon analyses of diabase from the Curaçao Lava Formation (CLF).
**Figure A3:** Reduced data from both baddeleyite and zircon analyses of the Matijs Diabase of Bonaire.
Figure A4: Reduced data from zircon analyses of the gabbro unit of Gran Roque.