Geochemistry, Weathering and Diagenesis of the Bermuda Paleosols:

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GEOCHEMISTRY, WEATHERING AND DIAGENESIS

OF THE

BERMUDA PALEOSOLS

A thesis

by

JOEL A. FRISCH

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of the requirements

for the degree of

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GEOCHEMISTRY, WEATHERING AND DIAGENESIS OF THE BERMUDA PALEOSOLS

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ABSTRACT

Pleistocene-age terra rossa paleosols are situated on and are intercalated with eolianite and marine carbonate units across the Bermuda Islands. These clay-rich soils were originally thought to the derived from weathering of the volcanic seamount and/or from dissolution of the carbonate units, the paleosols are now believed to be primarily the result of atmospheric dust deposition from Saharan North Africa and the Sahel via long range transport, with some local inputs. If so, these soil units are mixtures of atmospheric deposition during one or more glacialinterglacial cycles. Previous investigations have been conducted on the paleosols to determine their provenance, age, and to identify unique characteristics for island wide mapping. We conducted comprehensive geochemical analyses to determine the degree of chemical weathering and diagenesis, and to identify processes responsible for their formation and development. The paleosols were found to be geochemically similar across all ages, and to show an increased degree of alteration with age rather than with their duration of subaerial exposure, indicating diagenesis by infiltrating meteoric waters as well subaerial weathering. Evidence of paleosol diagenesis suggests vadose flow across the island may not be limited to preferential pathways and that while flow through the limestones is complex, infiltrating waters appear to have allowed for additional alteration of the soils.

In addition to the paleosols, clay-rich deposits with paleosol-like textures were identified during coring operations in Harrington Sound and Hungry Bay, beneath present-day sea level. The source and development histories of these materials were previously unknown. Since these clay deposits are situated beneath present-day sea level it is likely that they were deposited and chemically weathered exclusively during glacial low-sea level climate conditions. Geochemical analyses were conducted on the submarine clay samples to determine if they were related to the above-sea level paleosol and to identify their sources. Major and trace element signatures showed the submarine clay deposits to be chemically similar to the paleosols and to be derived from a similar upper continental crust-like parent. Trace element fingerprinting showed the samples to be derived from a parent similar to that of the paleosols; primarily atmospheric dust with some volcanic contributions. These findings provide additional evidence that trade wind vectors for dust transport were present during Pleistocene glacial climate conditions. Weathering indicators reveal the submarine clay samples to be somewhat less weathered than paleosols of similar age and comparable periods of exposure. Like the paleosols, the submarine clays underwent an initial period of rapid subaerial weathering which suggests warm humid climate conditions during glacial low sea level periods. However, the submarine clays did not experience extended periods of diagenesis, which may explain the somewhat lower degree of weathering. Evidence of inputs from the volcanic platform to the paleosols was limited, but comparisons with shallow volcanic rock and highly weathered volcanic residual known as the Primary Red Clay showed some similarities, suggesting that in-situ chemical weathering of the volcanic platform could produce a laterite with some characteristics similar to the Bermuda paleosols.

Geochemical analysis of volcanic sands collected at Whalebone Bay showed the igneous fragments to be a result of mechanical weathering and sorting of heavy refractory minerals and we interpret these sediments to be best described as a beach placer deposit. These materials are enriched in insoluble trace elements and REE, and their contribution to the paleosols is limited.

DEDICATION

To my advisor Prof. Rudi Hon for his patience, faith in me, his guidance and support throughout this project. May he enjoy a long, happy and healthy retirement on Cape Cod. I will be forever grateful. Thanks to Dr. Stan Herwitz for providing the samples, his time and attention, and the excellent advice he offered me. Thanks to Prof. Jeremy Shakun for his insightful recommendations that significantly helped to focus and improve this thesis. Thanks also to Dr. Dan Muhs of the U.S. Geological Survey and Dr. Mark Rowe of Birbeck University of London for their time and helpful recommendations. In special appreciation for my friendship with Prof. David Roy, who through thoughtful discussions during long canoe trips and swarms of black flies in the wilds of Maine, taught me a lot about myself. May he rest in peace. To the Department of Earth and Environmental Sciences and the Morrissey Graduate School of Arts and Sciences for allowing me the opportunity to complete my degree. To Steve and Sally Dean, who gave me an amazing opportunity and foundation, and without whose mentorship I would never be where I am today professionally. To my business partner Jay Billings who has been a mentor and friend. To my wife's parents, Bill and Judy for all their care and support over the years. To my parents that gave me everything I ever needed and taught me by example, the value of hard work. To my children who constantly inspire me.

Especially to my wife Suzanne, who from the day we met made me believe in myself, whose encouragement and support helped me through my years at Clark University and Boston College, starting a family and a business, and all years since, and whose caring and thoughtfulness know no bounds; to you I dedicate this work.

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January 1832 Cape Verde

"Generally, the atmosphere is hazy; and this is caused by the falling of impalpably fine dust, ... From the direction of the wind whenever it has fallen, ... we may feel sure that it all comes from Africa ... The dust falls in such quantities as to dirty everything on board, ... It has fallen on ships when several hundred, and even more than a thousand miles from the coast of Africa, ... I was much surprised to find particles of stone above the thousandth of an inch square, mixed with finer matter."

> Charles Darwin Voyage of the Beagle

1.0 INTRODUCTION

Reddish-brown, clay-rich paleosols are present across the Bermuda archipelago, on and intercalated with carbonate dune deposits of Pleistocene age (Sayles, 1931). The source of the paleosols has been debated for nearly a century, with theories ranging from insoluble residue from the dissolution of the carbonates (Verrill, 1907; Sayles, 1931) to weathering of the basement volcanics (Blackburn and Taylor, 1969; Prognon et al, 2011) and even deposition from birds or wind blow sources (Sayles, 1931). The prevailing model is that the paleosols are mostly the result of deposition and weathering of atmospheric dust from North Africa during one or more glacial-interglacial cycles, and that the fossil soils represent periods of long-term stability of the underlying carbonate sands (Bricker and Prospero, 1969; Muhs et al, 2012). In addition to the paleosols, clay-rich soil-like deposits have been identified below present-day sea level, whose composition and provenance are to date, unknown.

Previous investigations have described the mineralogy and geochemistry of the paleosols in an effort to identify unique characteristics for island wide mapping (Bretz, 1960; Land et al, 1967) or for provenance determinations (Bricker and Prospero, 1969; Herwitz and Muhs, 1995; Herwitz et al, 1996, and Muhs et al, 2012). This study seeks to understand the development history of the soils and to identify processes responsible for their present composition.

The geochemistry and mineralogy of the Bermuda paleosols and submarine clays are a function of the weathering and diagenesis of the original parent materials. We conducted geochemical comparisons and fingerprinting techniques (after Muhs et al, 2012) which showed the submarine clays to be derived from a similar parent material as the paleosols and to have undergone subaerial weathering. Based on their position below present day sea-level, the submarine clays would have been deposited and weathered during low-sea level conditions only, during glacial periods. These deposits represent a unique opportunity to understand chemical

weathering conditions different from those experienced by the paleosols positioned above present day sea-level. In addition, we employed a common weathering indicator for the paleosols, submarine clays and parent dust materials, and made comparisons of the Bermuda paleosols with similar soils on other Caribbean Islands. Geochemical comparison of the paleosols of different ages were found to be of similar composition to each other. The degree of weathering of the paleosols and submarine clays was found to be extensive, with most alteration occurring within the first 100 ka. Paleosol weathering was more strongly correlated with the age of the underlying carbonate units than with the duration of exposure, suggesting the paleosols have undergone diagenesis as well as subaerial weathering. The submarine clays were of a similar to somewhatlower degree of weathering, suggesting comparable, wet sub-tropical conditions were present on the island during glacial low-sea level conditions. In addition, we compared the paleosols and submarine clays to previously published data for shallow volcanic samples, clays derived from weathering of the volcanic platform, as well as volcanic beach sands to determine their potential contributions to the paleosols.

2.0 GEOLOGY OF THE BERMUDA ISLANDS

The Bermuda Islands are located approximately 1,000 km east of Cape Hatteras, North Carolina in the western North Atlantic Ocean (Figure 1). The Bermuda Islands sit atop an Eocene age volcanic seamount capped by Pleistocene age carbonate deposits of eolian and marine origin, that are interlayered with clay-rich paleosols. Bermuda is one of four seamounts forming a short chain or cluster atop the Bermuda Rise, a mid-plate positive depth anomaly in the western North Atlantic. The seamount cluster includes Bowditch Seamount to the northeast, and Challenger Bank and Argus Bank to the southwest (Figure 1). The Bermuda Rise may have formed along the Mid-Atlantic Ridge around 100-125 Ma (Vogt and Jung, 2007) with the



Figure 1a. Map showing the location of the Bermuda Islands and other North Atlantic atmospheric dust sampling locations and Pleistocene Age lateritic soils.



Figure 1b. Map of the Bermuda Rise showing the Bermuda Seamount Cluster, the Bermuda platform and the Bermuda Archipelago.

seamounts having formed as volcanic islands during a later episode of volcanic activity around 35-45 Ma. The trend of the Bermuda cluster is nearly perpendicular to that of the North American Plate vector, upon which the Bermuda Rise is fixed, suggesting a non-hot spot formation (Figure 1). This unusual trend has given rise to alternate theories for the origin of the seamounts, such as arising from a large fissure formed along a temporary crustal weakness or from a pulsating "lava-lamp-type" hot spot (Vogt & Jung, 2007). The seamounts extend approximately 4,000 m above the surrounding ocean floor. Measurements of volcanic sediments around their flanks, suggest the peaks rose to elevations of 1,000 meters above sea level at the end of volcanic activity, approximately 35 Ma (Vogt & Jung, 2007). Since the end of volcanism, the peaks have been eroded to the submarine platforms observed today.

2.1 VOLCANIC PLATFORM

Drilling operations and seismic measurements have located the surface of the volcanic platform at depths ranging from approximately -15 m below present day sea level, in the vicinity of Castle Harbour, to -75 m across the platform, with an average of approximately -45 m to -50 m beneath the carbonate deposits (Pirsson, 1914; Wollard and Ewing, 1939; Officer et al, 1952; Stanley and Swift, 1967; Gees and Medioli, 1970; Reynolds and Aumento, 1974; Vogt, 1991; Vacher and Rowe, 1997; Vogt and Jung, 2007; Figure 2). A deep borehole, drilled in the vicinity of the Bermuda Institute of Ocean Sciences at Ferry Reach (Well 1972; Figure 2), reached a depth of 802 m and identified numerous submarine pillow lavas of hydrothermally altered tholeiitic basalts (Reynolds and Aumento, 1974). The lavas consisted of approximately 65% altered lavas and approximately 35% intrusive sheets. Shallow basement rock samples obtained near the Government Quarry (Well 1980b; Figure 2) consisted of a series of albitised and chloritised mafic and ultramafic Oceanic Island Basalts (OIB) into which were injected sheeted dikes or cone sheets of a rare type consisting of low-silica content titaniferous alkaline rock



known as Bermudite (Aumento and Gunn, 2005; Vogt and Jung, 2007).

Figure 2. Map of Bermuda Islands showing places of interest and sample locations, including deep boring locations.

The Bermuda seamount was historically considered tectonically stable and was believed to act as a Pleistocene sea level tide gauge (Land et al, 1967). However, glacio-isostatic adjustment modeling has indicated that loading and unloading of continental ice sheets may have had significant impacts on Bermuda sea levels (Rowe, 2014). Recent field evidence at Bermuda suggesting anomalously high sea levels (>6 m; Rowe et al, 2014) and anomalously low sea levels (~2-7 m; Muhs, 2020) indicate a more complex sea level history than previously thought.

No outcrops of the volcanic basement rocks are present above sea level. Thin laminae of black sands are found within the Pleistocene carbonate Upper Town Hill formation at Whalebone Bay, and black sands have been observed within modern beach sands on Coney Island. These materials are believed to be fragments of the eroded volcanic platform (Woolard and Ewing, 1939; Blackburn and Taylor, 1969; Prognon, 2011; Figure 2). The presence of these deposits

may indicate that outcrops of the volcanic basement coexisted for some time with the carbonates (Rowe, 2020) or perhaps these materials originated from the outer platform and were transported episodically during one or more storm surges (Herwitz et al, 1996). Blackburn and Taylor (1969), via X-ray diffraction and microscopic analysis, reported the presence of apatite, chromite, pink garnet, gold, goethite, perovskite and sphene in the black sands. With the exception of goethite, these minerals are not commonly found in the carbonates or in the paleosols, and are unlikely to have contributed significantly to their composition.

A reddish-brown clay of at least 3 m thickness was identified at depths of 12-15 m in the vicinity of Castle Harbour (Moore and Moore, 1946; Figure 2). This unit, known as the Primary Red Clay, was situated above the volcanic surface and underlies limestones in that area (Foreman, 1951). Samples of this material were obtained from dredging operations during construction of the former American Air Force Base and at Ferry Reach between 1941 and 1943 (Foreman, 1951; Figure 2). The clay was found to contain a mineral assemblage similar to that of the volcanic basement rocks including apatite, augite and other pyroxenes (Moore and Moore, 1946) as well as quartz, zircon, chromite, magnetite, titanite, epidote and feldspar, and the minerals grains were described as being coated by iron oxide (Foreman, 1951). The samples contained relatively high percentages of Al₂O₃ (~22%), Fe₂O₃ (~12.5%), SiO₂ (~12.2%) and TiO₂ (~7.8%) and were described as a laterite, derived from weathering of the volcanic platform, under subaerial weathering conditions in a humid tropical environment (Foreman, 1951).

2.2 CARBONATE UNITS

The volcanic seamount of Bermuda is capped by a series of mid-Pleistocene age limestone deposits, consisting of corals, mollusks, forams and algae fragments (Sayles, 193; Land, McKenzie and Gould, 1967; Vollbrecht, 1990; Figure 3). The limestones include

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Figure 3. Generalized geologic map of the Bermuda Islands showing the mapped carbonate units and an idealized geologic cross-section showing the onlapping nature of the eolianites units (*after Vacher et al, 1989*).

nearshore marine deposits and eolianite. Bermuda is the type locality for eolianite (Sayles, 1931) consisting of lithified dune fields that formed near shore (Bretz, 1960; Vacher, 1971; Vacher, 1973) from onshore winds in all directions (Sayles, 1931; Mackenzie, 1964a; 1964b; Rowe and Bristow, 2015) during a small number of large storms (Vacher, 1973). The majority of the carbonate units that make-up the Bermuda Islands by volume are eolianite (Vacher and Rowe, 1997), but some marine facies are present showing beach-to-dune transitions (Bretz, 1960; Land et al, 1967) and are a record of a marginal marine, strandline environment (Vacher, 1971). The eolianites are lenticular deposits, laterally accreted (Hearty and Vacher, 1994) with a shoreward superposition from both the north and south shores; placing the oldest formations generally in the center of the island and the younger units generally along the shores (Vacher et al, 1995; Figure 3). The carbonate units are, from oldest to youngest: the Walsingham Formation (Qw >880 ka);

the Lower Town Hill Formation (Q_{tl} 450 ka); the Upper Town Hill Formation (Q_{tu} 325 ka); the Belmont Formation (Q_b 125 ka); the Rocky Bay Formation (Q_r) 125 ka); and the South Hampton Formation (Q_s 80 ka) (Vacher et al, 1997; Table 1). Nowhere on the island can a complete section be seen in outcrop and most locations show only a few units together (Muhs et al, 2012; Figure 3 and Figure 4). This is due in part to the onlapping lenticular nature of the deposits and their irregular aerial extent.



Figure 4a. Photograph showing large landward advancing dune overriding a thin, poorly-developed fossil soil known as a protosol; note persons on beach for scale (*photo by Mark Rowe <u>www.bermudageology.com</u>*, *used with permission*).



Figure 4b. Photograph showing eolianite leeward slip-face bedding of dune advancing right to left (*photo by Mark Rowe <u>www.bermudageologv.com</u>*, *used with permission*).



Figure 4c. Photograph showing eolianite leeward slip-face stratification (A) overlain by windward cross-stratification (C) and separated by a bounding surface (B) (*photo by Mark Rowe <u>www.bermudageology.com</u>*, used with permission).

The younger carbonate units consist of recognizable shell fragments with a high degree of primary porosity (Figure 5a). The grains are relatively pure, containing very little insoluble material and are composed of unstable high-magnesium calcite and aragonite (Land, et al, 1967; Vacher et al, 1995; Figure 4). Dissolution of the carbonates over time by infiltrating rainwaters results in a progressive loss of the primary lithic fragments, which dissolve and are reprecipitated within the limestone pore spaces as low-magnesium calcite cement. This diagenetic alteration leads to a higher degree if lithification, the gradual loss of primary porosity, and the development of a moldic secondary porosity (Plummer et al, 1976; Figure 5b). The degree of eolianite alteration and lithification from meteoric and phreatic waters roughly correlates with formation age (Land et al, 1976), but local factors can influence the degree of consolidation (Vacher, 1971) making in-situ identification of individual units based on lithologic criteria difficult (Land et al, 1967). In addition, the variable degree of carbonate diagenesis and karst development (Mylroie et al, 2009), along with fractures systems that are present within the carbonates (Hartsock et al, 1995), and the onlapping nature of the deposits, complicate vadose flow through the units (Vacher, 1974). Evidence for the non-uniform nature of vadose flow can be seen in the variable nature of the limestone lithification, which can result in the formation of vugs containing loose unconsolidated sand within an otherwise well cemented limestone unit.

The timing and processes involved with eolianite formation have also been long debated (Sayles, 1934; Bretz, 1960; Hearty and Vacher, 1994; Rowe and Bristow, 2015). Through facies analysis Rowe and Bristow (2015) have synthesized dune formation as involving an oscillating interglacial relative sea level relating to eustatic and/or isostatic sea level changes. In general, the process typically begins with a prolonged period of offshore interglacial sediment buildup near present day sea level. A rise in sea level triggers sediment delivery to the shores, as onshore waves are able to overcome the barrier effects of the platform shoals. A subsequent sea



Figure 5a. Close-up photograph of the Southampton Formation showing primary lithic fragments and primary porosity (*photo by Mark Rowe <u>www.bermudageologv.com</u>*, used with *permission*).



Figure 5b. Close-up photograph of the Lower Town Hill Formation showing significant loss off primary lithic fragments and primary porosity by dissolution and reprecipitation to form moldic porosity (*photo by Mark Rowe <u>www.bermudageology.com</u>*, used with permission).



Figure 5c. Close-up photograph of the Walsingham Formation showing near total loss of original grains with secondary calcite cements and secondary porosity (*photo by Mark Rowe www.bermudageology.com*, used with permission).

level regression at the onset of glacial climate conditions, exposes the near-shore sediments to transport by on-shore winds (Vacher and Hearty, 1989; Vacher et al, 1995; Rowe and Bristow, 2015). The eolianite deposits often consist of large volumes of material containing very long, continuous laminae, indicating they were deposited during a small number of strong wind storms (Vacher et al, 1995; Vacher and Rowe, 1997; Figure 4). Therefore, while the eolianites represent the majority of the deposits on the island by volume, they represent the shortest depositional time intervals.

Between episodes of dune formation, clay rich soils formed on the eolianite formations. Later episodes of dune formation often resulted in soil burial and preservation as fossil soils. Early investigators, due in part to the observable island stratigraphy, the degree of limestone alteration, and the lack of other obvious sources, suggested dissolution of the carbonates as the main source of the surface soils and paleosols (Verrill, 1907; Sayles, 1931). However, considering the relative purity of the carbonates with minimal insoluble fractions and the unrealistically large volumes of material required to produce the soils of the thickness observed, carbonate dissolution has been shown to be an insignificant source of mineral inputs to the paleosols (Ruhe, Cady, and Gomez, 1961).

2.3 PALEOSOLS

Intercalated between the eolianite and marine carbonate units are paleosols. These units vary from well developed, deep-red to chocolate-brown clay-rich "terra rossa" soils found in hollows (2.5YR, 10R to 5YR, 7.5 YR; Muhs, et al, 2012), to poorly developed buff colored/gray calcareous soils often found on hillsides (Figure 6a & 6b). The paleosols range in thickness from a few centimeters to several meters, and sometimes contain fossil snail shells (Nelson, 1838; Verrill, 1907; Sayles, 1931; Land et al, 1967). The textural differences between soils formed on hillsides versus hollows are believed to be due to rain-wash from slopes (Sayles, 1931, Ruhe et al, 1961). No defining characteristics to identify the individual paleosols are known (Sayles, 1931) and island wide correlation of the paleosols is tenuous except where stratigraphic position has been determined or where direct tracing is possible (Ruhe et al, 1961; Land, McKenzie and Gould, 1967). The more developed and extensive soils are considered to represent long intervals (10 ka to 100 ka) between deposition of the carbonate units (Verrill, 1907; Sayles, 1931). The light-colored poorly developed soils, containing mixtures of clay and calcarenite materials, have been termed regosols (Ruhe et al, 1961) or protosols (Vacher, 1973) and often occur at the transitions between marine and eolianite deposits representing back-beach environments, or at the base of forsets in distal dune environments (Hearty et al, 1992). In some cases, protosols are believed represent short local pauses in eolianite formation, while more extensive protosols are thought to represent intervals of island-wide depositional stability (Hearty and Vacher, 1994).



Figure 6a. Photograph showing a typical terra-rossa paleosol between Lower Town Hill and Upper Town Hill Formations (*photo by Mark Rowe <u>www.bermudageology.com</u>*, used with permission).



Figure 6b. Photograph showing a typical terra-rossa paleosol between Walsingham and Town Hill Formations (*photo by Mark Rowe <u>www.bermudageology.com</u>*, used with permission).

Paleosols are present throughout the islands in a variety of stratigraphic positions and were originally thought to represent residue from limestone dissolution (Verrill, 1907; Sayles, 1931) or the weathering of materials from the volcanic platform. While there are some indications of localized inputs to the soils from volcanic materials (Blackburn and Taylor, 1969; Blackburn and Taylor, 1970; Prognon et al, 2010; Muhs et al, 2012), there is a growing record of evidence for atmospheric dust deposition, primarily by long range transport from North Africa, as a principal parent material for the Bermuda paleosols (Bricker and Prospero, 1969; Bricker and Mackenzie, 1970; Glaccum and Prospero, 1980; Arimoto et al, 1992; Sholkovitz et al, 1993; Herwitz and Muhs, 1995; Herwitz et al, 1996; Muhs et al, 2010; Muhs et al, 2012).

Some of the paleosols formed on carbonate units and were never buried, some paleosols were deposited between carbonates units of sequential age, and some soils were situated between carbonate units of non-sequential age. The soils that were never buried and the paleosols situated between carbonate units of non-sequential age are considered composite soils, they represent more than one soil depositional sequence, and were exposed during more than one glacial-interglacial cycle (Bretz, 1960). Paleosols situated between carbonates units of sequential age are termed Geosols, since they constitute named stratigraphic units (Vacher et al., 1995) and represent soil formation and development during a single glacial-interglacial cycle. The Geosols from oldest to youngest include: Castle Harbour Geosol (>880 ka); Harbour Road Geosol (350 - 450 ka); Ord Road Geosol (250 - 350 ka); and Shore Hills Geosol (170 - 200 ka) (Table 1).

The terra rossa soils show obvious signs of weathering, especially soils formed in depressions, with distinct soil horizons being common, leaching of A horizons, and the accumulation of clay minerals in the B horizons (Bretz, 1960; Ruhe at al, 1961; Muhs et al, 2012; Figure 7). Secondary calcite cement is present in cracks and pore spaces in the A and B horizons (Ruhe et al, 1961) and the general degree of weathering has been suggested to be proportional to the duration of exposure (Hearty and Vacher, 1994). The well-developed paleosols have been



Figure 7. Photograph showing a well-developed terra-rossa paleosol containing soil horizons (*photo by Mark Rowe <u>www.bermudageology.com</u>*, used with permission).

described as aluminous laterites (Hearty and Vacher, 1994), which typically form in warm to hot, wet climates as a result of intense weathering and often contain the minerals gibbsite, boehmite, hematite, goethite and quartz (Patterson, 1967). According to the U.S. Soil Taxonomy, Bermuda soil classifications would range from aluminum and iron-rich Ultisols (intensively-weathered) to Alfisols (less intensively-weathered) (Vacher et al, 1995).

The Bermuda paleosols consist primarily of the minerals kaolinite (20-50%), chlorite (20-40%) and interstratified chlorite/vermiculite and/or hydroxy-interlayered clays (Prognon et al, 2010; Muhs et al, 2012) as well as gibbsite (Ruhe et al, 1961) but no mica or smectite, the lack of which has been cited by as evidence of significant weathering (Herwitz and Muhs, 1995). While previous investigators have identified traces of minerals in the paleosols that may be related to the igneous platform (Verrill, 1907; Sayles, 1931; Prognon et al, 2010) and other minerals more common to continental rocks (Sayles, 1931), the general lack of primary rock forming minerals is an indication of a high degree of pedogenesis. As a result, few age-related differences, with the exception of the presence of Al-rich boehmite occurring only in the oldest of soils (Castle Harbour Geosol), have been noted (Muhs, 2012).

A "soil base" described by Salyes (1931) as consisting of a fine, cream-colored microcrystalline limestone is present beneath the paleosols at soil-limestone interfaces (Herwitz, 1993; Herwitz and Muhs, 1995). These features are also sometimes present between composite paleosols and can be seen beneath modern surface soils or paleosols where eroded. This suggests that the soil base develops during each soil forming episode prior to limestone burial, and that soil formation is episodic, since such features would not likely develop with continuous deposition. Ruhe et al (1961) noted an accumulation of organic carbon and aluminum above the soil base indicating that, once formed, the soil base may be a barrier to vadose flow. Such a barrier would likely limit post-burial chemical alteration of the paleosols by infiltrating meteoric waters.

2.4 BERMUDA CLIMATE

Bermuda is located at 32° 20' N, 64° 45' W and is situated near the current northern limit of subtropical climate conditions (Vollbrecht, 1990; Figure 1). Bermuda's sub-tropical climate is due in part to the Gulf Stream with ocean temperatures ranging from a minimum of 19.3°C (February-March) to a maximum of 27.3°C (August-September) which support hermatypic coral reef species (Vacher and Rowe, 1970). Sea surface temperatures in winter can drop below 18°C for short periods (Vollbrecht, 1990). Rainfall is generally evenly distributed throughout the year with 10-12 cm/month December to July and approximately 13-17 cm/month August through November (~146 cm/yr) (Plummer at al, 1976; Vacher and Rowe, 1997). The yearly average air temperature is 21°C with a monthly temperature range of 10°C and the rainfall and potential evapotranspiration vary seasonally with an annual excess rainfall over evapotranspiration of approximately 10 cm (Plummer et al, 1976). Estimates of recharge through soils are on the order of 18 cm/year November through March, with a soil water deficit occurring July through August (Vacher, 1974).

Recharge from rainfall is rapid and there is no surface runoff on Bermuda, no streams are present, and surface waters are limited to small lakes, ponds and marshes whose surfaces reflect the water table or the ocean surface (Plummer et al, 1976).

2.5 ATMOSPHERIC DUST

Strong evidence has been presented for the deposition of atmospheric dust throughout the western North Atlantic, in quantities sufficient to produce soils over time periods of 10^4 - 10^5 years. Atmospheric dust samples determined to be from North African source areas have been collected throughout the Atlantic Ocean including at Cape Verde, Bermuda, Barbados and Miami, among others (Prospero, 1968; Bricker and Prospero, 1969; Glaccum and Prospero, 1980; Figure 1; Table 2). Through a variety of techniques, including satellite photo interpretation (Prospero et al, 1970); Muhs et al, 2012), radiosonde data (Carlson and Prospero, 1972), satellite measurements with back-trajectory calculations (Husar et al, 1997), and atmospheric modeling (Mahowald et al, 2006), the suspended minerals have been traced to large dust storms emanating from North Africa along the trade winds. In fact, materials derived from North African dust storms have also been identified throughout southern Europe and the Mediterranean (Chester and Johnson, 1971, Muhs et al, 2010) the Middle East (Ganor, 1991) and South America (Prospero et al, 1981). Dust collection programs at Bermuda, Miami, and Barbados have shown that dust delivery to the Western Atlantic varies seasonally and as a function of latitude, with the highest concentrations occurring May through September at lower latitudes (Prospero and Lamb, 2003; Arimoto, 2001). Differences in median grain sizes have also been observed for dust collected at

Cape Verde ($6.16 \mu m$) versus Barbados ($2.06 \mu m$) and Miami ($2.05 \mu m$), and have been attributed to settling during the first few days of the five to seven-day journey across the Atlantic Ocean (Glaccum and Prospero, 1980). Climate modeling has shown increased dust generation and atmospheric transport from North Africa to the Western Atlantic during the last glacial period relative to modern climate conditions (Mahowald et al, 2006).

Assuming atmospheric dust from the Sahara and Sahel regions of North Africa as a primary parent material for the Bermuda soils, a generalized conceptual model of soil development can be made by considering the alteration of the parent minerals due to chemical weathering. Infiltrating meteoric waters are naturally acidic due to the dissolution of atmospheric CO_2 and the formation of carbonic acid (H₂CO₃). Interstitial CO_2 partial pressures (*p*CO₂) in soils are often more than 10 times atmospheric CO_2 levels further reducing soil water pH. In addition, *p*CO₂ is substantially higher in vegetated soils than non-vegetated soils (up to 100 times atmospheric CO_2) as a result of plant and microorganism respiration in the root zone and organic decomposition (Birkeland, 1999). This hydrogeochemical process results in locally acidic waters with a significantly increased chemical weathering potential.

Due to the small grain size of atmospheric dust reaching Bermuda, the current prevailing humid sub-tropical climate, and the soil water conditions present, it can be assumed that the relatively high rainfall acting on the high surface areas of the mineral grains would impart an intense degree of chemical weathering on the soils. The result would lead to: (1) alteration of the feldspars and mica/illite to form hydroxy-interlayered clays, kaolinite, and/or gibbsite; (2) alteration of the Mg/Fe bearing minerals such as biotite/chlorite to form vermiculite, iron oxides and/or hydroxides; (3) a reduction in the amount of silica and mobile elements (Na, Mg, K) relative to other immobile elements such as Al and Ti; (4) elimination of parent-derived calcium silicates and calcite due to dissolution; and (5) the accumulation of amorphous materials and the

more refractory minerals such as quartz. Some idealized mineralogical alterations with the addition of meteoric water and time are summarized as follows:

Orthoclase Feldspar:

Plagioclase Feldspar:

Plagioclase \Rightarrow		Montmorillonite			\Rightarrow	
NaAlSi ₃ O ₈ * CaAlSi	$_{2}O_{8}$	$(Na, Ca)_{0.33}$	$(Al, Mg)_2(Si_4O_{10}$)(OH) ₂		
Montmorillo	onite	\Rightarrow	Kaolinite	\Rightarrow	Gibbsite	

 $Al_2Si_2O_5(OH)_4$

Al(OH)₃

 $(Na, Ca)_{0.33}(Al, Mg)_2(Si_4O_{10})(OH)_2$

Ferromagnesian Minerals:

Biotite K(Mg.I	\Rightarrow Fe) ₃ AlSi ₃ O ₁₀ (OH) ₂	Chlorite Mg ₅ Al(AlSi ₃ O ₁	0)(OH)8	+	Iron Hydroxide Fe(OH) ₃	\Rightarrow
	Chlorite \Rightarrow Mg ₅ Al(AlSi ₃ O ₁₀)(OH) ₈	Kaolinite Al ₂ Si ₂ O ₅ (OH)	⇒)4	Gibbsit Al(OH)	e 33	
	and					
	Iron Hydroxide \Rightarrow Fe(OH) ₃	Hematite Fe ₂ O ₃	⇒	Goethit FeOOH	e [

These incongruent dissolution reactions remove soluble elements (Na, Mg, K, Ca as well as Sr) and result in an enrichment of Al₂O₃, Fe₂O₃ and other immobile elements (Rb, Ti, Zr, Hf, Y, Sc and the Rare Earth Elements) (Taylor and McLennan, 1985).

In summary, paleosols bracketed by eolianites are considered to represent long periods of soil accumulation and weathering between brief episodes of dune formation, and represent accumulations from at least one glacial-interglacial cycle. Atmospheric dust deposition on Bermuda has been continuous during both glacial and interglacial periods, with higher deposition rates suspected during glacial periods. Inputs to the soils from the volcanic basement rocks is evident, at least locally. Episodes of dune formation were initiated due to initial sea level transgressions, followed by sea level lowering that exposed the calcareous near-shore sediments to the wind. Succeeding each dune forming episode, atmospheric dust accumulation increased across the island resulting in the intercalation of soils on and between the limestone units across the entire platform. As each glacial period ended, the rising sea level terminated dust deposition in the areas below present-day sea level. Areas above present-day sea level continued to receive smaller inputs of atmospheric dust during the following interglacial period, until the onset of another glacial period where eolianite formation would have buried the soil units to form paleosols.

3.0 PURPOSE

The purpose of this thesis was to perform a comprehensive geochemical analysis of Bermuda paleosol samples of different ages and stratigraphic positions to assess the degree of chemical weathering and diagenesis as a means to understand the timing, intensity and mechanisms involved. Soil geochemistry and mineralogy generally reflect the climate and hydrologic conditions to which a soil has been exposed, which result in the enrichment or depletion of some elements and distinct mineralogical changes. Comparisons between soils of different ages, different exposure durations, and different stratigraphic positions were used to understand the development history of these soils. Of particular interest were: the differences between soils located above and below present-day sea level; whether the submarine clays had the same parent materials and development histories; how the paleosols and submarine clays compared to the basement volcanics, volcanic sands and weathered volcanic residue (i.e. Primary Red Clay) and what these differences might tell us about the degree of volcanic inputs to the paleosols; whether differences in the soils could be correlated to soil age or soil exposure; what these differences may indicate about glacial vs interglacial climate differences; and what these findings might reveal about the movement of meteoric waters through the vadose zone.

In addition, the geochemistry of the Bermuda paleosols were compared to soils on other Atlantic islands of similar age and source materials, but experienced different climate conditions. Such a comparison may shed light on soil weathering rates and diagenetic differences throughout the tropical and subtropical portions of the western North Atlantic.

4.0 METHODS

4.1 SAMPLES AND COMPARISON DATA

4.1.1 Paleosol Samples

A total of 26 paleosol samples, three submarine clay samples, and one sample of volcanic sands were analyzed as part of this investigation (Table 3). The sample locations are shown on Figure 2. The samples were collected by Dr. Stanley Herwitz, currently of the UAV Collaborative at the NASA Ames Research Center in California, during several research trips to Bermuda between 1993 and 1996. The author accompanied Dr. Herwitz during one such trip in November 1992. The samples collected were in-situ and were located at lateral exposures along road or railway cuts, walls of abandoned quarries, and sites of coastal erosion. Samples were collected from observable B soil horizons after Herwitz et al (1996) and the age and stratigraphic positions were based on the Geologic Map of Bermuda by Vacher and Hearty (1989). At the Marriot Harbor site, several samples were collected to determine geochemical variations throughout a soil profile (Figure 2). Of the Marriot Harbor samples, MH-8 was collected from A horizons contained very low trace element concentrations (Herwitz et al, 1996) that were not useful for geochemical analysis. Therefore, data obtained for sample MH-8 in this study were included in average elemental values but were not included in data correlations.
4.1.2 Submarine Clay Samples

A total of three submarine clay samples were analyzed. Two clay samples recovered from submarine cores advanced at Harrington Sound were acquired from Dr. Rudiger Vollbrecht of the Institute and Museum for Geology and Paleontology, Universitat Gottingen, Germany. Submarine core sample HS-CS/1 was collected at Harrington Sound at a water depth of 21.5 m in which the paleosol was found between 5.3-5.5 m of the core sample, while core sample HS-SP 04 was collected from Harrington Sound at a water depth of 20.7 m in which the soil was identified between 5.8-6.2 m of the core sample (Figure 2). At each of the Harrington Sound core sample locations the soils represent the transition from Holocene to Pleistocene sediments (Vollbrecht, 1990). Logs of the Harrington Sound cores from which the samples were obtained and seismic profiles of the portion of Harrington Sound at each core location are also included in Appendix A.

An additional clay sample was recovered at Hungry Bay (HUN-1) a depth of approximately 2.3-3.0 m during a shallow coring operation through mangrove sediments (Figure 2).

4.1.3 Volcanic Sands

A sample of black sands collected from Whalebone Bay and believed to be weathered volcanic fragments, was analyzed to investigate the source, weathering history, and potential contribution to the Bermuda paleosols (Figure 2).

4.1.4 Published Data Used for Geochemical Comparison

The geochemical data obtained during this investigation were compared and with shallow volcanic core data (-38 m msl) presented by Muhs et al (2012), major element data for the Primary Red Clay (Foreman, 1951), and major element data for African dust (Scheuven, 2013). The data were normalized using CI chondrite and Upper Continental Crust (UCC) data presented by Taylor and McLennan (1985).

4.2 SAMPLE PREPARATION AND LABORATORY ANALYSIS

The Bermuda paleosols vary in composition from mostly clayey unconsolidated deposits to carbonate cemented. In order to perform geochemical analyses, it was necessary to powder the samples. The majority of the samples were loosely consolidated and were gently ground into a coarse powder via a mortar and pestle. Samples containing larger portions of carbonate cement were powdered using a rock shatter box. Aliquots of the coarse powders were separated and analyzed via Instrumental Neutron Activation Analysis and X-ray diffraction. The remaining coarse powders were ball-milled for approximately 5 minutes to 10 minutes to achieve a finer matrix, which was then analyzed via X-ray Fluorescence. Prior to powdering to volcanic sands, the carbonate fraction was removed by immersion in a solution of HCl (6N) and then rinsing with deionized water three times.

4.2.1 X-ray Fluorescence Analysis

X-ray Fluorescence Analysis (XRF) was performed at the Regional Geochemical Laboratory at St. Mary's University, Nova Scotia, Canada. Samples were analyzed for Major Elements: Si, Al, Ti, Fe, Mn, Mg, Ca, K, Na, P and trace elements Cl, V, Cr, Zr, Sc, Be, Nd, La, Ni, Cu, Zn, Ge, Rb, Sr, Y, Nb, Sn, Pb, Th, U utilizing a Philips PW-2400 sequential x-ray fluorescence spectrometer with a Philips PW-2510 sample changer (Table 4a and Table 4b; Table 5a and Table 5b). The fluorescence spectra of each sample are measured, and the results quantified by comparison with a set of geologic standards of known composition. To determine the major elemental concentrations, a 10 g sample was mixed with 1.5 g Bakelite resin, pressed in a Herzog-type hydraulic press to a disk 4 cm in diameter and 5 mm thick. The pellets were baked in an oven at 190 °C for 15 minutes, then stored in a desiccator or analyzed immediately. Sample preparation for trace element analysis involved 1 g of sample mixed with 5 g lithium tetraborate, 300 mg of LiF, and 35 mg of LiBr. The samples were placed in platinum crucibles and melted using a Claisse Fluxy fusion device. The disks, 3 cm in diameter and 3 mm thick, were stored in a desiccator or analyzed immediately.

4.2.2 X-Ray Diffraction Analysis

X-Ray Diffraction Analysis (XRD) was performed at the former Tobin X-ray Laboratory at Boston College utilizing a Phillips PW-1840 X-Ray Diffraction Analyzer to determine general mineralogy (Table 6). Powdered samples were mixed with deionized water to form a slurry and were placed on frosted slides. The slides were allowed to air dry overnight. One of the slides was left untreated. A second slide was placed in a vapor chamber and exposed to ethylene glycol for at least one hour. A third slide was heated to 550°C for one hour to remove interstitial water and organic matter.

Due to limited sample volumes it was not possible to conduct mineralogical analyses of all of the soil samples. Mineralogical analyses were conducted on two of the submarine clay samples and two of the Geosol samples in an effort to make a qualitative comparison of the general composition of the soils and to determine if the submarine clay sample mineralogy was significantly different from the paloeols. Geosol samples MH-8 (Castle Harbour Geosol) and SHR-2 (Shore Hills Geosol) and submarine cores samples HUN-1 and HSCS/1 were selected for analysis by XRD to identify the minerals present (Figure 2).

4.2.3 Instrumental Neutron Activation Analysis

Instrumental Neutron Activation Analysis (INAA) was performed at the former Boston College Department of Geology and Geophysics INAA Laboratory utilizing a Ge 4096 multichannel analyzer to determine the major elements Na₂O, CaO, FeO, Co, Cr, Sc, and Cs; the trace elements Hf, Ta, Th, and U; and the Rare Earth Element La, Ce, Nd, Sm, Eu, Gd, Tb, Ho, Yb and Lu (Table 7a and Table 7b). Refer to Denechaud et al (1970) for a complete description of the INAA method. Following the powdering process as described in section 4.2, samples were sent to the Massachusetts Institute of Technology, in Cambridge, Massachusetts and irradiated at the Nuclear Reactor Laboratory under the Department of Energy Reactor Sharing Program. To ensure data accuracy and precision, a U.S. Geological Survey standard (RGM-1) was employed. Data quality analyses are included in Appendix C.

5.0 **RESULTS**

Comparisons were made between the Geosols and the composite paleosols with the assumed parent materials (African dust and volcanic samples) to characterize their general geochemistry and degree of alteration. In addition, the geochemistry of the submarine clays was analyzed to determine if the clays could be considered paleosols and to see what clues they provide about their origin, the chemical weathering they experienced, and the relative development of the above-sea level paleosols. As noted above, the Bermuda paleosols include Geosols (soils situated between eolianites of sequential age and deposited during a single glacial-interglacial cycle) and composite soils (soils situated between eolianites of non-sequential age and deposited during more than one glacial-interglacial cycle). In contrast, submarine clays were deposited during lower-sea level conditions, may have been influenced by the basement volcanics, and may have experienced somewhat different weathering conditions.

5.1 MINERAL COMPOSITION

Based on comparisons with published d-spacing data for common minerals, the tentatively identified minerals in the paleosols and submarine clay samples from the XRD analyses included quartz, illite, chlorite, gibbsite and possibly woodhouseite, kaolinite and hydroxy-interlayered clays (Table 6). In all of the samples analyzed, there appeared to be a significant amount of amorphous material based on the x-ray diffractograms (Appendix B). Calcite was more strongly associated with the Geosol samples while quartz was more strongly associated with the submarine clay samples. This result can be explained by the lack of calcite precipitation within the submarine clays. No other obvious differences were noted between the paleosol and submarine clay mineralogy.

Previous XRD analysis of the insoluble soil residues by Ruhe et al. (1961) identified vermiculite, kaolinite and gibbsite. Analyses by Muhs et al (2012) identified quartz, woodhouseite or crandallite, and interpreted the dominant clay mineral to be hydroxy-interlayered clays which ranged from smectite/vermiculite to chlorite, boehmite, and goethite. These results indicate a high degree of pedogenesis but not a specific source. By comparison, the Primary Red Clay is described as including primary minerals associated with mafic igneous rocks including plagioclase feldspar, chromite, magnetite and zircon, while the shallow volcanics also included similar igneous minerals as well as olivine, mica/illite and chlorite. Volcanic sands and mineral fragments analyzed by Prognon et al (2011) also identified chromium and titanium rich minerals.

5.2 ORGANIC CARBON AND WATER

Paleosol major and trace element concentrations vary widely due in part to the presence of organic carbon, interstitial water and OH⁻ molecules within clays, and variable carbonate contributions, which occur as primary biolithic grains or as secondary calcite cement. XRFmeasured CaO ranged from 3.02% to 53.15% with an average of 31.23%. Loss on ignition analyses (LOI; 1,100°C) ranged from -17.3% to -44.0% with an average of -33.7% and correlate well with CaO concentrations ($R^2 = 0.97$), indicating the majority LOI is due to loss of CO₂ from calcite (Table 7a; Figure 8). Due to the partially lithified nature of the paleosols with secondary calcite, it has been assumed that LOI due to organic carbon and interstitial water is limited.



correlations with major elements indicate CaO is a diluting agent within the paleosols. Positive correlation with LOI indicates that the majority of LOI is due to CO₂ from calcite and H₂O from interstitial water. Figure 8. Graphs showing correlations of CaO with SiO₂, Fe₂O₃, Al₂O₃ and LOI in the Bermuda paleosols. Negative

Raw CaO data display strong negative correlations with the major elements SiO₂ $(R^2=0.86)$, Al₂O₃ $(R^2=0.97)$ and Fe₂O₃ $(R^2=0.97)$ (Table 8a; Figure 8). As noted above, the majority of CaO is present as calcite or aragonite (CaCO₃), as primary lithic fragments and secondary calcite cement. As a result: (1) the calcite can be considered a diluting agent, reducing overall major element concentrations as a function of increasing CaO; (2); its presence at varying amounts does not allow for direct comparison of major element concentrations; and (3) the presence of CaO as a diluting agent, results in higher correlations between elements, that can be misleading. Therefore, the geochemical data have been corrected to remove the carbonate components, similar to the method employed by Foos (1991) and a description is included in Appendix D as well as a summary of additional LOI data analysis performed at Boston College. The data are presented in both raw and carbonate-corrected forms. However, all discussion and analyses have been made using carbonate-corrected data only. It is important to note that while the carbonate correction allows for a more direct comparison of individual element concentrations between samples, it does not change the elemental ratios within samples.

5.3 MAJOR ELEMENTS

The major elements SiO₂, Al₂O₃ and Fe₂O₃ are the most abundant in the paleosol samples (Table 3b). The more insoluble Al₂O₃ ranged from 9.7% to 44.6% with an average of 35.1 %, while the more soluble SiO₂ ranged from 20.3% to 42.3% with a mean concentration of 31.3%. Fe₂O₃ is present from 9.4% to 22.8% with an average of 16.4%. Comparison of the major elements with SiO₂ shows moderate negative correlations with Al₂O₃ (R²=0.44), and Fe₂O₃ (R²=0.38) and with P₂O₅ (R²=0.38) indicating the loss of SiO₂ with increased weathering and the formation of gibbsite and iron oxy-hydroxides (in the case of Fe₂O₃ and Al₂O₃) (Table 8b; Figure 8). The majority of SiO₂ within the soils is likely as quartz, which is stable under tropical soils (Foos, 1991). Concentrations of K₂O average approximately 1% for the paleosols and indicate

little if any potassium feldspar or mica are present. A weak negative correlation of P_2O_5 with SiO_2 ($R^2 = 0.38$) is unexplained. The presence of P_2O_5 bearing minerals in the paleosols has been previously noted such as woodhouseite (Muhs et al, 2012) or crandallite (Prognon et al, 2011). Apatite has been observed in the volcanic materials (Pirsson, 1914) and in limestone cave deposits (Hearty and Olson, 2010) and intense weathering of apatite can produce crandallite (Prognon et al, 2011). Muhs et al (2012) interpreted the XRF data for the P_2O_5 bearing mineral to most likely be woodhouseite and noted a more widely distributed source for P_2O_5 was likely from bird guano. Phosphate was also detected in the Primary Red Clay (14%) and was explained by Foreman (1951) as being a later addition to the deposits from an unknown source.

The remaining major elements show weaker correlations and large scatter versus SiO₂. Fe₂O₃ shows a strong positive correlation with Al₂O₃ (R^2 =0.69) possibly indicating association with illite and/or chlorite and more strongly with Fe/Al oxyhydroxides. A weak positive correlation of Fe₂O₃ with TiO₂ (R^2 =0.27) suggests inputs from chromite either as an accessory mineral from the parent or minor inputs from the volcanic platform.

MgO concentrations show a strong negative correlation with both Al_2O_3 (R²=0.69) and Fe_2O_3 (R²=0.73) (Figure 9) and a moderate positive correlation with CaO (R² = 0.46) (Table 8b). The stronger negative correlations with Fe_2O_3 and Al_2O_3 suggests the association of MgO with illite and/or chlorite while the weaker positive correlation with CaO indicates Mg substitution in calcite and aragonite.

Na₂O concentrations in the paleosols show a strong positive correlation with the trace element Cl ($R^2 = 0.79$), indicating of the presence of sea water aerosols within the soils (Figure 9).

UCC normalized spider plot of the major elements for average African dust, average paleosol, shallow volcanics, the volcanic sands and the average submarine clay (Figure 10a),







shows the average African dust is very similar to the upper continental crust, but is somewhat depleted in Na₂O. The average paleosol and shallow volcanics show similar trends with enrichment in TiO₂, Fe₂O₃, MnO, MgO and P₂O₅, as well as depletion in SiO₂, Na₂O and K₂O. However, whereas the average paleosol is enriched in Al₂O₃, the shallow volcanics are depleted. The average submarine clay shows similar trends with enrichment in TiO₂, Al₂O₃, Fe₂O₃ and P₂O₅, and similar depletion in SiO₂, MnO and Na₂O. Like the paleosols, the average submarine clay is also depleted in Na₂O and K₂O. The volcanic sands show a similar trend to the shallow volcanics with enrichment in TiO₂, Fe₂O₃, MnO and P₂O₅, and depletion in SiO₂, Al₂O₃, MgO, Na₂O, and K₂O, but unlike the shallow volcanics the volcanic sands are also depleted in P₂O₅. In addition, the volcanic sand enrichment in TiO is significantly greater than that of the shallow volcanics (10x) and the volcanic sand depletion in SiO₂, Al₂O₃, Na₂O and K₂O are greater than that of the shallow volcanics.



Figure 10a. UCC normalized major element spider plot comparing atmospheric dust (*Scheuvens, 2013*), shallow volcanics (*Muhs et al, 2012*), volcanic sands, average paleosol, and average submarine clay.



Figure 10b. UCC normalized major element spider plot comparing Geosols, protosol, and average submarine clay.



Figure 10c. UCC normalized major element spider plot comparing submarine clays and average paleosol.

The Geosols, protosol and modern surface soil are of similar composition to each other (Figure 10b) generally showing slight depletion in SiO₂, Na₂O and K₂O and enrichment in TiO₂,

 Al_2O_3 , Fe2O3, MnO, MgO and P₂O₅, with the Harbour Road Geosol showing the most TiO₂ and P₂O₅ enrichment and the most MgO and Na₂O depletion.

The submarine clays are of similar composition to each other (Figure 10c) showing depletion in SiO₂, MnO, Na₂O and K₂O but are enriched in TiO₂, Al₂O₃, Fe₂O₃, MgO and P₂O₅. However, the submarine clays differ from the average paleosol in MnO depletion, where the paleosols are enriched. The two core samples collected from Harrington Sound are more similar to the each other than the core sample collected from the Hungry Bay mangroves, with the mangrove sample having somewhat higher Fe₂O₃, MnO, MgO and CaO and somewhat lower K₂O and P₂O₅. These differences may reflect differences in the generally anoxic pore waters found in mangroves (Hatcher et al, 1982) versus that found within the seasonally anoxic, brackish to salt water Harrington Sound (Neumann, 1965). Also of importance may be the somewhat longer exposure time for the shallower mangrove sample (-2.3 to -3 m msl) versus the Harrington Sound cores (-27 m msl), since shallower sample depths would be exposed sooner during falling sea levels and would remain exposed later during rising sea levels.

5.3.1 Soil Profile

A plot of major element oxides along a soil profile was developed with samples collected from the oldest, longest exposed and presumably best develop paleosol, the Castle Harbour Geosol ("MH samples"), in order to understand the weathering throughout the soil column (Figure 11). The Fe₂O₃/Al₂O₃ data show a uniform ratio throughout the soil profile, indicating immobility of both oxides and the complete oxidation of soluble FeO to insoluble Fe₂O₃ from the parent minerals. The profiles for SiO₂/Fe₂O₃ and SiO₂/Al₂O₃ are similar throughout, showing enrichment of SiO₂ in the A horizon with more uniform and depleted ratios at depth. This could be evidence of quartz accumulation within the A horizon and vertical movement of clays to the B horizons (Herwitz and Muhs, 1996). These data suggest all three major element oxides are stable



Figure 11. Elemental ratios across soil profile of Castle Harbour Geosol.

and are likely present as quartz (SiO₂) and oxy-hydroxides (Fe₂O₃ and Al₂O₃). All of the other ratios of the more mobile oxides (MgO/Fe₂O₃, MnO/Fe₂O₃, K₂O/Fe₂O₃) show enrichment in the A to AB horizons and uniform and somewhat depleted ratios at depth.

5.3.2 General Chemical Weathering Trends

The elements Al_2O_3 , Fe_2O_3 and SiO_2 are the three most abundant elements in the paleosols and their behavior during pedo-diagenesis differ due to solubility, with SiO_2 being the most soluble and Al_2O_3 being the least soluble. Comparisons between these elements can show relative changes from the parent materials as a result of weathering. A ternary plot of the Al_2O_3 ,

 Fe_2O_3 and SiO_2 data for the paleosols shows the progressive alteration of the soils from the youngest to the oldest soils, with the loss of SiO_2 relative to Al_2O_3 and Fe_2O_3 , along a distinct linear trend (Figure 12). This trend suggests a uniformity of source materials for the paleosols



Figure 12. Si-Al-Fe ternary diagram showing paleosol and volcanic weathering trends (*Muhs et al, 2012*). The paleosol weathering trend suggests an initial period of rapid chemical weathering and a uniformity of source materials from an atmospheric dust (*Scheuvens, 2013*) UCC source (*Taylor and McLennan, 1985*). The figure also shows the result of chemical insitu weathering (Primary Red Clay; *Foreman, 1951*) and chemical weathering and sorting (volcanic sands).

of different ages and similar, consistent weathering conditions over time. Exceptions to this trend include the Ord Road Geosol samples (325 ka), which plot as somewhat more weathered than the

older Harbour Road Geosol (450 ka). In addition, composite soils samples SG-1 (Q_b/Q_{tl}) and FS-1 (Q_r/Q_{tl}) are shown to be less weathered than would be predicted by this model. These inconsistencies may be a result of the limited sample size for these units, uneven weathering or possibly incorrectly identified or mapped soil positions.

Projecting backward along the paleosols weathering trend (i.e. less-weathered) points to a UCC-like parent. The submarine clay samples plot along the weathering trend of the paleosols and are among the least weathered of the soils, falling between the modern surface soil (NON-99SS, 80 ka) and the youngest Geosol (Shore Hills; 200 ka).

A gap in the weathering trend, between the presumed parent dust and the youngest paleosols, indicates an initial period of rapid weathering (Figure 12), since even the modern soil shows significant weathering. This initial weathering likely occurred during the periods of deposition and subaerial exposure.

The shallow volcanics, volcanic sands, and the Primary Red Clay are also included on Figure 12. The shallow volcanics plot between UCC and the younger paleosols. The Primary Red Clay, presumed to be derived from the shallow volcanics by subaerial weathering, plots in the area of the most weathered paleosols, while the volcanic sands plot well outside the trend of the paleosols. These results appear to show two different weathering processes acting on the shallow basement volcanics including mechanical weathering and sorting in the case of the volcanic sands, and in-situ chemical weathering in the case of the Primary Red Clay. The volcanic sands plot well outside the trend of the soil samples demonstrating significant differences in mineralogy with higher Fe₂O₃ and lower SiO₂. In addition, paleosol sample LD-5 plots between the soils and the volcanics sands, suggesting some volcanic inputs at that location. Sample LD-5 was collected off Ferry Reach Road within the influence of the volcanic fragments observed at Whalebone Bay (Figure 2). These results are consistent with previous investigations and suggest at least some local volcanic influence on the paleosols (Prognon et al, 2011; Muhs et al 2012). However, the Primary Red Clay plots within the range of the oldest and most weathered paleosols indicating a very differ petrology from the volcanic sands (Figure 14). Assuming the volcanic sands and the Primary Red Clay are both derived from a material similar to the shallow volcanics, these differences could be explained by the Primary Red Clay having chemically weathered in-situ and retaining clays and secondary minerals, while the volcanic sands were likely mechanically weathered from the seamount and selectively sorted and transported to the marine carbonates at Whalebone Bay (see Figure 2). If true, this result suggests that the highly altered volcanics could produce a lateritic clay of similar composition to the paleosols via in-situ chemical weathering, at least in terms of major element geochemistry. This result also suggests that the volcanic fragments would appear to be minor accessory inputs to the soils, with limited extent.

In order to further illustrate the relative change in the volcanic sands from a presumed shallow volcanic parent, we applied a simplified geochemical model of major element differences to determine absolute gains or losses (Table 10). This analysis takes advantage of the weight percentage of the major elements summing to ⁺/. 100% by assuming the values are equal to grams/100 gram. Assuming no change in the insoluble elements (TiO₂ or Al₂O₃) from the parent, the major elements can be recalculated to show real gains and losses. The results indicate that the volcanic sands have resulted from the removal of nearly 90% of the original basement volcanics and a loss of over 90% of the majority of the major elements, through extensive mechanical weathering and resorting processes.

Modeling of major element changes from weathering was also conducted on the African dust from an assumed UCC-like parent, and on the average paleosol and average submarine clay from an African dust parent, based on the insolubility of both Al₂O₃ and TiO₂, and the results show good agreement (Table 10). This simplified model suggests a cumulative loss of approximately 32% of the original UCC materials to form a typical African dust. Such modeling also suggests an additional loss of 63% to 74% of the African dust to form an average paleosol, and a loss of 61% to 71% to form the average submarine clay.

5.3.3 Degree of Chemical Alteration by Weathering and Diagenesis

Common weathering indices, such as Weathering Index of Parker (WIP)(Parker, 1970), Chemical Index of Alteration (CIA) (Nesbit, 1979; Nesbit and Young, 1982) or Chemical Index of Weathering (CIW)(Harnois, 1988) among others, which show progressive degrees of weathering and mineralogical maturation, could not be performed due to the lack of information on silica-related CaO or non-carbonate calcium. This study applied the Ruxton Ratio (Ruxton, 1968) or the SiO₂/Al₂O₃ ratio as a measure of the relative degree of paleosol weathering. The Ruxton Ratio relates silica loss to total-element loss and considers alumina and other sesquioxides to be immobile during weathering. During hydrolysis SiO₂ is lost through mineral alteration while Al₂O₃ generally is retained (Herwitz & Muhs, 1995), therefore the SiO₂/Al₂O₃ values usually decrease as chemical weathering proceeds. The soil profile shown on Figure 11 of Fe₂O₃/Al₂O₃ vs soil horizon depth shows that Fe₂O₃ and Al₂O₃ are both stable and generally immobile in the Bermuda soils. Therefore, both SiO₂/Al₂O₃ and SiO₂/Fe₂O₃ ratios should be indicators of silicate mineral alterations by chemical weathering. Average SiO₂/Al₂O₃ (2.66) and SiO₂/Fe₂O₃ (5.56) ratios for African dust (Bergametti et al, 1989; Chiapello, 1997) were used as a baseline for the degree of chemical alteration (Muhs, personnel communication).

A review of SiO₂/Al₂O₃ and SiO₂/Fe₂O₃ values for paleosol samples of different ages reveals a general pattern of SiO₂ decreasing with the age of the soils and is consistent for nearly all the samples (Figure 15). The oldest and longest exposed paleosol sample (Q-10, >880ka) has the lowest SiO₂/Al₂O₃ value (0.48), while the youngest paleosol sample (NON-99S, 80ka) has the highest SiO₂/Al₂O₃ value (1.59). In addition, all of the soils resting upon the oldest carbonate unit







Figure 13a & 13b. Graphs of Si/Al and Si/Fe ratios vs Soil Age for the Bermuda Geosols.

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(Walsingham Formation, >880 ka) have SiO₂/Al₂O₃ values of less than 0.62 (not including sample LD-5 which appears to be influenced by volcanic inputs).

The submarine clays appear to be somewhat less weathered than the paleosols of similar age based on SiO₂/Al₂O₃ values (Figure 13). The SiO₂/Al₂O₃ values calculated for the submarine clay samples (1.24 to 1.49, \overline{X} =1.35) fall between the range of values for the younger, less weathered paleosols such as the 85 ka modern surface soil (1.59) and the 200 ka Shore Hill Geosol (1.17 to 1.35; \overline{X} =1.26). The SiO₂/Al₂O₃ value for the Primary Red clay was 0.55 and is similar to the oldest and most weathered paleosols.

The weathering gap noted between the unweathered parent dust and the youngest and least-weathered paleosols on Figure 12, is also present on Figure 13 and is characterized by an initially rapid loss of SiO₂ over the first 80ka – 120ka, followed by a more gradual loss for the older soils. This trend is interpreted to result from the instability of parent materials (plagioclase, mica/illite, etc.) under Bermuda climatic conditions and rapid chemical alteration to more stable minerals (kaolinite, chlorite, and hydroxy-interlayered clays). Muhs (2001) identified a similar degree of rapid weathering in soils on Barbados and noted that the greatest rate of change in SiO₂/Al₂O₃ occurred in the first 100 ka of exposure. The more rapid loss of SiO₂ is likely to have occurred during the period of subaerial exposure, while the more gradual alteration may have occurred after carbonate burial by infiltrating waters, or may simply reflect the decreased rate of weathering over time due to increasing mineral stability.

The SiO₂/Al₂O₃ and SiO₂/Fe₂O₃ values can also be considered a rough indicator of sample mineralogy. For example, higher SiO₂/Al₂O₃ ratios indicate the presence of minerals that are more parent-like, less-stable and with more complex crystal structures (e.g. feldspars, micas, hydroxy-interlayered clays; smectite or vermiculite) and so less weathered. Samples with lower SiO₂/Al₂O₃ values indicate less parent-like, more stable minerals, with simpler crystal structures

(e.g. quartz, kaolinite, gibbsite, and boehmite) and so more weathered. Similarly, with SiO₂/Fe₂O₃ ratios, higher values indicate the presence of minerals with more complex crystal structures (e.g. illite and chlorite) while lower values the presence of simpler minerals (e.g. hematite and goethite). The relative SiO₂/Al₂O₃ ratios (based on molar abundances after Muhs, 2007) for the minerals smectite, kaolinite and gibbsite are superimposed on Figure 13 and show the change soil in mineralogy over time.

5.3.4 Soil Age vs Soil Exposure

The SiO₂/Al₂O₃ ratios for the paleosols show moderate negative correlations with both total soil age ($R^2 = 0.63$) (based on the age of the underlying carbonate unit) and the duration of subaerial weathering/exposure ($R^2 = 0.46$) (based on the age difference between the underlying and overlying carbonates) and both are significant at the 99% level (Table 7a and 7b). In addition, SiO₂/Fe₂O₃ ratios show a moderate negative correlation with age ($R^2 = 0.59$) and exposure ($R^2 = 0.44$) and these were also significant at the 99% level (Table 7a and 7b). The significant major element ratio correlations with both soil age and soil exposure may be due in part to the covariance of age versus exposure, as the oldest soils have the longest periods of exposure (Table 1).

5.4 TRACE ELEMENTS

The trace elements can be subdivided into groups based on elements that have similar geochemical characteristics. These trace element groups consist of the transition metals (V, Cr, Co, Ni, Cu, Zn), the Large Ion Lithophile Elements which include the Low Field Strength Elements (Rb, Sr and Ba), the High Field Strength Elements (Sc, Y, Zr, Nb, Pb, Th, and U) and the Rare Earth Elements (La, Ce, Nd, Sm, Eu, Gd, Tb, Ho, Yb and Lu). Figure 14a compares the shallow volcanics and volcanic sands with average paleosols and average clays. Figure 14b



Figure 14a., 14b. and 14c. UCC normalized (*Taylor & McLennan, 1985*) trace element spider plots of paleosols, submarine clays, and shallow volcanics (*Muhs et al, 2012*)

compares the Geosols to each other and to the average submarine clay. Figure 14c compares the submarine clays to each other and to the average paleosol.

Due to insufficient sample volumes, the transition elements Ni, Cu, Zn and Ga could not be determined for the samples F-2, PY-3, and RBD-5. In addition, sample NON-99SS prepared for INAA was damaged and no results were obtained.

5.4.1 Transition Elements

The first series of the transition elements are often present in near surface environments as ions with charges of 2+ or 3+ and form a variety of oxides, hydroxides and some carbonates, with varying degrees of solubility.

The transition element data for the average paleosol showed strong enrichment for Cr (>10x) and a generally uniform enrichment (2x - 5x) for the remaining transition metals V, Co, Ni, Cu, and Zn (Figure 14a). The submarine clay samples showed similar enrichment for Cr and the other transition elements but no enrichment for Cu and the average submarine clay was less enriched in transition elements than the average paleosol. The Geosols showed similar transition element trends and degrees of enrichment to each other (Figure 14b). Transition element correlations for the paleosols were limited to Cu vs Co ($R^2 = 0.46$; Table 8b). The shallow volcanics showed enrichment in the transition elements (Cr, Co, Ni; Figure 14a) but the Cr enrichments (3x) was the lowest of all the samples, while the volcanic sands were the most enriched in Cr (>100x).

The submarine clay samples showed similar transition element trends relative to each other and Cr enrichment (10x) similar to the average paleosol (Figure 14c). In general, the submarine clays were less enriched in transition elements than the paleosols or, in some cases, slightly depleted (Co, Cu, Zn; Figure 14c).

5.4.2 Low Field Strength Elements

The Low Field Strength Elements (LFSE: Rb, Sr, Cs and Ba) are considered incompatible elements in the lower crust and mantle and are generally incorporated into the upper crust by substituting for elements of similar ionic radius such as Ca and K in feldspars and phyllosilicates, and under surficial weathering conditions form soluble cations that are often adsorbed by clays (Birkeland, 1999). In addition, Sr can substitute for Ca and Mg in carbonates, while Rb is associated with K in feldspars and micas.

The LFSE Rb and Sr are strongly correlated with each other ($R^2=0.96$) while Ba is moderately correlated with Rb ($R^2=0.38$) and Sr ($R^2=0.34$). Rb and Sr also show moderate correlations the transition metals Co and Cu and with CaO. The high concentration of Rb and Sr inputs are likely related to carbonate mixing where they substitute for Ca in aragonite and calcite. Rubidium shows a moderate correlation with K₂O ($R^2=0.52$) most likely in chlorite. Cs is also associated with K₂O ($R^2=0.38$) in K-spar and may be present in daughter products (Table 8b).

The trace element spider plot Figure 14a shows the shallow volcanics and volcanic sands to be depleted in Rb and Cs while the average paleosol and submarine clays are slightly enriched. The volcanics, average paleosol and average submarine clay are all enriched in Sr and Ba, with the volcanics the least enriched in Sr (<10x) and the paleosols the most enriched (>10x). This is likely due to Sr substitution in calcite in the soils.

The LFSE are all enriched in, and show similar trends between the Geosols (Figure 14b). Sr enrichment is highest in the protosol (>100x) due to high calcite content. The submarine clays exhibit similar Sr, Ba and Cs enrichment as the paleosols (Figure 14c), however, only one of the submarine clays shows enrichment in Rb (HSCS/1) and one shows depletion (HUN-1). The slight enrichment in Rb in the paleosols and the clays and the slight depletion in K₂O suggests the breakdown of K-spar and/or micas due to weathering, the loss of soluble K₂O in solution and the adsorption of the more immobile Rb to secondary clay minerals.

5.4.3 High Field Strength Elements

The High Field Strength Elements (HFSE: Sc, Y, Th, U, Pb, Zr, Hf, Nb, Ta) are not easily incorporated into the silicate minerals, due in part to difficulties with charge imbalances, are typically incorporated into the more refractory accessory minerals, and are generally regarded as being immobile (Rollinson, 1993). Under chemical weathering, as a result of their relatively high ionic potential, the elements in this group tend to be adsorbed on to clays near the point of release from their host mineral nearly quantitatively (Taylor and McLennan, 1985) and their ratios can often provide useful geochemical signatures for provenance studies (Muhs, 2007).

A spider plot of UCC normalized HFSE for the average paleosols, average submarine clays and the volcanics (Figure 14a) shows the paleosols to be generally enriched in HFSE with the exception of Sc and U. The average submarine clay and shallow volcanics are slight enriched in Sc and slightly depleted in U. The volcanic sands are depleted in Pb and enriched in Th and U (>10x), and greatly enriched in Nb and Ta (>100x). The Geosols HFSE trends are similar to each other (Figure 16b) as are the submarine clay samples (Figure 14b). A notable difference between the submarine clay samples and the paleosols was the slight enrichment in Sc in the clays and slight depletion of Sc in the paleosols (Figure 14c).

5.4.4 Rare Element Elements

The Rare Earth Elements (REE: La, Ce, Nd, Sm, Eu, Gd, Tb, Ho, Yb and Lu) form a group of elements with chemical properties that change systematically and gradually across the series (Sholkovitz et a, 1993). This unique property provides a method of detecting the relative effects of fractionation and differentiation due to weathering, often resulting in a distinct geochemical signature. A UCC normalized spider plot of the REE data is presented for comparison with a presumed African dust-like parent (Figure 15a). In addition, a chondrite

normalized spider plot is presented to allow for comparisons with previous investigations (Muhs et al, 2012; Figure 15b).



Figure 15a & 15b. UCC and chondrite normalized (*Taylor and McLennan, 1985*) REE spider plots of paleosols, submarine clays, volcanic sands, atmospheric dust (*Scheuvens et al, 2013*) and shallow volcanics (*Muhs et al, 2012*).

Due to a limited number of samples available, limited sample volumes and the ultimate failure of the laboratory equipment used for these analyses, the REE data are incomplete. In addition, some of the individual element results contained large errors relative to the known standard, specifically Ce, Gd and Ho. However, the majority of the data are acceptable to identify general REE trends. At least one representative sample was analyzed for each of the named paleosols and at least three samples were analyzed from the two oldest paleosols (Castle Harbor and Harbor Road). REE concentrations were also determined for the three submarine clay samples and for the insoluble fraction of the volcanic sands collected from Whalebone Bay (Table 7a and Table 7b; Figure 2).

Chondrite normalized plots of REE abundances for the Bermuda Geosols (Figure 15b) show light REE enrichment relative to heavy REE with apparent negative Eu anomalies and relatively flat heavy REE trends. This trend follows the overall trend for African dust (Muhs, 2007) and shows significant enrichment for all paleosols REE (>100x chondrite). By comparison, the African dust REE enrichment is generally <100x chondrite for the light REE and generally <20x chondrite for the heavy REE.

The shallow volcanics and volcanic sands also show light REE enrichment vs heavy REE, but the range across the suite (La to Lu) is much steeper, with no obvious anomalies in the shallow volcanics and a slight negative Eu anomaly in the volcanic sands. However, the shallow volcanics are only slightly more REE enriched than the African dust, while the volcanic sands are the most enriched off all the samples, indicating a high degree of weathering.

The paleosol samples also show negative Ce anomalies, which was also noted by Muhs et al (2012). The submarine clay trends are similar to the paleosol results showing light REE enrichment versus heavy REE, Eu and Ce anomalies, and are at the lower end in terms of absolute abundances, suggesting a lesser degree of weathering. The REE trend for paleosol sample LD-5, which also showed anomalous major and trace element chemistry, appears to be more similar to the volcanic sands, which show significantly more enriched light REE relative to heavy REE and a negative Eu anomaly. It is important to note that previously published results for volcanic fragments at Whalebone Bay (Muhs et al, 2012) did not show an Eu anomaly. However, those results were from a sample described as "volcanic-grain rich beach sand" which likely contained some carbonate component, whereas the sample analyzed in this study contained only the non-carbonate residual mineral fragments.

Upper Continental Crust (UCC) normalized REE spider plots of the Bermuda Geosols, the average submarine clay, average African dust and the volcanics (Figure 15a) show the paleosols and submarine clays to follow a similar trend of heavy REE enrichment relative to light REE. Conversely, the shallow volcanics, the volcanic sands, and paleosol sample LD-5 show light REE enrichment vs heavy REE.

The UCC normalized paleosols show negative Ce and positive Eu anomalies and the older and longer exposed soils are generally the most enriched. The UCC normalized submarine clays have similar REE trends to the paleosols and show the least absolute abundances, suggesting a less weathered condition.

The UCC normalized REE trend for the volcanic fragments shows the very different nature of this sample to the paleosols. These results indicate the significant effect this material would have on the paleosol geochemistry, as shown by paleosol sample LD-5, and suggests that a significant contribution to the paleosols and submarine clays from these highly weathered fragments is unlikely. However, this does not rule out any volcanic contribution to the paleosols.

5.6 SUBMARINE CLAY PROVENANCE

The major element weathering trends, trace element signatures, and the REE signatures for the submarine clay samples are similar to the paleosols and indicate that they are derived from a similar parent material. While the similar weathering trend of the submarine clay samples and the spider plots point to a related upper continental crust parent for the submarine clays, they do not indicate a specific source. Trace element ratios are commonly used as a fingerprinting technique to determine provenance of rocks and sediments (Taylor and McLennan, 1985) and have been used with great effect to identify African dust as a significant source of the Bermuda paleosols (Herwitz and Muhs, 1995; Herwitz et al, 1996). Muhs et al (2012) used trace element ratios of Nd-Cr-Ta and La-Sc-Th as well as Eu/Eu* vs Gd_N/Yb_N and Eu/Eu* vs La_N/Yb_N to differentiate geochemical signatures for the potential sources of the Bermuda paleosols. Potential source materials include African dust, North American loess, and contributions from two different suites of basement rocks from the submerged volcanic seamount.

The trace element results for the submarine clay samples were plotted versus the results presented by Muhs et al (2012) and show that that the clay samples fall within or very near to the African dust fields and/or within the fields for the paleosols, with some spread (Figure 16). Submarine clay sample HS-SP 04 plots somewhat nearer to the deeper volcanics on the Eu/Eu* vs Gd_N/Yb_N and Eu/Eu* vs La_N/Yb_N scatter plots. However, all the submarine clay samples plot closer to the shallow volcanic samples on the Nd-Cr-Ta and La-Sc-Th ternary diagrams, suggesting a potential for some shallow volcanic contribution.

5.7 STRATIGRAPHIC INTERPRETATION OF SUBMARINE CLAYS

Two of the submarine clay samples were obtained from Harrington Sound and one from Hungry Bay (Figure 2). Hungry Bay is a small coastal feature, likely formed in an interdune depression that was deepened and enlarged by dissolution during successive glacial-interglacial periods (M. Rowe, personal communication; after the process described by Vacher, 1978; Vacher and Mylroie, 1991). Submarine clay sample HUN-1 was obtained from the mangrove portion of Hungry Bay (Figure 2).



Figure 16a, 16b, 16c, & 16d. – Trace element plots for submarine clay provenance determination. Dark blue stars represent Harrington Sound samples. Light blue star represents Hungry Bay sample (*after Muhs et al, 2012*).

Harrington Sound is a larger enclosed embayment. Land et al (1967) described Harrington Sound as an intermediate area boxed by dunes that was probably deepened by solution because it was already low. However, they did not believe that Harrington Sound was solely a result of solution and was not considered a karst feature. A depositional sequence for the Bermuda carbonate formation described by Hearty and Vacher (1994) suggests the progressive formation of the eolianite units around the Harrington depression over time. Seismic surveys of Harrington Sound (Gees and Medioli, 1970) suggested the basin morphology might be related to the underlying volcanic bedrock surface, possibly revealing the presence of a submerged caldera. However, considering that Bermuda is a seamount and that the peaks around the Bermuda Rise reached elevations of 1,000 m, it is likely that any trace of the calderas would have been removed by erosion, and that the volcanic basement rocks near Harrington Sound represent lavas that erupted below present day sea level (M. Rowe, personal communication).

Sample HS-CS/1 was obtained from a portion of Harrington Sound known as Patton's Basin while sample HS-SP was obtained from Harrington Sound's South Basin (Figure 17). The core samples collected from Harrington Sound are interpreted as being of Lower Town Hill (450 ka) to Upper Town Hill age (325 ka), while the age of the Hungry Bay sample is interpreted as being of Belmont Formation age (200 ka). This interpretation is based on our assessment of the age of the underlying carbonate units, which is the convention applied to the paleosols located above sea level. Since the geochemistry of the paleosols has been shown to correlate with the age of the underlying carbonates, it is reasonable to assume that the ages of the submarine clays are similarly related to the ages of the carbonates upon which they were deposited. The samples obtained from Harrington Sound are located at the Pleistocene-Holocene sediment boundary within the cores (Vollbrecht, 1996). Carbonate outcrops in and around the sample locations consist of Walsingham Formation (>880 ka) to the northeast, Lower Town Hill Formation (450 ka) at Trunk Island and in outcrops to the east and southwest, and Upper Town Hill Formation (325 ka) in outcrops to the east and north (Figure 3). Outcrops of Belmont Formation (200 ka) and Rocky Bay Formation (125 ka) are present to the northwest, but these units onlap shoreward and are underlain by the older units. The Vollbrecht (1990) interpretation of Harrington Sound seismic line HS 1.2.1 (Figure 18) designates the carbonate units in the vicinity to range from



Figure 17. – Bathymetric map of Harrington Sound (figure from Vollbrecht, 1996).

isotope stage 9 to 15 (325 ka to >500 ka) which are within the range of the outcrops in and around Harrington Sound. By superposition, this suggests the submarine clay samples are likely of Lower Town Hill (450 ka) to Upper Town Hill age (325 ka). Given the depth at which the sample were obtained (-32 m) and the onlapping nature of the carbonate units around Harrington Sound, we interpret the clay samples to most likely be of Lower Town Hill age (450 ka) and likely represent composite soils (i.e. multiple glacial periods) of mid to late Pleistocene glacial deposition.



The bedrock outcrops around Hungry Bay consist entirely of Belmont Formation (200 ka), which onlaps Upper Town Hill Formation (325 ka) (Figure 3). This stratigraphic orientation suggests the core sample obtained from Hungry Bay is probably no younger than 200 ka and also is likely a composite soil of mostly Pleistocene glacial deposition. However, due to its elevation (approximately -3 m), this sample may include some late-interglacial Pleistocene deposits as sea level was falling, and possibly some early Holocene inputs as sea level was rising at the end of the last glacial.

As noted above, the Hungry Bay and Harrington Sound clays appear to be of different ages. However, due to their relative elevation differences the submarine clays have likely experienced similar periods of subaerial exposure. Subaerial weathering of all three submarine clays would have taken place during lower-sea level periods only. Estimates of subaerial exposure durations for the submarine clays were made using the eustatic sea level curve presented by Rowe and Bristow (2015) and show that the Harrington Sound cores would have a cumulative subaerial weathering duration of up to 188 ka to 225 ka while the Hungry Bay sample period of weathering would range from 175 ka to 200 ka, depending on the age of the underlying carbonate units (Figure 19). It is possible that the submarine clays were deposited after both basins were lowered by dissolution and rest on an unconformity, which would make the units significantly younger than the underlying carbonates.



Figure 19. – Determination of subaerial exposure of the submarine clay samples based on Bermuda sea-level curve (*figure from Rowe & Bristow*, 2015).

6.0 DISCUSSION

6.1 WEATHERING FACTORS

In order to understand the weathering processes responsible for the development of the Bermuda paleosols, is it helpful to characterize the climate on Bermuda during their formation, including temperature, precipitation and prevailing wind directions. Of these parameters, the carbonates provide the best indicator of average temperature while the paleosols are more likely to contain a record of precipitation and prevailing wind directions, through weathering indicators and source area determination, respectively.

The Bermuda Islands are situated near the northern limit of subtropical climate conditions due in part to warming by the Gulf Stream (Vollbrecht, 1990) and the current Bermuda climate is classified as humid sub-tropical (Kottek et al, 2006). This geographic position supports a number of temperature sensitive species including hermatypic corals (Vacher and Rowe, 1997), and even minor sea surface temperature decreases or a shift in the Gulf Stream during glacial climate conditions would have dramatic effects at Bermuda. The ages of the Bermuda carbonate materials correspond to warm interglacials. Nearshore accumulations of carbonate sediments accumulated off-shore during high-sea level conditions. These materials were transported to the shorelines to form prograding beach deposits as a result of an initial sea level rise at the onset of glacial conditions, or were blown into dunes during sea level regressions by a few large storms. The eolianite deposits show the winds blew from all directions around the perimeter of the islands and so do not record a prevailing wind direction.

No glacial age coral deposits have been identified to date (Hearty and Olson, 2010; Rowe and Bristow, 2015). It is possible that Pleistocene temperatures at Bermuda were warm enough to support coral growth during glacial low-sea level conditions, but the evidence is not preserved above present-day sea level. Glacial sea levels at times reached depths of over -100 m (Figure 19), which would have placed the shore line below to top of the volcanic platform and these deeper waters could have limited coral growth. However, these extreme events were not the norm and glacial sea levels were near or above the platform surface (-75 m) for the majority of the past 900 ka. It is also possible that the lack of glacial age carbonates was a result of cooler glacial temperatures which terminated coral growth altogether. Wilkenstern et al (2017) note that Bermuda Rise sediments show evidence that glacial meltwaters traveled as far south as Bermuda

during Heinrich Events bringing cooler waters to the Sargasso Sea near Bermuda. Summer ocean temperatures during the last glacial period have been estimated at around 24 °C (Climap, 1976). Yung et al (1996) have suggested that the hydrologic cycle during the last glacial maxima was reduced by half and that tropical sea surface temperatures at Bermuda were approximately 5°C cooler than during the current interglacial. Such a temperature change would prohibit coral growth. However, according to Land et al (1967), the fossil land snail Poecilozonites nelsoni, found within the paleosol and carbonate units, prefers heavily vegetated moist soils. Some of the Bermuda *Poecilozonites* land snails are believed to have spanned multiple glacial-interglacial periods (Hearty and Olson, 2010). According to Outerbridge and Sarkis (2018), the Poecilozonites bermudensis and Poecilozonites circumfirmatus, well documented on the island for the past 1.0 Ma, require a temperature range of 20-25 °C and a generally moist vegetated environment. This narrow temperature range provides an excellent constraint on the Bermuda temperature during their time on the islands and indicate that Bermuda has likely remained generally warm and humid during the Quaternary. In addition, submarine excavations around the islands have identified multiple submerged cedar trees, presumably Bermuda Cedar Juniperus bermudiana in growing position at elevations of up to -15 m (Knox, 1940). Juniperus bermudiana is present on the islands today and is considered a subtropical species that grows in a moist forest habitat (Wingate et al, 2011) but also has a frost hardiness limit of -6.6 °C to -1.1 °C (Zone 9), indicating that it is relatively hardy and can withstand colder winter temperatures (Bannister and Neuner, 2001). The presence of Bermuda Cedar at such depths below present-day sea level indicates that the volcanic platform likely supported subtropical forest species during glacial low-sea level conditions and may have temporarily experienced freezing temperatures in winter, which is similar to the climate seen on Bermuda today. Considering that the presence of the Gulf Stream is partially responsible for maintaining the current climate on Bermuda, it is likely that similar ocean circulation patterns existed during low sea level conditions.

6.2 BERMUDA PALEOSOL AND VOLCANIC WEATHERING

The Bermuda paleosols have a complex formation and development history, which includes atmospheric deposition of mineral-rich dust from a currently arid/semi-arid climate to a humid sub-tropical climate by long range transport across the Atlantic Ocean during both glacial and interglacial periods; the physical sorting and chemical weathering of the materials during the 5-7 day transit; extended periods of subaerial chemical weathering during both glacial and interglacial climate conditions; post-depositional inputs from locally derived minerals (carbonates and volcanics); and further diagenetic alteration by infiltrating meteoric waters upon burial by carbonate sediments. All of these factors complicate the history of the soils. However, the geochemical results presented here show that the soils are remarkably similar in composition.

The similarities of the major, trace, and REE geochemical signatures for the Bermuda paleosol samples (Figure 10, Figure 12-15), which were collected from locations well distributed across the island and from a variety of different ages spanning nearly 900 ka, indicate a common parent and development history. In addition, the geochemical signatures for the submarine clay samples collected from Harrington Sound and Hungry Bay were shown to be derived from a similar UCC source, and geochemical fingerprinting indicate that the submarine clays are derived from a similar parent as the paleosols with a strong atmospheric dust signature (Figure 16). These similarities indicate that atmospheric circulation patterns similar to today around Bermuda likely persisted during the Pleistocene. Since the submarine clays were deposited during low sea level conditions only and the paleosols received some dust inputs during both glacial and interglacial periods, the compositional similarities between the units supports prevailing wind patterns similar to today across the tropical North Atlantic Ocean during both glacial and interglacial periods. The information provided from this limited investigation of submarine clay samples suggests significant potential for future research of these types of deposits to assist in understanding conditions on Bermuda during glacial periods.

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The sea level curve presented by Rowe and Bristow (2015) suggests that glacial low-sea level conditions were more common on Bermuda than high-sea level interglacial conditions over the past 900 ka (Figure 19). In addition, evidence of stacked soil profiles (Sayles, 1931), low modern dust-fluxes (Bricker and Prospero, 1969; Prospero and Nees, 1977; Glaccum and Prospero, 1980; Arimoto et al 1992), and the geochemical similarity of the glacial-age submarine clays to the paleosols, all support the hypothesis that the majority of dust deposition occurred during glacial climate conditions.

The weathering gap observed between the atmospheric dust and the paleosols (Figure 12 and Figure 13) indicates that the soil parent materials underwent rapid post-depositional weathering. This interpretation is reasonable when considering the relatively small median grain size of the dust ($\sim 2 \mu m$) and the unstable mineralogy (feldspars and mica) exposed to the humid, sub-tropical climate of today. While a similarly warm climate is not necessary on Bermuda during glacial climate conditions to generate such rapid weathering, comparable precipitation is. Therefore, we consider it likely that the Bermuda climate has remained wet for at least the past 900 ka.

The SiO₂/Al₂O₃ and the SiO₂/Fe₂O₃ ratio graphs (Figure 13), appear to show two different weathering trends. The initial trend is a rapid alteration by loss of SiO₂ as silica from primary minerals during the initial 100 ka period of subaerial chemical weathering, followed by a more gradual weathering trend. The later mode gradual weathering trend is likely due to a combination of slower rates of digenetic alteration and a more stable mineral assemblage, consisting of secondary minerals and quartz.

The degree of weathering exhibited by the submarine clays, which experienced only lowsea levels climate conditions, supports the conclusion that a significant amount of the paleosol soil alteration occurred during subaerial exposure (Figure 12 and Figure 13), while the more gradual weathering trend over time exhibited by the paleosols, may result from long-term diagenesis by infiltrating waters following carbonate burial. The most developed soils are often observed in outcrop where the paleosols occur within hollows. Soil development would be improved at these locations during subaerial exposure due to pooling water. Similarly, these areas would also result in temporary perched groundwater lenses during precipitation events, increasing chemical weathering. The fact that paleosol SiO₂/Al₂O₃ ratios correlate somewhat more strongly with age than with the exposure duration, further supports this interpretation and suggest that continuous paleosol flushing by meteoric waters may have occurred (Table 8a and Table 8b). This signal is surprising when considering the variety of carbonate and paleosol characteristics that should result in preferential pathways for vadose flow such as onlapping dune formations, fault and fracture patterns, high permeability eolianites vs low permeability paleosols, secondary calcite cementation in the soils and the presence of soil bases, etc. All of these factors should combine to limit direct and well distributed flushing of the Bermuda paleosols. It is possible that the time periods involved for paleosol alteration $(10^4-10^5 \text{ years})$ are sufficient for the presumably small amount of meteoric water that would be available to continuously alter the soils after burial in such a complex flow regime. However, the fact that the oldest soils have the lowest SiO₂/Al₂O₃ ratios and that the trend is progressive with increasing age, indicates that diagenesis may have been a continuous process, and this progressive loss of SiO₂ across soils of all ages, is further evidence for a continuously wet Bermuda. The decreasing rate of weathering over time may also result from the progressive burial of the paleosol units over time by younger onlapping dunes, further complicating the vadose flow system.

Some authors have considered the volcanic platform to be the primary source of the Bermuda paleosols (Blackburn and Taylor, 1969; Prognon et al, 2011) while others have considered the volcanics to have only locally contributed to the paleosol development (Muhs et al, 2012). An important aspect required to explain a volcanic source for the paleosols is a transport mechanism from the platform surface to the carbonates. The presence of volcanic grains identified within the paleosols and volcanic grains in sands identified at Whalebone Bay and Coney Island (Figure 2) have been cited as evidence of a volcanic original of Bermuda paleosols by Prognon et al (2011). They offered erosion and sorting of the volcanic materials during low sea levels as a potential explanation. This process involves erosion and reworking of the volcanics by wave action during a lowered sea level at the platform margin, sorting of lighter minerals (quartz, albite or chlorite) from the heavier minerals (magnetite chromite and titanite) to form of lag deposits, and transportion of the lighter minerals inland by wind as the parent for soil formation. We agree that the deposits at Whalebone Bay are such a lag deposit. Based on the mineralogy and depositional environment, we believe the Whalebone Bay volcanics can best be characterized as a beach placer deposit. Evidence offered by Prognon et al (2011) and additional evidence provided here, supports this interpretation. However, we considering their sorting and transport mechanism unlikely to preferentially transport only the lighter minerals inland, especially since the only direct evidence of volcanic materials being transported from the platform surface to higher elevations on the present islands are the heavy volcanic sands. According to their model, these materials should have remained behind on the platform. We also believe that if the volcanics were the primary source for the paleosols, evidence in the form of volcanic sand lenses would be more widely distributed throughout the carbonates, and that immobile and trace element abundances would be more enriched in the soils, especially the HFS and REE.

The major element modeling conducted here suggests an unreasonably high degree of weathering would be required to generate the Bermuda paleosols directly from the shallow volcanic rock. This conclusion is supported by the abundance of primary minerals reported in the Primary Red Clay (Foreman, 1951) and the lack of similar primary minerals in the paleosols.

The geochemical signatures for the paleosols and submarine clays are markedly different from the volcanic sand deposits at Whalebone Bay and even small contributions from these volcanic fragments are apparent (Figure 10, Figure 12, Figure 14-Figure 16). These placer-like deposits consist of heavy, highly refractory minerals with enriched Ti, Cr, Th, U and REE concentrations, which differ greatly from shallow volcanic basement rocks. Therefore, it is reasonable to assume that weathering and transport of these volcanic sands across the seamount during glacial low-sea level periods were limited and that their influence on paleosols formation was not widespread. However, the paleosols and the submarine clays showed similar major and trace element trends to the shallow volcanics, while the submarine clays showed some similarities to the Primary Red Clay, which is itself highly weathered volcanic residue. These results, though limited, suggest that a more detailed geochemical investigation of the Primary Red Clay is warranted to further understand the potential contributions of the volcanic platform to the paleosols and submarine clays.

6.3 COMPARISON WITH OTHER CARIBBEAN TERRA ROSSA SOILS

In order to better understand the development history of the Bermuda paleosols, comparisons were made with other terra rossa soils found throughout the western North Atlantic (Figure 1). These soils have been shown to share the same parent materials (atmospheric dust and volcanics), so mineralogical and chemical weathering differences may be explained in part by age and climate differences including temperature and rainfall. Assuming tropical to sub-tropical climates across the Caribbean during the mid to late Pleistocene, the major factors influencing the soils in this region would be a combination of rainfall and dust flux (Muhs, 2001). In addition, evaporation and rainfall intensity can also be important factors in chemical weathering. According to Foos (1991) repeated wet/dry cycles can accelerate soil weathering. This can occur in areas where potential evapotranspiration exceeds precipitation, and especially where thin soils are underlain by a well-drained substrate, accelerating drying. The Bermuda paleosols and submarine clays show a similar mineralogy and degree of weathering to soils found on the Bahamas (Foos, 1991) and the Florida coast (Muhs et al, 2007) and appear to be somewhat more weathered than terra rossa soils found on Jamaica (Muhs and Budahn, 2009) and Barbados (Muhs et al, 1987; Muhs et al, 1990; Muhs, 2001; Muhs et al, 2012; Figure 20). Given the lower rates of dust flux at these higher latitudes, wet/dry cycles have been used to explain the degree of weathering at Florida and the Bahamas (Foos, 1991). Precipitation on Bermuda is generally consistent throughout the year with an excess rainfall over evapotranspiration of approximately 10 cm (Plummer at el, 1976). However, a soil water deficit does occur July through August (Vacher, 1974) and the highly permeable nature of the underlying limestones, similar to the Bahamas, should lead to wet/dry cycles and increased chemical weathering.

The SiO₂/Al₂O₃ and the SiO₂/Fe₂O₃ ratios for the Bermuda, Bahama and Florida soils all show an initial rapid period of weathering, while the Barbados and Jamaican weathering trends are more gradual and consistent over similar time periods (Figure 20). The differences may be a result of slightly different parent materials, as the Barbados and Jamaican soils contain volcanic ash inputs and appear to have parent materials with higher initial SiO₂ concentrations. However, even with higher SiO₂/Al₂O₃ parents, the oldest Jamaican soil is as weathered as the oldest and most weathered Bermuda paleosols. Soils on Barbados and Jamaica occur on a series of elevated reef terraces that increase in age with increased elevation. While rainfall is higher at the higher elevations, average temperatures and potential evapotranspiration are lower (Muhs, 2001), resulting in a more consistent weathering pattern over time for these soils. While the Barbados soils follow a similar trend to the Jamaican soils, they do not achieve the degree of weathering seen on Bermuda or Jamaica. This difference is explained by an increased dust flux on Barbados, having the effect of refreshing the soils and reducing the observed rate of weathering (Muhs, 2001).



Figure 20a & 20b. Comparison of Bermuda paleosols and submarine cores with soils on Barbados (*Muhs et al, 1987; Muhs et al, 1990; Muhs, 2001; Muhs et al, 2012*), Bahamas (*Foos, 1991*), Jamaica (*Muhs & Budhan, 2009*) and Florida (*Muhs, 2007*).

At all locations, decreasing SiO₂/Al₂O₃ ratios over time reflect increasing mineral stability over time, as the loss of silica results in more refractory minerals with simpler crystal forms. In addition, burial of the terra rossa soils that occur as paleosols (Bermuda and the Bahamas) acts to reduce the rate of chemical change, as diagenesis is generally slower than subaerial weathering. Hence, the rate of chemical change generally decreases with soil age. As a result, while the relative soil weathering rates across the Caribbean differ, given enough time they can achieve similar compositions.

7.0 CONCLUSIONS

- Terra Rossa paleosols of Pleistocene age on Bermuda are believed to result from the long-range transport of atmospheric dust from Saharan North Africa (Bricker and Prospero, 1969; Herwitz and Muhs, 1995; Herwitz et al, 1996; Muhs et al, 2012) and locally contain inputs from the volcanic platform. Geochemical comparison of the paleosols with an upper continental crust-derived parent show the paleosols to be of a similar composition to each other and to have undergone extensive alteration.
- Two submarine clay samples collected from Harrington Sound are likely of Upper Town Hill to Lower Town Hill Formation Age (325 ka - 450 ka) or younger, while a clay sample collected from Hungry Bay is likely of Belmont Formation to Upper Town Hill Formation Age (200 ka – 325 ka) or younger.
- The submarine clay samples are geochemically similar to the paleosols located above sea level and likely represent materials deposited and chemically weathered during low-sea level periods only. Based on a current Pleistocene sea level curve for Bermuda the submarine clays would have experienced up to 200 ka of subaerial exposure during glacial periods and are somewhat less weathered that paleosols of similar age.

- Trace element fingerprinting of the submarine clay samples indicates they are likely derived from the same parent materials as the paleosols and have undergone a similar degree of weathering as paleosols of similar age/exposure, suggesting similar climatic conditions and persistent atmospheric circulation from North Africa to Bermuda during glacial climate conditions.
- Evidence of stacked soil profiles (Sayles, 1931), low modern dust-fluxes (Bricker and Prospero, 1969; Prospero and Nees, 1977; Glaccum and Prospero, 1980; Arimoto at al 1992), and the geochemical similarity of the glacial-age submarine clays to the paleosol described here, suggest the majority of dust deposition on Bermuda likely occurred during glacial climate conditions.
- Plots of SiO₂/Al₂O₃ vs age show that, like on other western Atlantic islands and the Florida coast, the greatest degree of soil development occurs within the first 100,000 years of subaerial exposure.
- Ternary plots of Si-Al-Fe ratios show a distinct linear weathering trend of the paleosols
 from the youngest, least-weathered modern soil to the oldest, most weathered paleosols.
 The submarine clays also plot along this weathering trend indicating a similar parent and
 development processes. In addition, projecting the weathering trend backwards (in the
 direction of the least weathered soils) points to a UCC/African dust-like parent for the
 paleosols and submarine clays.

- The paleosol and submarine clay geochemistry differed significantly from the weathered volcanic sands found at Whalebone Bay based on major, trace and REE spider plots. The volcanic sands were found to be highly enriched relative to the shallow volcanics, the paleosols, and the submarine clays in Ti, Cr, Th, U and the REE. The Si-Al-Fe plot for the volcanic sands and other geochemical data suggest they are the result of a high degree of mechanical weathering and sorting, rather than in-situ chemical weathering. We characterize the volcanic sands as a beach placer deposit of heavy resistant minerals whose island wide contribution to the paleosols and submarine clays is limited.
- Some similarities were noted between the major and trace element geochemistry of the paleosols and the shallow volcanics, as well as the major element trends for the submarine clays and the weathered volcanic residue known as the Primary Red Clay. The similarity of the Primary Red Clay to the older paleosols (as noted in the Si-Al-Fe ternary weathering trends and the SiO₂/Al₂O₃ ratios) suggest that in-situ chemical weathering of the volcanic platform may produce a lateritic soil with similar major element composition to the paleosols or submarine clays, though additional research is warranted to more fully characterize the mineralogy and trace element geochemistry of the Primary Red Clay. The presence of original volcanic minerals in the Primary Red Clay and their absence in the paleosols is problematic for assigning a volcanic source to the paleosols. In addition, a transport mechanism to emplace a suitable mineral assemblage to the carbonates from the volcanic platform to form the paleosols without highly enriched trace elements, is lacking.
- The degree of weathering of the paleosols and submarine clays is extensive and more strongly correlated with the age of the underlying carbonate units than with the duration

of exposure, suggesting the paleosols have undergone both subaerial weathering and diagenesis. Age-correlated diagenesis may indicate that vadose flow through the carbonate units is not exclusively by preferential pathways and has resulted in continuous alteration of the soils. The decrease in paleosol diagenesis over time is likely due to a combination of soil mineral stability over time as well as a reduction in carbonate permeability due to the loss of primary porosity and additional carbonate/soil burial.

8.0 REFERENCES CITED AND RELATED RESEARCH

- Arimoto, R., R. A. Duce, D. L. Savoie, and J. M. Prospero (1992), Trace elements in aerosol particles from Bermuda and Barbados: Concentrations, sources and relationships to aerosol sulfate, J. Atmos. Chem., 14, 439–457, doi:10.1007/BF00115250.
- Bannister P., Neuner G. (2001). "Frost resistance and the distribution of conifers," in Conifer Cold Hardiness eds Bigras F. J., Colombo S. J. (Dordrecht: Kluwer Academic Publisher;), 3–22 10.1007/978-94-015-9650-3 1.
- Bergametti, G., L., da silva Gomes, and E.Remoundaki, (1989), Present Transport and Deposition Patterns of African Dusts to the North-Western Mediterranean. *in* Paleoclimatology and Paleometeorology: Modern and past patterns of global atmospheric transport. NATO ASI SERIES 282, Kluwer Academic Publishers, Editors: M. Leinen and M. Sarnthein. doi: 10.1007/978-94-009-0995-3 9.
- Birkeland, Peter, (1999), Soils and Geomorphology. 3rd ed., Oxford University Press, New York.
- Blackburn, G., and R. M. Taylor (1969), Limestones and red soils of Bermuda, Geol. Soc. Am. Bull., 80, 1525–1598, doi:10.1130/0016-7606(1969)80[1595:LARSOB]2.0.CO;2.
- Blackburn, G., and R. M. Taylor (1970), Limestones and red soils of Bermuda: Reply, Geol. Soc. Am. Bull., 81, 2525–2526, doi:10.1130/0016-7606(1970)81[2525:LARSOB]2.0.CO;2.
- Bretz, J. H. (1960), Bermuda: A partially drowned, late mature, Pleistocene karst, Geol. Soc. Am. Bull., 71, 1729–1754, doi:10.1130/0016-7606 (1960)71[1729:BAPDLM]2.0.CO;2.
- Bricker, O.P., and F. T. Mackenzie (1970), Limestones and red soils of Bermuda: Discussion, Geol. Soc. Am. Bull., 81, 2523–2524, doi:10.1130/0016-7606(1970)81[2523:LARSOB]2.0.CO;2.
- Bricker, O.P., and Prospero, J.M., (1969), Airborne dust on the Bermuda Islands and Barbadose [abs]: Eos (Transactions, American Geophysical Union), v.50, p. 176.
- Chester, R. and Johnson, L. R., (1971), Atmospheric dusts collected off the Atlantic coasts of North Africa and the Iberian Peninsula. Marine Geol., 11: 251-260. doi:10.1016/0025-3227(71)90027-2.
- Chiapello, I, G. Bergametti, B. Chatenet, P. Bousquet, F. Dulac, and E.S. Soares, (1997), Origins of African dust transported over the northeastern tropical Atlantic, Journal of Geophysical Research Atmospheres 102:13701-13709. doi: 10.1029/97JD00259.
- CLIMAP Project Members (1976), The surface of the ice-age earth. Science, 191(4232), 1131-1144. doi: 10.1126/science.191.4232.1131
- Denechaud, E.B., Helmke, P.A. and Haskin, L.A., (1970), Analysis for rare-earth elements by neutron activation and Ge(Li) spectrometry. J. Radional. Chem., 6: 97-113. doi: 10.1007/BF02513904.

- Foos, A., (1991), Aluminous lateritic soils, Eluthera Bahamas: A modern analog to carbonate paleosols. Journal of Sedimentary Petrology 61: 340-348. doi:10.1306/D4267703-2B26-11D7-8648000102C1865D.
- Foreman, Fred (1951), Study of Some Bermuda Rocks. Bulletin of The Geological Society of America. Vol 62, pp 1297-1330. doi:0.1130/0016-7606(1951)62[1297:SOSBR]2.0.CO;2
- Ganor, E., (1991), The composition of clay minerals transported to Israel as indicators of Saharan dust emission, Atmos, Environ., 25A, 2657-2664. doi: 10.1016/0960-1686(91)90195-D.
- Gees, R.A., and Medioli, F., (1970), A continuous seismic survey of the Bermuda Platform, Part I: Castle Harbour. Maritime Sediments, Vol.6, No.1, pp. 21-25. doi:10.4138/1899.
- Glaccum, R. A., (1978), The mineralogy and elemental composition of mineral aerosols over the tropical North Atlantic; The influence of Saharan dust [M.S. thesis]: Miami, Florida, University of Miami, 161 p.
- Glaccum, R. A., and J. M. Prospero (1980), Saharan aerosols over the tropical North Atlanticmineralogy, Mar. Geol., 37, 295–321, doi:10.1016/0025-3227(80)90107-3.
- Harnois, L. (1988) The CIW Index: A New Chemical Index of Weathering. Sedimentary Geology, 55, 319-322. doi: 10.1016/0037-0738(88)90137-6.
- Hartsock, J.K., D.L. Woodrow, and D.B. McKinney, (1995), Fracture systems in northeastern Bermuda. *in* Terrestrial and Shallow Marine Geology of the Bahamas and Bermuda. ed. Allen Curran and Brian White, The Geological Society of America, Special Paper 300. doi:10.1130/0-8137-2300-0.325.
- Hatcher, P.G., B.R.T. Simoneit, F.T. MacKenzie, A.C. Neumann, D.C. Thorstenson, and S.M. Gerchakov, (1982), Organic geochemistry and pore water chemistry of sediments from Mangrove Lake, Bermuda. Organic Geochemistry, 4 (2), 93-112. doi:10.1016/0146-6380(82)90012-2.
- Hearty, P.J., Olson SL, (2010) Geochronology, biostratigraphy, and changing shell morphology in the land snail subgenus Poecilozonites during the Quaternary of Bermuda. Palaeogeography, Palaeoclimatology, Palaeoecology 293:9–29. doi:10.1016/j.palaeo.2010.04.026.
- Hearty, P. J., H. L. Vacher, and R. M. Mitterer (1992), Aminostratigraphy and ages of Pleistocene limestones of Bermuda, Geol. Soc. Am. Bull., 104, 471–480, doi:10.1130/0016-7606(1992)104<0471:AAAOPL>2.3. CO;2.
- Hearty, P.J., and H.L. Vacher, (1994) Quaternary stratigraphy of Bermuda: A high-resolution pre-Sangamonian rock record, Quat. Sci. Rev., 13, 685- 697, 1994. doi: 10.1016/0277-3791(94)90099-X.
- Herwitz, S. R. (1993), Stemflow influences on the formation of solution pipes in Bermuda eolianite, Geomorphology, 6, 253–271, doi:10.1016/0169-555X(93)90050-C.

- Herwitz, S. R., and D. R. Muhs (1995), Bermuda solution pipe soils: A geochemical evaluation of eolian parent materials, Spec. Pap. Geol. Soc. Am., 300, 311–323, doi:10.1130/0-8137-2300-0.311.
- Herwitz, S. R., D. R. Muhs, J. M. Prospero, S. Mahan, and B. Vaughn (1996), Origin of Bermuda's clay-rich Quaternary paleosols and their paleoclimatic significance, J. Geophys. Res., 101, 23,389–23,400, doi:10.1029/96JD02333.
- Husar, R. B., J. M. Prospero, and L. L. Stowe (1997), Characterization of tropospheric aerosols over the oceans with the NOAA advanced very high-resolution radiometer optical thickness operational product, J. Geophys. Res., 102, 16,889–16,909, doi:10.1029/96JD04009.
- Isphording, W.C., (1978), Mineralogical and physical properties of Gulf Coast limestone soils: Transactions Gulf Coast Association Geological Societies, v. 28, p. 201-214. doi:10.1306/C1EA52EC-16C9-11D7-8645000102C1865D
- Jeong, G.Y. and E.P. Achterberg, (2014), Chemistry and mineralogy of clay minerals in Asian and Saharan dusts and the implications for iron supply to the oceans. Atmospheric Chemistry and Physics. 14, 12415–12428, doi:10.5194/acp-14-12415-2014.
- Kottek, M., J. Grieser, C. Beck, B. Rudolf, and F. Rubel, (2006): World Map of the Köppen-Geiger climate classification updated. Meteorol. Z., 15, 259-263. DOI: 10.1127/0941-2948/2006/0130.
- Land, L. S., F. T. MacKenzie, and S. J. Gould (1967), Pleistocene history of Bermuda, Geol. Soc. Am. Bull., 78, 993–1006, doi:10.1130/0016-7606 (1967)78[993:PHOB]2.0.CO;2.
- Mackenzie, F.T., (1964a), Bermuda Pleistocene Eolianites and Paleowinds, Sedimentology, v.3, 52-64. doi:10.1111/j.1365-3091.1964.tb00275.x.
- Mackenzie, F.T., (1964b), Geometry of Bermuda Calcareous Dune Cross-Bedding, Science, v.144, No. 3625, p.1449-1450. doi: 10.1126/science.144.3625.1449.
- Mahowald, N. M., D. R. Muhs, S. Levis, P. J. Rasch, M. Yoshioka, C. S. Zender, and C. Luo (2006), Change in atmospheric mineral aerosols in response to climate: Last glacial period, preindustrial, modern, and dou- bled carbon dioxide climates, J. Geophys. Res., 111, D10202, doi:10.1029/2005JD006653.
- Moore, H. B. and Moore, D.M. (1946), Preglacial History of Bermuda. Bulletin of the Geological Society of America. Vol. 57, pp. 207-222.
- Mylroie John, E.; Carew, James, L.; Vacher-H-L, (1995), Karst development in the Bahamas and Bermuda, in Shallow Marine Geology of the Bahamas and Bermuda: Geol. Soc. America SP-300. doi:10.1130/0-8137-2300-0.251.
- Muhs, D.R., (2001), Evolution of Soils on Quaternary Reef Terraces of Barbados, West Indies. Quaternary Research, 56(1):66-78. doi:10.1006/qres.2001.2237.

- Muhs, D. R., and J. R. Budahn (2009), Geochemical evidence for African dust and volcanic ash inputs to terra rossa soils on carbonate reef terraces, northern Jamaica, West Indies, Quat. Int., 196, 13–35, doi:10.1016/j.quaint.2007.10.026.
- Muhs, D. R., J. Budahn, A. Avila, G. Skipp, J. Freeman, and D. Patterson (2010), Geochemical evidence for African dust additions to Quaternary soils on the island of Mallorca, Spain, Quat. Sci. Rev., 29, 2518–2543, doi:10.1016/j.quascirev.2010.04.013.
- Muhs, D.R., J.R. Budahn, J.M. Prospero, and S. Carey, (2007), Geochemical evidence for African dust inputs to soils of western Atlantic islands: Barbados, the Bahamas, and Florida. Journal of Geophysical Research Atmospheres 112(F2) doi: 10.1029/2005JF000445.
- Muhs, D. R., J. R. Budahn, J. M. Prospero, G. Skipp, and S. R. Herwitz (2012), Soil genesis on the island of Bermuda in the Quaternary: The importance of African dust transport and deposition, J. Geophys. Res., 117, F03025, doi:10.1029/2012JF002366.
- Muhs, D.M., C.A. Bush, K.C. Stewart, TR. Rowland, and R.C. Crittenden, (1990), Geochemical Evidence of Saharan Dust Parent Material for Soils Developed on Quaternary Limestones of Caribbean and Western Atlantic Islands. Quaternary Research 33, 157-177. doi:10.1016/0033-5894(90)90016-E
- Muhs, D.R., R.C. Crittenden, J.N. Rosholt, C.A. Bush, and K.C. Stewart, (1987), Genesis of marine terrace soils, Barbados, West Indies: evidence from mineralogy and geochemistry: Earth Surface Processes and Landforms, v. 12, p. 605-618. doi:10.1002/esp.3290120604.
- Daniel R. Muhs, Kathleen R. Simmons, R. Randall Schumann, Eugene S. Schweig, Mark P. Rowe, (2020), Testing glacial isostatic adjustment models of last-interglacial sea level history in the Bahamas and Bermuda, Quaternary Science Reviews, Volume 233: doi.org/10.1016/j.quascirev.2020.106212.
- Nelson, Richard J. (1840) On the Geology of the Bermudas. Trans. Geolog. Soc. London, 2d ser., v, pp. 103-123, with woodcuts and map, 1837 (1840), based on observations made between 1827 and 1833. doi: 10.1144/transgslb.5.1.103.
- Nesbitt, H.W., (1979), Mobility and fractionation of rare earth elements during weathering of a granodiorite. Nature, 279, p.206-210. doi:10.1038/279206a0.
- Nesbitt, H.W. & Young G.M., (1982), Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. Nature. 299. 715-717. doi:10.1038/299715a0.
- Neumann, A. Conrad, (1965), Processes of Recent Carbonate Sedimentation in Harrington Sound, Bermuda, Bulletin of Marine Science, Volume 15, Number 4, pp. 987-1035(49)
- Officer, C. B., M. Ewing, and P. C. Wuenschel, (1952) Seismic refraction measurements in the Atlantic Ocean, IV, Bermuda, Bermuda Rise, and Nares Basin, Geol. Soc.Am. Bull., 63, 777-808. doi:10.1130/0016-7606(1952)63[777:SRMITA]2.0.CO;2.

- Outerbridge, M. E. and Sarkis, S.C. 2018. Recovery plan for the endemic land snails of Bermuda: Poecilozonites bermudensis and Poecilozonites circumfirmatus. Department of Environment and Natural Resources, Government of Bermuda. 26 pages. doi: 10.13140/RG.2.2.18957.79848.
- Parker, A., (1970). An index of weathering for silicate rocks. Geological Magazine 107, 501-504. doi:10.1017/S0016756800058581.
- Patey, M.D., E.P Achterberg, M.J. Rijkenberg, and R. Pearce, (2015), Aerosol time-series measurements over the tropical Northeast Atlantic Ocean: Dust sources, elemental composition and mineralogy. Marine Chemistry 174, 103–119. doi:10.1016/j.marchem.2015.06.004.
- Patterson, S.H., (1967), Bauxite reserves and potential aluminum resources of the world: U.S. Geological Survey Bulletin 1228, 176 p.
- Pirsson, L.V., (1914), Geology of Bermuda Island: The igneous platform: American Journal of Science, v.38, p. 189-206, 331-334. doi: 10.2475/ajs.s4-38.225.189.
- Plummer, L.N., H.L. Vacher, F.T. Mackenzie, O.P. Broicker, and L.S. Land (1976), Geological Society of America Bulletin, v. 87, p. 1 3 0 1 - 1 3 1 6, Doc. no. 60912. doi:10.1130/0016-7606(1976)87<1301:HOBACH>2.0.CO;2.
- Prognon, F., I. Cojan, P. Kindler, M. Thiry, and M. Demange (2010), Mineralogical evidence for a local volcanic origin of the parent material of Bermuda Quaternary paleosols, Quat. Res., 75, 256–266, doi:10.1016/j.yqres.2010.08.002.
- Prospero, J. (1968). Atmospheric dust studies on Barbados. Bulletin of the American Meteorological Society, 49(6), 645-652. Retrieved March 31, 2020, from www.jstor.org/stable/26251506.
- Prospero, J.P, R.A. Glaccum, and R.T Nees, (1981) Atmospheric transport of soil dust from Africa to South America. Nature 289(5798) doi: 10.1038/289570a0.
- Prospero, J. M., and P. J. Lamb (2003), African droughts and dust transport to the Caribbean: Climate change implications, Science, 302, 1024–1027, doi:10.1126/science.1089915.
- Prospero, J. M., and R. T. Nees (1977), Dust concentration in the atmosphere of the equatorial North Atlantic: Possible relationship to the Sahelian drought, Science, 196, 1196–1198, doi:10.1126/science.196.4295.1196.
- Reynolds, P.R., and Aumento, F.A., (1974), Deep Drill 1972, Potassium-argon dating of the Bermuda drill core: Canad. Jour. Earth Sci., v. 11, p. 1269- 1273. doi: 10.1139/e74-118.
- Rollinson, Hugh, (2014), Using Geochemical Data: evaluation, presentation, interpretation. Routledge, London.
- Rowe, M.P., (2020), website: Bermuda Geology, An Explanation of the Geology of Bermuda. www.bermudageology.com

- Rowe, M.P. and C.S. Bristow, (2015), GSA Bulletin; November/December; v. 127; no. 11/12; p. 1645–1665; doi:10.1130/B31237.1
- Mark P. Rowe, Karine A.I. Wainer, Charlie S. Bristow, Alex L. Thomas (2014), Anomalous MIS 7 sea level recorded on Bermuda, Quaternary Science Reviews, Volume 90, Pages 47-59: doi.org/10.1016/j.quascirev.2014.02.012.
- Ruhe, R. V., J. G. Cady, and R. S. Gomez (1961), Paleosols of Bermuda, Geol. Soc. Am. Bull., 72, 1121–1142, doi:10.1130/0016-7606(1961)72 [1121:POB]2.0.CO;2.Sayles, R. W. (1931), Bermuda during the ice age, Proc. Am. Acad. Arts Sci., 66, 381–467, doi:10.2307/20026356.
- Ruxton, B.P., (1968), Measures of the degree of chemical weathering of rocks. Journal of Geology 76, 518-527. doi:10.1086/627357.
- Sayles, R.W., (1931), Bermuda during the Ice Age: Proceedings of the American Academy of Arts and Sciences, v. LXVI, no. 11, p. 382–467. doi: 10.2307/20026356.
- Scheuvens, D., L. Schutz, K. Kandler, M. Ebert, and S. Weinbruch, (2013), Bulk composition of northern African dust and its source sediments — A compilation. Earth-Science Reviews Volume 116, Pages 170-194. doi: 10.1016/j.earscirev.2012.08.005.
- Sholkovitz, E. R., T. M. Church, and R. Arimoto (1993), Rare earth element composition of precipitation, precipitation particles, and aerosols, J. Geophys. Res., 98, 20,587–20,599, doi:10.1029/93JD01926.
- Strom, R., & Kim, J. (1987). GENESIS OF DIOCTAHEDRAL CHLORITE-LIKE CLAYS IN TERRA ROSSA SOILS IN SOUTH FLORIDA. Florida Scientist, 50(4), 253-263. Retrieved March 31, 2020, from www.jstor.org/stable/24320183.
- Taylor, S. R., and S. M. McLennan (1985), The Continental Crust: Its Composition and Evolution, 312 pp., Blackwell Sci., Oxford, U. K.
- Vacher, H. L., (1971), Late Pleistocene sea-level history: Bermuda evidence [Ph.D. thesis]: Evanston, Illinois, Northwestern University, 147 p.
- Vacher, H. L., (1973), Coastal dunes of Younger Bermuda, in Coates, D. R., ed., Coastal Geomorphology: Binghamton, State University of New York, p. 355–391.
- Vacher, H.L., (1974), Groundwater hydrology of Bermuda: Hamilton, Government of Bermuda, Public Works Dept., 87 p.
- Vacher, H.L., & Hearty, (1989), History of stage 5 sea level in Bermuda: review with new evidence of a breif rise to present sea level during substage 5A. Quaternary Science Reviews 8:159-168. doi: 10.1016/0277-3791(89)90004-8.
- Vacher, H.L., Hearty, P.J., and Rowe, M.P., (1995), Stratigraphy of Bermuda; nomenclature, concepts, and status of multiple systems of classification. *in* Shallow Marine Geology of

the Bahamas and Bermuda: Geol. Soc. America SP-300. doi:10.1130/0-8137-2300-0.271.

- Vacher, H. L., and M. P. Rowe (1997), Geology and hydrology of Bermuda, in Geology and Hydrology of Carbonate Islands, Dev. Sedimentology., vol. 54, edited by H. L. Vacher and T. M. Quinn, pp. 35–90, Elsevier, Amsterdam.
- Verrill, A.E., (1907), The Bermuda Islands; Part IV, Geology and paleontology, and Part V, An account of the coral reefs: Conn. Acad. Arts and Sciences Trans., v. 12, p. 45-348. doi:10.5962/bhl.part.28921.
- Vogt, Peter, R., (1991), Bermuda and Appalachian-Labrador rise: Common non-hotspot processes?, Geology, v. 19, p. 41-44, January 1991. DOI: 10.1130/0091-7613(1991)019<0041:BAALRC>2.3.CO;2.
- Vogt, P.R. & W.-Y. Jung, (2007) Origin of the Bermuda volcanoes and Bermuda Rise: History, observations, models, and puzzles, in Plates, Plumes, and Planetary Processes, G.R. Foulger and D.M. Jurdy (Eds.), Geological Society of America Special Paper 430. doi:10.1130/2007.2430(27).
- Vollbrecht, R., (1990), Marine and meteoric diagenesis of submarine Pleistocene carbonates from the Bermuda carbonate platform: Carbonates and Evaporites, v. 5, p. 13–96, doi:10.1007/BF03174319.
- Vollbrecht, Rudiger, (1996), Postglacial rise of sea level, palaeoclimate and hydrography, recorded in sediments of the Bermuda inshore waters. Ph.D. Dissertation, Institute und Museum fur Geologie Palaontologie, Universitat Gottingen. http://hdl.handle.net/11858/00-1735-0000-0006-B35B-4.
- Wingate, D.B., Adams, R & Gardner, M. 2011. Juniperus bermudiana . The IUCN Red List of Threatened Species 2011: e.T30376A9532928. doi.org/10.2305/IUCN.UK.2011-2.RLTS.T30376A9532928.en.
- Winkelstern, I. Z., M. P. Rowe, K. C. Lohmann, W. F. Defliese, S. V. Petersen, and A. W. Brewer (2017), Meltwater pulse recorded in Last Interglacial mollusk shells from Bermuda, Paleoceanography, 32, 132–145, doi:10.1002/2016PA003014.
- Woolard G.P and M. Ewing, (1939) Structural Geology of the Bermuda Islands, Nature, 5(27). doi: 10.1038/143898a0.
- Yung, Yuk, L., Typhoon Lee, Chung-Ho Wang and Ying-Tzung Shieh, (1996), Dust: A Diagnostic of the Hydrologic Cycle During the Last Glacial Maximum. Science, Vol. 271, Issue 5251, pp. 962-963. doi: 10.1126/science.271.5251.962.

TABLES

	Approx. Age	Approx. Exposure	Correlated MIS
Limestone Units Soils	(ka)	(ka)	δ^{18} O Stage
St. Georges/Surface Soil		80	1 to 5e
Southampton Formation - Qs	80		5a
Protosol		45	5e/5c
Rocky Bay Formation - Qr	125		Late 5e
Shore Hills Geosol		125	6
Belmont Formation - Qb	200		7
Ord Road Geosol		125	8
Upper Town Hill Formation - Qtu	325		9
Harbour Road Geosol		125	10
Lower Town Hill Formation - Qtl	450		11
Castle Harbour Geosol		430	12-26?
Walsingham Formation - Qw	>880		27/35?

Table 1 - Mapped Geologic Units of Bermuda¹

 1 - after Hearty, 2002 and Muhs et al, 2012 and Rowe et al, 2014

	Shallow	Primary	Bermuda	Bermuda	Miami	Barbados	Cape Verde
Mineral	Volcanics ¹	Red Clay ²	Paleosols ¹	Dust ³	Dust ⁴	Dust ⁴	Dust ⁴
Mica/Illite	*				62.2%	64.3%	53.8%
Quartz		*		* * *	15.0%	13.8%	19.6%
Kaolinite			20-50%	*	7.0%	8.3%	6.6%
Calcite	*				6.1%	3.9%	8.2%
Plagioclase	*	*		*	4.5%	4.1%	5.4%
Chlorite	*		20-40%	*	3.9%	4.1%	4.3%
Microcline					1.3%	1.5%	2.2%
Montmorillinite							
Pyrophylite							
Gypsum							
Gibbsite			*				
Goethite							
Hydroxy-interlayed Clays			*				
Chromite		*					
Magnetite	*	*					
Zircon		*					
Garnet		*					
Diopside	*						
Zeolite	*						
Olivine	*						
Muhs et al, 2012	³ - Bricker and	Prospero, 1969					
Louise 1051	4 ~~	() ()	(

Table 3 - Samples analyze	ed for this investigation					
~ .		Stratigraphic	Soil	Soil	Age	Exposure
Sample PALEOSOLS	Location	Position	Туре	Horizon	(ka)	(ka)
F-2	Ferry Road	O_P/O_{T_0}	Composite	Bw	325	200
LD-1	Ferry Road	$O_{\rm P}/O_{\rm Tu}$	Composite	В	325	200
PR-3	Ferry Road	$O_{\rm P}/O_{\rm Tu}$	Composite	Bt1	325	200
Z-10	Mount Zion Church	$O_{\rm P}/O_{\rm Tu}$	Composite	Bt1	325	200
Z-3	Mount Zion Church	O_R/O_{Tu}	Composite	Bt2	325	200
SG-1	Somers Garden	$O_{\rm B}/O_{\rm TI}$	Composite	В	450	250
FS-1	Fort Scaur	$O_{\rm R}/O_{\rm TI}$	Composite	В	450	325
LD-5	Ferry Road	O_{Tu}/O_W	Composite	В	880	555
Q-10	Government Quarry	/Qw	Composite	Bt	880	880
NON-99SS	Non-Such Island	$/Q_{S}$	St. Georges ⁶	В	80	80
SOM-2	Somerset Island	Q_S/Q_R	Protosol	Bt	125	45
SHR-2	Spice Hill Road	Q_R/Q_B	Shore Hills	Bt	200	125
PG-1	Paget Island	Q_R/Q_B	Shore Hills	В	200	124
RBD-5	Rocky Bay	Q_R/Q_B	Shore Hills	Soil Pipe	200	125
ORD-1	Ord Road	$Q_{\rm B}/Q_{\rm Tu}$	Ord Road	Bt1	325	125
GRR-1	Grape Bay Rail Road	$Q_{\rm B}/Q_{\rm Tu}$	Ord Road	Bt2	325	125
GRR-3	Grape Bay Rail Road	$Q_{\rm B}/Q_{\rm Tu}$	Ord Road	Bt1	325	125
PY-3	Paynter's Road	$\mathbf{Q}_{\mathrm{Tu}}/\mathbf{Q}_{\mathrm{Tl}}$	Harbour Road	В	450	125
IQL	Island Quarry	Q_{TI}/Q_W	Castle Harbour	В	880	430
IRE-1	Ireland Island	Q_{TI}/Q_W	Castle Harbour	B1	880	430
MH-5	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	BA	880	430
MH-10	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	AB	880	430
MH-6	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	Bt1	880	430
MH-7	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	Bt2	880	430
MH-8	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	Bt	880	430
MH-9	Marriott Castle Harbour	Q_{TI}/Q_W	Castle Harbour	Bs	880	430
UBMARINE CLAYS						
HSCS/1	Harrington Sound	Core/Q _{Tl}	Composite		450	200
HSSP-01	Harrington Sound	Core/Q _{Tl}	Composite		450	200
HUN-1	Hungy Bay	Core/Q _B	Composite		200	200
VOLCANIC SANDS		0				
WBB/VOL	Whalebone Bay	Q _{Tu}	Volcanics			

 Table 3 - Samples analyzed for this investigation

			Soil	Age 1	Exposure											
Sample	Position	Type	Horizon	(ka)	(ka)	SiO_2	TiO_2	Al_2O_3	${\rm Fe}_2{\rm O}_3$	MnO	MgO	CaO	Na_2O	$\mathbf{K_2O}$	P_2O_5	IOI
Chondrite ¹		Standard				34.20	0.11	2.44	35.88	0.19	23.70	1.22	0.740	0.70	0.28	6.96
Volcanics ²		Reference				37.70	5.02	11.4	13.6	0.23	6.89	13.90	3.00	2.06	1.08	5.31
UCC ³		Reference				65.89	0.50	15.17	4.49	0.07	2.2	4.19	3.89	3.39	0.20	
Average Dust ⁴		Reference				60.14	0.72	12.02	5.82	0.06	2.29	5.24	0.610	1.71	0.24	8.64
Red Clay ⁵		Reference				12.18	7.82	21.96	12.460	0.02	1.74	8.31			14.14	15.03
PALEOSOLS																
F-2	Q_{R}/Q_{Tu}	Composite	Bw	325	200	2.75	0.29	3.00	1.56	0.01	0.51	48.76	0.21	0.07	0.56	41.70
LD-1	$Q_{I\!R}/Q_{T\!u}$	Composite	В	325	200	13.53	0.89	13.31	6.09	0.05	1.28	30.92	0.46	0.29	0.50	32.60
PR-3	$Q_{I\!R}/Q_{T\!u}$	Composite	Bt1	325	200	18.20	1.21	20.64	8.98	0.04	1.50	20.37	0.50	0.55	0.54	27.00
Z-10	$Q_{I\!R}/Q_{T\!u}$	Composite	Bt1	325	200	2.65	0.13	2.38	1.01	0.00	0.72	50.91	0.16	0.05	0.08	42.80
Z-3	$Q_{I\!R}/Q_{T\!u}$	Composite	Bt2	325	200	26.23	1.19	22.79	9.19	0.04	1.63	13.81	0.35	0.30	0.21	24.60
SG-1	Q_{B}/Q_{TI}	Composite	В	450	250	1.50	0.09	0.97	0.55	0.00	0.64	53.15	0.53	0.04	0.06	43.90
FS-1	Q_{R}/Q_{TI}	Composite	В	450	325	2.01	0.08	1.32	0.55	0.00	1.39	51.49	0.29	0.05	0.10	43.70
LD-5	$Q_{Tu} \! / \! Q_W$	Composite	В	880	555	1.75	1.37	0.59	1.03	0.01	0.34	53.14	0.54	0.04	0.40	42.40
Q-10	/Qw	Composite	Bt	880	880	14.70	1.03	30.65	13.45	0.06	0.70	5.28	0.28	0.80	10.75	20.90
SS66-NON	$Q_{\rm s}$	St. Georges ⁶	В	80	80	2.86	0.10	1.80	0.75	0.02	1.44	47.31	0.22	0.07	0.49	44.00
SOM-2	$Q_{\rm S}/Q_{\rm R}$	Protosol	Bt	125	45	1.86	0.10	1.92	0.72	0.01	0.50	51.85	0.15	0.01	0.14	43.60
SHR-2	Q_{R}/Q_{B}	Shore Hills	Bt	200	125	6.15	0.32	4.85	2.33	0.02	0.69	45.80	1.02	0.22	0.09	40.80
PG-1	Q_{R}/Q_{B}	Shore Hills	В	200	124	7.47	0.38	6.38	2.96	0.07	1.30	41.55	0.19	0.22	0.23	38.50
RBD-5	Q_{R}/Q_{B}	Shore Hills	Soil Pipe	200	125	15.97	0.90	11.81	6.85	0.03	2.28	27.77	0.73	0.29	0.29	32.80
ORD-1	Q_{B} / Q_{Tu}	Ord Road	Bt1	325	125	2.55	0.19	3.23	1.42	0.01	0.55	49.10	0.75	0.11	0.43	43.30
GRR-1	Q_{B} / Q_{Tu}	Ord Road	Bt2	325	125	7.81	0.65	11.62	5.22	0.22	1.05	35.75	0.41	0.41	1.01	35.80
GRR-3	$Q_{B}\!/Q_{Tu}$	Ord Road	Bt1	325	125	5.69	0.44	7.72	3.50	0.13	1.08	41.81	0.79	0.92	0.92	38.50
РҮ-3	Q_{Tu} / Q_{Tl}	Harbour Road	В	450	125	24.56	4.01	25.73	12.95	0.19	0.88	3.39	0.12	0.64	7.98	17.30

Table 4a - Raw major element data by XRF

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Table 4a - Raw	major elen	nent data by XRF	ſŦ													
			Soil	Age E:	xposure											
Sample	Position	Type	Horizon	(ka)	(ka)	SiO_2	TiO_2	Al_2O_3	${\rm Fe}_2{\rm O}_3$	MnO	MgO	CaO	Na_2O	$\mathbf{K_2O}$	P_2O_5	LOI
IQL	Q_{TI} / Q_{W}	Castle Harbour	В	880	430	12.91	3.27	22.56	12.62	0.11	1.37	17.81	0.18	0.25	1.88	26.70
IRE-1	Q_{TI}/Q_{W}	Castle Harbour	B1	880	430	4.55	0.48	8.64	3.74	0.03	0.83	42.93	0.75	0.14	0.19	38.90
MH-5	Q_{TI}/Q_{W}	Castle Harbour	BA	880	430	8.45	0.80	13.52	6.06	0.03	0.92	31.22	0.97	0.42	1.71	36.80
MH-10	Q_{TI}/Q_{W}	Castle Harbour	AB	880	430	15.68	1.52	25.91	11.66	0.06	1.39	10.53	1.36	0.77	3.27	27.00
9-HM	Q_{TI}/Q_{W}	Castle Harbour	Bt1	880	430	19.53	1.54	32.09	14.15	0.10	1.21	3.02	0.37	0.83	4.40	21.80
7-HM	Q_{TI}/Q_{W}	Castle Harbour	Bt2	880	430	19.58	1.07	33.05	13.34	0.06	1.15	3.29	0.29	0.91	4.94	21.00
8-HM	Q_{TI}/Q_{W}	Castle Harbour	Bt	880	430	25.57	1.62	26.94	11.92	0.22	1.75	5.85	0.35	1.23	3.31	20.60
6-HM	Q_{TI}/Q_W	Castle Harbour	\mathbf{Bs}	880	430	11.53	0.64	19.26	7.76	0.03	0.85	25.26	0.26	0.55	2.80	30.70
					Min:	1.50	0.08	0.59	0.55	0.00	0.34	3.02	0.12	0.01	0.06	17.30
					Mean:	10.62	0.94	13.56	6.17	0.06	1.07	31.23	0.47	0.39	1.82	33.76
					Max:	26.23	4.01	33.05	14.15	0.22	2.28	53.15	1.36	1.23	10.75	44.00
BMARINE CL.	AYS															
HSCS/1	$Core/Q_{TI}$	Composite		450	200	35.96	1.84	28.92	4.23	0.02	1.90	1.81	1.71	2.02	2.97	17.70
HSSP-01	$Core/Q_{TI}$	Composite		450	200	31.92	1.82	21.49	9.13	0.02	1.29	1.05	1.82	1.18	2.16	27.40
HUN-1	Core/Q _B	Composite		200	200	30.77	1.74	23.12	10.52	0.04	2.75	6.11	1.61	0.49	0.36	20.90
					Mean:	32.88	1.80	24.51	7.96	0.02	1.98	2.99	1.71	1.23	1.83	22.00
OLCANIC SAN	SO															
WBB/VOL	Volcanics	Volcanics				4.92	47.03	1.86	8.19	0.19	0.99	32.39	0.33	0.02	0.13	0.9
	All data	presented in %														
	\int_{1}^{1} - Taylc	or and McLennan	, 1985 - Ta	able 11. I		4 - Sche	uvens, 2	013 - av	verage of	.14 "dus	t" samp	les				
	² - Shall	ow volcanic sam	ole Muhs, .	2012		^ر - "Prii	nary Re	d Clay"	after Fo	reman.	1951					
	- I aylo	or and McLennan	197 - 0861 ,	cI .7 əldı		- after	Hearty,	7007								

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	Non-Metal		Tr	ansition	Metal			LILE/L	ow Field S	trength
Sample	CI	Λ	Cr	Co	Ni^4	Cu	Zn	$\mathbf{R}\mathbf{b}$	\mathbf{Sr}	Ba
Chondrite ¹		85	3975	764	1.65	168	462	3.45	11.9	3.41
Upper Volcanic ²			103	47.5	62.8		108	58.5	1140	1120
ucc ³		60	35	10.0	20.0	25	71	112.0	350	550
PALEOSOLS										
F-2	0.009	32	136	0						317
LD-1	0.883	61	420	6	31	10	64	39	1,977	559
PR-3	0.408	120	731	13	35	13	56	62	1,711	812
Z-10	0.106	46	83	5	9	8	٢	22	2,963	479
Z-3	0.121	66	414	23	29	٢	50	40	554	381
SG-1	0.681	15	44	6	٢	٢	4	27	4,761	133
FS-1	0.104	32	51	6	5	6	5	22	2,865	354
LD-5	0.461	24	164	6	0	11	5	25	3,770	0
Q-10	0.111	136	722	31	24	38	252	127	14,958	1,021
SS66-NON	0.008	26	47	9	0	10	25	25	3,448	176
SOM-2	0.000	9	73	7	1	10	٢	23	3,354	545
SHR-2	2.414	69	76	8	41	6	15	39	3,889	208
PG-1	0.000	29	177	13	17	15	24	38	3,297	420
RBD-5	0.979	68	342	٢						528
ORD-1	1.144	31	107	9	11	6	8	31	3,797	238
GRR-1	0.663	80	209	18	46	24	154	62	3,433	398
GRR-3	1.364	52	158	6	68	21	76	54	4,698	561
PY-3	0.000	114	757	24						877
IQL	0.000	111	1,133	41	87	29	86	41	3,924	450
IRE-1	0.753	40	201	S	18	17	16	30	2,963	336
MH-5	1.354	88	489	19	55	18	82	57	5,119	512

Table 5a - Raw trace element data by XRF

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		Non-Metal		Tra	insition	Metals			LILE/L	ow Field	Strength
Sample		CI	V	\mathbf{Cr}	C0	Ni^4	Cu	$\mathbf{Z}\mathbf{n}$	Rb	\mathbf{Sr}	Ba
Chondrite ¹			85	3975	764	1.65	168	462	3.45	11.9	3.41
Upper Volcanic ²				103	47.5	62.8		108	58.5	1140	1120
UCC ³			60	35	10.0	20.0	25	71	112.0	350	550
MH-10		1.619	151	929	23	101	30	166	91	6,547	989
9-HW		0.217	144	872	24	64	37	202	110	8,319	1,075
MH-7		0.051	137	917	19	58	34	197	124	9,449	1,121
MH-8		0.067	147	534	24	24	30	215	138	7,545	649
0-HM		0.371	82	501	12	22	21	106	71	4,648	717
	Min:	0.000	9	44	0	0	٢	4	22	554	0
	Mean:	0.534	75	396	14	33	18	80	56	4695	533
	Max:	2.414	151	1133	41	101	38	252	138	14958	1121
SUBMARINE CLAYS											
HSCS/1		1.690	135	361	٢	44	18	207	174	8,643	656
HSSP		1.542	169	684	14	48	22	136	76	4,342	601
HUN-1		1.726	120	458	20	35	11	57	37	574	555
	Mean:	1.653	141	501	14	42	17	133	96	4520	604
VOLCANICS											
WBB/VOL		0.175	0	13,288	40	17	0	37	54	1,808	1,297
		1 - REE Dat	a from	Taylor a	nd McL	ennan,]	- 286	Table	11.1 (Fe	as FeO)	
		2 - Muhs et i	al, 201	2							
		3 - Taylor a	nd McI	ennan, 1	985 - T	able 2.	15 (Fe	as Fe(()		

Table 5a - Raw trace element data by XRF

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Table 5a - Raw trace element data by XRF												ĺ
				HFSE					REE		Met	als
Sample	Sc	Υ	Тh	U	Pb	Zr	Nb	La	Ce	Nd	Ga	Sn
Chondrite ¹	8.64	2.25	0.0425	0.0122	3.65	5.54	375	0.367	0.957	0.711	15.2	2.52
Upper Volcanic ²	18.2		10.8	3.01		337		103	193	88.2		
UCC ³	11.0	22	10.7	2.8	20	190	25.0	30	64	26.0	17	5.5
PALEOSOLS												
F-2	0					143		14	49	22		
LD-1	0	270	19	0	39	98	31	132	142	69	12	б
PR-3	0	407	23	0	52	140	41	154	73	38	15	7
Z-10	0	40	5	0	٢	82	б	70	10	4	9	1
Z-3	0	368	23	б	53	142	35	140	164	62	19	٢
SG-1	0	5	5	0	6	123	0	0	34	15	4	7
FS-1	0	14	5	0	11	84	7	19	15	С	9	0
LD-5	0	10	14	0	9	125	88	101	199	92	7	ю
Q-10	0	844	44	0	102	35	28	184	140	72	8	4
SS66-NON	0	52	5	0	15	72	æ	13	ю	3	٢	1
SOM-2	0	28	9	0	6	146	7	70	26	6	4	0
SHR-2	0	99	10	0	18	138	9	27	81	34	9	б
PG-1	0	221	13	0	22	46	14	43	60	41	6	0
RBD-5	0					144		83	70	33		
ORD-1	0	75	6	0	13	87	4	72	56	26	5	4
GRR-1	0	319	16	0	39	62	21	101	63	31	6	0
GRR-3	0	181	13	0	27	94	11	67	45	21	9	б
PY-3	14					969		237	274	126		
IQL	0	240	23	0	46	269	53	130	128	59	12	б
IRE-1	0	106	13	0	25	113	11	17	28	15	٢	0
MH-5	0	272	21	0	39	133	19	85	72	38	6	7
											Page	: 3 of 4

Table 5a - Raw trace element data by	XRF												
					HFSE					REE		Met	als
Sample		Sc	Υ	Тћ	U	Pb	\mathbf{Zr}	Nb	La	Ce	Nd	Ga	\mathbf{Sn}
Chondrite ¹		8.64	2.25	0.0425	0.0122	3.65	5.54	375	0.367	0.957	0.711	15.2	2.52
Upper Volcanic ²		18.2		10.8	3.01		337		103	193	88.2		
UCC ³		11.0	22	10.7	2.8	20	190	25.0	30	64	26.0	17	5.5
MH-10		3	548	37	0	LL	178	39	229	81	45	14	10
9-HW		22	750	44	0	66	139	43	241	85	46	15	8
7-HM		22	928	41	0	76	0	38	229	45	33	14	9
MH-8		10	379	33	0	73	267	33	165	191	88	14	5
6-HM		0	478	25	0	56	10	23	112	49	26	11	7
	Min:	0	5	5	0	9	0	0	0	з	з	2	0
	Mean:	ς	287	19	0	41	138	24	105	85	41	6	ŝ
	Max:	22	928	44	ŝ	102	696	88	241	274	126	19	10
SUBMARINE CLAYS													
HSCS/1		16	195	35	0	68	404	29	138	126	59	19	9
HSSP		23	296	31	0	70	269	45	182	169	80	20	5
HUN-1		8	221	29	4	68	326	39	183	160	71	22	б
	Mean:	16	237	32	1	69	333	38	168	152	70	20	5
VOLCANICS													
WBB/VOL		0	299	378	40	7	1,324	3,503	3,213	5,260	2,483	0	2
		4 - afte	r Heart	v, 2002									
		5 - Coi	<i>idrite</i> N	i value in	%, all othe	er in ppr	и						

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	Non-Metal			Fransitio	n Metals			LILE/	Low Field	Strngth
Sample	G	Λ	Ċ	°C	Ni ⁸	Сп	Zn	Rb	Sr	Ba
Chondrite ¹		85	3975	764	1.65	168	462	3.45	11.9	3.41
Upper Volcanic ²			103	47.5	62.8		108	58.5	1140	1120
Upper Continental Crust ³		60	35	10.0	20.0	25	71	112.0	350	550
PALEOSOLS										
F-2	0.101	358	1,520	0						3,543
LD-1	2.426	168	1,154	25	85	27	176	107	5,432	1,536
PR-3	0.782	230	1,402	25	67	25	107	119	3,281	1,557
Z-10	1.475	640	1,155	70	83	111	76	306	41,217	6,663
Z-3	0.195	160	699	37	47	11	81	65	895	615
SG-1	15.578	343	1,006	206	160	160	91	618	108,907	3,042
FS-1	1.801	554	883	156	87	156	87	381	49,611	6,130
LD-5	7.595	395	2,702	148	0	181	82	412	62,109	0
Q-10	0.153	188	797	43	33	52	348	175	20,654	1,410
SS66-NON	0.103	335	606	77	0	129	323	323	44,490	2,271
SOM-2	0.000	111	1,346	37	18	184	129	424	61,847	10,050
SHR-2	15.390	440	618	51	261	57	96	249	24,794	1,326
PG-1	0.000	151	921	68	88	78	125	198	17,158	2,186
RBD-5	2.502	174	874	18						1,349
ORD-I	12.385	336	1,158	65	119	97	87	336	41,106	2,577
GRR-1	2.335	282	736	63	162	85	542	218	12,090	1,402
GRR-3	6.438	245	746	42	321	66	458	255	22,176	2,648
PY-3	0.000	148	982	31						1,138
TÒI	0.000	201	2,054	74	158	53	156	74	7,114	816
IRE-1	3.891	207	1,039	26	93	88	83	155	15,312	1,736

 ${\bf Table~5b}$ - Carbonate corrected trace element data by XRF

		Non-Metal			Fransitio	n Metals			LILE/]	Low Field	Strngth
Sample		CI	٧	Cr.	Co	Ni^8	Cu	$\mathbf{Z}\mathbf{n}$	$\mathbf{R}\mathbf{b}$	\mathbf{Sr}	Ba
Chondrite ¹			85	3975	764	1.65	168	462	3.45	11.9	3.41
Upper Volcanic ²				103	47.5	62.8		108	58.5	1140	1120
Upper Continental Crust ³			60	35	10.0	20.0	25	71	112.0	350	550
MH-5		0.067	147	534	24	24	30	215	138	7,545	649
MH-10		4.117	268	1,487	58	167	55	249	173	15,567	1,557
9-HW		2.628	245	1,508	37	164	49	269	148	10,626	1,605
7-HM		0.292	194	1,175	32	86	50	272	148	11,211	1,449
MH-8		0.069	184	1,233	26	78	46	265	167	12,701	1,507
6-HM		0.849	188	1, 147	27	50	48	243	163	10,641	1,641
	Min:	0.00	111	534	0	0	11	81	65	895	0
	Mean:	3.12	265	1, 140	56	102	81	199	233	26,369	2,323
	Max:	15.58	640	2,702	206	321	184	542	618	108,907	10,050
SUBMARINE CLAYS											
HSCS/1		2.12	170	454	6	55	23	260	219	10,862	824
HSSP		2.18	239	996	20	68	31	192	107	6,131	849
HUN-1		2.42	168	641	28	49	15	80	52	804	777
	Mean:	2	192	687	19	57	23	177	126	5,932	817
VOLCANICS WBB/VOL		0.18	0	13,288	40	17	0	37	54	1,808	1,297
		1 - REE Dat	a from T	aylor and N	AcLennan,	1985 - Ta	able 11.1	(Fe as Fe	(<u>0</u>		
		2 - Muhs et	al, 2012								
		3 - Taylor a	nd McLe	nnan, 1985	- Table 2	.15 (Fe as	: FeO)				

Table 5b - Carbonate corrected trace element data by XRF

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				HFSE					REE		Me	tals
Sample	Sc	Y	Тћ	U	Pb	Zr	Nb	La	Ce	Nd	Ga	\mathbf{Sn}
Chondrite ¹	8.64	2.25	0.0425	0.012	3.65	5.54	375	0.367	0.957	0.711	15.2	2.52
Upper Volcanic ²	18.2		10.8	3.01		337		103	193	88.2		
Upper Continental Crust ³	11.0	22	10.7	2.8	20	190	25.0	30	64	26.0	17	5.5
PALEOSOLS												
F-2	0					1,598		156	548	246		
LD-1	0	742	52	0	107	269	85	363	390	190	33	8
PR-3	0	780	44	0	100	268	79	295	140	73	29	4
Z-10	0	556	70	0	76	1,141	42	974	139	56	83	14
Z-3	0	594	37	5	86	229	57	226	265	128	31	11
SG-1	0	114	114	0	206	2,814	0	0	778	343	91	46
FS-1	0	242	87	0	190	1,455	35	329	260	52	104	0
LD-5	0	165	231	0	66	2,059	1,450	1,664	3,278	1,516	33	49
Q-10	0	1,165	61	0	141	48	39	254	193	66	11	9
SS66-NON	0	671	65	0	194	929	39	168	39	39	90	13
SOM-2	0	516	111	0	166	2,692	37	1,291	479	166	74	0
SHR-2	0	421	64	0	115	880	38	172	516	217	38	19
PG-1	0	1,150	68	0	114	239	73	224	468	213	47	0
RBD-5	0					368		212	179	84		
ORD-1	0	812	76	0	141	942	43	<i>611</i>	606	281	54	43
GRR-1	0	1,123	56	0	137	278	74	356	222	109	32	0
GRR-3	0	854	61	0	127	444	52	316	212	66	28	14
PY-3	18					903		308	356	163		
IQL	0	435	42	0	83	488	96	236	232	107	22	5
IRE-1	0	548	67	0	129	584	57	88	145	78	36	0

 ${\bf Table~5b}$ - Carbonate corrected trace element data by XRF

					HFSE					REE		Me	tals
Sample		Sc	Y	$\mathbf{T}\mathbf{h}$	Ŋ	Pb	Zr	Nb	La	Ce	Nd	Ga	\mathbf{Sn}
Chondrite ¹		8.64	2.25	0.0425	0.012	3.65	5.54	375	0.367	0.957	0.711	15.2	2.52
Upper Volcanic ²		18.2		10.8	3.01		337		103	193	88.2		
Upper Continental Crust ³		11.0	22	10.7	2.8	20	190	25.0	30	64	26.0	17	5.5
MH-5		10	379	33	0	73	267	33	165	191	88	14	5
MH-10		0	827	64	0	119	404	58	258	219	116	27	9
9-HW		5	889	60	0	125	289	63	372	131	73	23	16
MH-7		30	1,011	59	0	133	187	58	325	115	62	20	11
MH-8		30	1,247	55	0	130	0	51	308	60	44	19	8
6-HW		0	1,094	57	0	128	23	53	256	112	60	25	5
	Min:	0	114	33	0	73	0	0	0	39	39	11	0
	Mean:	4	710	72	0	128	762	113	388	395	181	42	12
	Max:	30	1,247	231	S	206	2,814	1,450	1,664	3,278	1,516	104	49
SUBMARINE CLAYS													
HSCS/1		20	245	44	0	85	508	36	173	158	74	24	8
HSSP		32	418	44	0	66	380	64	257	239	113	28	7
HUN-1		11	310	41	9	95	457	55	256	224	66	31	4
	Mean:	21	324	43	5	93	448	23	229	207	96	28	9
VOLCANICS		c		010	4	t					0 102 0	¢	ć
WBB/V0L		0	299	378	40	L	1,324	3,503	3,213	5,260	2,483	0	2
		4 - afte 5 - 6	r Hearty,	2002	- 11- 70								

 ${\bf Table~5b}$ - Carbonate corrected trace element data by XRF

5 - Condrite Ni value in %, all other in ppm

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Table 6.	Tentative	mineralogy	of selected	Bermuda	soils by]	XRD Analysis						
	d-space	d-space	d-space					d-space	d-space	d-space		
	untreat.	ethyl.gly.	heat	Peak		In terp reted		untreat.	ethyl.gly.	heat	Peak	In terp reted
Sample	Å	Å	Å	(inten.)	Notes	Minerals		Å	Å	Å	(inten.) Notes	Minerals
	3.34			small		quartz		2.59			small	illite
	4.25			small		quartz		2.99			small	
		2.56		mod.		Illite		3.34			small	quartz
		2.93		mod.		woodhouseite		4.25			mod.	quartz
		3.34		large		quartz		4.46			small	illite
		3.49		mod.	broad	chlorite		4.83			small	gibbsite
		4.25		small		quartz	LITIN 1	14.0			small	chlorite
		4.44		small		illite	1-11011		2.81		v.small	
		7.03		small	broad	kaolinite			3.34		mod.	quartz
		(7.69)		v.small					4.25		v.small	quartz
			3.34	v.small		quartz				3.33	large	quartz
			3.52	small		chlorite				3.49	mod.	chlorite
			4.25	small		quartz				4.25	mod.	quartz
			4.46	mod.		illite				4.46	small	illite
	2.93			mod.		woodhouseite		2.67			mod.	goethite
	3.34			mod.		quartz		3.02			v.small	calcite
	4.25			v.small		quartz		3.27			mod.	illite
	4.46			small		illite		3.34			small	quartz
	4.80			mod.	broad	HIC	SHR-2	3.38			large	illite
0-1114		2.93		mod.		woodhouseite		3.83			mod.	calcite
		3.03		large		calcite		14.0			v.small	chlorite
		3.34		mod.		quartz			3.03		mod.	calcite
		4.46		v.small		illite				3.03	mod.	calcite
		4.80		small	broad	HIC						

I able 7a. Kaw major, trace and KEE data by INAA											
Sample	Na_2O^2	CaO^2	Sc	\mathbf{Cr}	FeO^2	Co	Rb	Cs	La	Ce	Nd
Chondrite ¹	0.98	1.89	8.64	3,975	35.8	764	3.45	0.279	0.367	0.957	0.711
Upper Volcanics ³	1.52	9.21	18.2	103	9.33	47.5	58.5	1.36	103	193	88.2
UCC ⁴	3.9	4.2	11	35	4.5	10	112	3.70	30	64	26
Average Dust ⁵	0.55	2.03	15.33	87.01	4.49	18.28	98.4	4.90	46.90	100.24	43.57
PALEOSOLS											
F-2	0.24	40.32	6.12	140.92	1.41	2.66	6.07	1.22	33.46	47.44	32.37
LD-1	0.50	27.09	20.10	452.41	5.34	9.11	20.90	5.30	110.85	140.34	122.54
PR-3	0.31	23.67	19.78	465.08	5.25	8.40	30.94	6.30	96.90	66.89	50.02
Z-10	0.21	41.63	4.24	82.67	0.90	1.33	3.24	0.75	16.87	22.82	17.38
Z-3	0.39	8.69	36.94	480.57	9.17	13.30	24.91	10.98	130.61	206.21	139.10
SG-1	0.57	34.71	2.04	47.13	0.50	1.13	1.80	0.59	8.48	12.07	8.91
FS-1	0.29	37.46	2.86	65.59	0.64	1.09	4.02	0.81	9.92	15.25	10.41
LD-5	0.47	45.93	2.62	178.68	0.93	2.43	0.00	0.25	91.22	189.05	85.46
Q-10	0.08	3.38	47.09	681.28	10.85	23.80	51.88	10.45	188.84	237.03	166.42
SS66-NON											
SOM-2	0.19	44.57	3.47	74.91	0.67	1.11	0.00	0.59	13.72	18.67	4.02
SHR-2	1.00	43.37	7.31	93.54	1.93	3.79	10.85	2.74	36.57	51.58	36.32
PG-1	0.24	38.50	16.10	185.02	2.68	5.31	15.95	3.47	58.72	75.60	40.39
RBD-5	0.81	20.36	18.92	91.74	6.09	7.49	13.61	4.64	91.67	29.39	60.98
ORD-1	0.73	31.99	7.38	106.93	1.35	2.18	5.97	1.53	30.15	36.48	30.72
GRR-1	0.43	31.69	21.29	212.63	4.67	11.87	39.53	8.89	108.91	137.09	115.76
GRR-3	0.82	35.15	13.09	159.80	3.07	8.51	25.68	5.55	73.06	89.09	79.13
PY-3	0.11	2.67	41.93	811.01	11.90	15.90	38.49	16.56	221.63	353.75	43.55
IQL	0.08	8.59	73.96	1,371.43	13.15	41.59	0	6.28	160.69	25.85	47.82
IRE-1	0.67	31.86	11.09	226.32	3.60	4.79	12.04	3.84	44.76	69.96	43.10
MH-5	0.94	27.37	15.75	427.61	4.57	7.59	23.30	4.51	86.24	29.32	18.94
MH-10	1.22	10.39	30.57	831.59	9.07	14.08	43.58	8.83	167.33	11.98	165.26

d R E E data hv INAA - 2 Table 7a D.

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Table 7a. Raw major, trace and REE data by I	INAA											
Sample]	Na_2O^2	CaO^2	Sc	\mathbf{Cr}	FeO^2	Co	Rb	\mathbf{Cs}	La	Ce	Nd
Chondrite ¹		0.98	1.89	8.64	3,975	35.8	764	3.45	0.279	0.367	0.957	0.711
Upper Volcanics ³		1.52	9.21	18.2	103	9.33	47.5	58.5	1.36	103	193	88.2
UCC ⁴		3.9	4.2	11	35	4.5	10	112	3.70	30	64	26
Average Dust ⁵		0.55	2.03	15.33	87.01	4.49	18.28	98.4	4.90	46.90	100.24	43.57
9-HW		0.36	2.61	38.95	844.57	11.74	13.27	67.76	11.05	207.72	268.11	213.01
MH-7		0.00	2.48	42.92	881.67	11.23	11.64	69.05	10.73	209.63	257.15	213.40
MH-8		0.20	3.72	40.14	544.49	10.47	21.28	96.68	12.91	135.12	207.78	128.39
6-HW		0.36	21.94	24.46	57.42	6.26	6.33	36.33	5.83	114.21	105.81	30.66
	Min:	0.53	11.91	35.96	612.81	9.26	15.26	50.26	9.47	154.06	149.46	109.91
N	Mean:	0.53	10.94	37.01	630.19	9.55	15.68	52.10	9.74	158.26	152.55	112.43
	Max:	0.54	9.98	38.06	647.57	9.83	16.11	53.94	10.02	162.46	155.63	114.96
SURMARINE CLANS		0 54	9 01	30 11	664 95	10 11	16 53	55 78	10.29	166.67	158 77	117 48
HSCS/1		0 55	8 05	40.15	687 33	10.39	16.95	57.61	10.57	170.87	161 81	120 01
HSSP-01		0.56	7.08	41.20	699.71	10.67	17.38	59.45	10.85	175.07	164.90	122.54
HUN-1		0.56	6.12	42.25	717.10	10.96	17.80	61.29	11.12	179.28	167.99	125.06
	Mean:	0.57	5.15	43.30	734.48	11.24	18.22	63.13	11.40	183.48	171.08	127.59
VOLCANIC SANDS		0.57	4.19	44.34	751.86	11.52	18.65	64.97	11.67	187.69	174.17	130.11
WBB/VOL		0.58	3.22	45.39	769.24	11.80	19.07	66.80	11.95	191.89	177.26	132.64
I - REE Data from Taylor and McLennan, 198	85 - Tab	le 11.1			4 - Tayloi	· and Mc	Lennan,	1985 - 2	Table 2.	15		
2 - Major data presented in %, all other result	ts in ppi	и			5 - Muhs,	2007 an	d Muhs,	2010				
3 - Muhs et al, 2012					- data in	blue have	e some Q	A errors				

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I able /a. Ivaw IIIajui, Have allu NEE uata UY IIVAN									
Sample	Sm	Eu	Gd	Тb	Ho	Υb	Lu	Ηf	Та
Chondrite ¹	0.231	0.087	0.306	0.058	0.085	0.248	0.038	0.179	0.026
Upper Volcanics ³	14.7	4.09	10.3	1.35	1.29	2.12	0.29	7.89	10.1
UCC ⁴	4.5	0.88	3.8	0.64	0.80	2.20	0.32	5.8	2.2
Average Dust ⁵	9.00	1.81	7.94	1.15	1.52	3.60	0.53	6.03	1.8
PALEOSOLS									
F-2	6.57	1.75	3.83	1.10	1.48	4.72	0.76	1.01	1.04
LD-1	22.02	5.52	12.09	3.55	6.43	16.87	2.56	4.41	2.43
PR-3	2.45	4.74	11.10	3.55	2.36	13.64	1.22	4.13	1.81
Z-10	3.56	0.92	2.27	0.66	0.98	3.09	0.49	0.70	0.22
Z-3	29.78	7.33	19.92	5.48	7.80	25.41	3.69	7.63	2.49
SG-1	1.92	0.48	1.11	0.33	0.70	1.54	0.26	0.39	0.13
FS-1	2.29	0.55	1.44	0.46	0.87	2.12	0.32	0.51	0.19
LD-5	10.45	3.27	7.80	1.27	0.80	1.78	0.11	0.93	10.08
Q-10	39.47	10.52	21.15	6.83	13.10	43.79	7.52	3.93	1.48
SS66-NON									
SOM-2	3.06	0.77	1.86	0.57	1.09	2.78	0.42	0.46	0.10
SHR-2	1.04	2.00	4.74	1.38	2.04	5.04	0.78	1.67	0.53
PG-1	1.21	3.68	6.76	2.65	5.31	13.61	1.84	2.19	0.27
RBD-5	4.51	4.54	12.40	3.73	3.11	13.14	0.00	4.68	0.99
ORD-1	6.88	1.86	2.53	1.25	1.99	5.71	0.82	0.98	0.30
GRR-1	24.11	6.58	10.94	4.37	8.00	21.11	2.91	3.43	1.27
GRR-3	16.31	4.43	7.15	2.85	4.64	12.29	1.84	2.33	0.77
PY-3	40.17	9.50	27.36	7.55	10.65	31.30	4.32	20.20	7.74
IQL	4.37	7.71	15.03	4.44	5.00	5.69	14.44	12.36	2.50
IRE-1	8.61	2.00	6.73	1.59	2.27	7.54	1.25	2.94	1.08
MH-5	17.77	4.52	9.86	0.77	4.81	2.15	0.54	3.70	1.26
MH-10	33.65	8.94	19.31	6.76	11.04	21.10	1.88	7.16	2.54

Table 7a. Raw major, trace and REE data by INAA

I aDIC / a. INAW IIIADI, HAV AILU INTEL HAIA	WWW KN									
Sample		\mathbf{Sm}	Eu	Gd	$\mathbf{T}\mathbf{b}$	Ho	Yb	Lu	Ηf	Ta
Chondrite ¹		0.231	0.087	0.306	0.058	0.085	0.248	0.038	0.179	0.026
Upper Volcanics ³		14.7	4.09	10.3	1.35	1.29	2.12	0.29	7.89	10.1
UCC ⁴		4.5	0.88	3.8	0.64	0.80	2.20	0.32	5.8	2.2
Average Dust ⁵		9.00	1.81	7.94	1.15	1.52	3.60	0.53	6.03	1.8
9-HW		42.29	11.47	23.31	9.07	13.32	27.01	3.64	7.28	2.73
MH-7		42.85	11.65	21.97	9.19	16.99	32.75	4.19	5.52	1.57
MH-8		25.63	6.40	18.29	4.69	8.27	23.91	3.70	8.18	2.75
0-HW		21.85	6.34	12.49	0.76	3.16	17.06	0.00	3.16	1.19
	Min:	28.03	8.10	17.38	5.22	9.19	20.39	3.53	7.35	2.15
	Mean:	28.89	8.33	17.84	5.36	9.47	20.86	3.62	7.57	2.17
	Max:	29.75	8.55	18.29	5.49	9.75	21.32	3.71	7.79	2.18
SUBMARINE CLAYS		30.61	8.78	18.75	5.63	10.03	21.78	3.79	8.01	2.20
HSCS/1		31.47	9.00	19.21	5.77	10.31	22.25	3.88	8.23	2.22
HSSP-01		32.33	9.23	19.67	5.90	10.59	22.71	3.97	8.45	2.24
HUN-1		33.20	9.45	20.13	6.04	10.87	23.17	4.05	8.67	2.26
	Mean:	34.06	9.68	20.58	6.18	11.15	23.64	4.14	8.89	2.28
VOLCANIC SANDS		34.92	9.90	21.04	6.31	11.43	24.10	4.23	9.11	2.30
WBB/VOL		35.78	10.13	21.50	6.45	11.71	24.56	4.31	9.33	2.32
	E 1001									

Table 7a. Raw major. trace and REE data hv INAA

REE Data from Taylor and McLennan, 1985 - Ta
 Major data presented in %, all other results in pj
 Muhs et al, 2012
-	AT 202	∩ ∩2	ζ	Ç	T 22	ζ	2	ζ	٠	ζ
Sample	Na20 ⁻	CaU ⁻	Sc	Cr	FeO ⁻	C0	Кb	CS	La	Ce
Chondrite ¹	0.98	1.89	8.64	3,975	35.8	764	3.45	0.279	0.367	0.957
Upper Volcanics ³	1.52	9.21	18.2	103	9.33	47.5	58.5	1.36	103	193
UCC ⁴	3.9	4.2	11	35	4.5	10	112	3.70	30	64
Average Dust ⁵	0.55	2.03	15.33	87.01	4.49	18.28	98.4	4.90	46.90	100.24
PALEOSOLS										
F-2	2.69	450.68	68.39	1,575.23	15.71	29.68	67.80	13.65	374.00	530.34
LD-1	1.36	74.43	55.22	1,242.93	14.66	25.04	57.41	14.56	304.53	385.57
PR-3	0.59	45.38	37.94	891.76	10.07	16.10	59.32	12.08	185.80	128.27
Z-10	2.93	579.10	58.98	1,150.01	12.52	18.49	45.03	10.39	234.64	317.42
Z-3	0.63	14.04	59.65	776.06	14.80	21.48	40.22	17.73	210.91	333.00
SG-1	13.07	793.97	46.59	1,078.19	11.44	25.84	41.08	13.52	193.97	276.18
FS-1	5.08	648.61	49.54	1,135.77	11.04	18.80	69.61	14.09	171.86	264.12
LD-5	7.76	756.64	43.24	2,943.59	15.24	40.08	0.00	4.11	1,502.76	3,114.42
Q-10	0.11	4.67	65.02	940.71	14.99	32.86	71.64	14.43	260.75	327.29
SS66-NON										
SOM-2	3.50	821.76	63.97	1,381.31	12.26	20.38	0.00	10.82	253.01	344.30
SHR-2	6.39	276.50	46.58	596.33	12.33	24.17	69.16	17.45	233.17	328.85
PG-1	1.24	200.37	83.81	962.89	13.97	27.64	82.99	18.06	305.58	393.45
RBD-5	2.06	52.03	48.36	234.44	15.57	19.15	34.77	11.86	234.25	75.10
ORD-1	7.91	346.36	79.90	1,157.59	14.57	23.55	64.61	16.59	326.42	394.91
GRR-1	1.51	111.60	74.98	748.85	16.43	41.80	139.23	31.31	383.55	482.81
GRR-3	3.89	165.91	61.80	754.32	14.51	40.16	121.21	26.22	344.84	420.51
PY-3	0.14	3.46	54.41	1,052.30	15.44	20.64	49.94	21.49	287.57	459.00
IQL	0.15	15.57	134.08	2,486.38	23.84	75.41	0.00	11.39	291.33	46.87
IRE-1	3.48	164.67	57.33	1,169.56	18.60	24.77	62.23	19.86	231.32	361.53

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Sample	Na2	0^{2}	CaO ²	Sc	Cr	FeO ²	Co	Rb	Cs	La	Ce
Chondrite ¹	0.0	98	1.89	8.64	3,975	35.8	764	3.45	0.279	0.367	0.957
Upper Volcanics ³	1	52	9.21	18.2	103	9.33	47.5	58.5	1.36	103	193
UCC ⁴	Э.	6	4.2	11	35	4.5	10	112	3.70	30	64
Average Dust ⁵	0	55	2.03	15.33	87.01	4.49	18.28	98.4	4.90	46.90	100.24
MH-5	2.5	86	83.24	47.90	1,300.35	13.90	23.09	70.84	13.71	262.25	89.17
MH-10	1.0	97	16.87	49.62	1,349.65	14.73	22.85	70.73	14.32	271.58	19.44
9-HW	0.4	49	3.51	52.49	1, 138.13	15.82	17.89	91.31	14.89	279.93	361.30
2-HM	0.0	00	3.33	57.70	1, 185.14	15.10	15.65	92.81	14.42	281.78	345.66
MH-8	0.	28	5.10	55.06	746.87	14.36	29.19	132.62	17.71	185.34	285.02
6-HM	0.8	83	50.24	56.00	131.45	14.34	14.49	83.17	13.36	261.48	242.25
Mi	in: 0.(00	3.33	37.94	131.45	10.07	14.49	0.00	4.11	171.86	19.44
Mea	in: 2.5	84	227.52	60.34	1, 125.19	14.65	26.77	64.71	15.52	314.90	413.07
Ma	ax: 13.	07 8	821.76	134.08	2,943.59	23.84	75.41	139.23	31.31	1,502.76	3,114.42
SUBMARINE CLAYS											
HSCS/1	5.	11	1.57	35.94	455.07	4.52	11.34	149.25	19.36	147.52	249.90
HSSP-01	5	52	0.61	40.14	1,051.15	12.07	15.52	63.64	13.85	242.50	362.28
I-NUH	5.	43	6.95	38.13	710.97	13.63	15.53	29.18	12.97	187.26	329.72
Mea	in: 2	35	3.04	38.07	739.07	10.07	14.13	80.69	15.39	192.43	313.97
VOLCANIC SANDS											
WBB/VOL	0.0	00	19.74	9.52	15,712	8.52	25.77	0.00	0.00	3,711.27	8,494.10
I - REE Data from Taylor and McLennan, 1985 - Table 11.	Γ	4	- Taylor	and McL	ennan, 198.	5 - Table	e 2.15				
2 - Major data presented in %, all other results in ppm 3 - Muhs et al, 2012		S	- Muhs, 2	2007 and	Muhs, 201	0					

Table 7b. Corrected, major, trace and REE data by INAA

Sample	Nd	Sm	Eu	Gd	$\mathbf{T}\mathbf{b}$	Ho	Yb	Lu	Hf	Та
Chondrite ¹	0.711	0.231	0.087	0.306	0.058	0.085	0.248	0.038	0.179	0.026
Upper Volcanics ³	88.2	14.7	4.09	10.3	1.35	1.29	2.12	0.29	7.89	10.1
ucc4	26	4.5	0.88	3.8	0.64	0.80	2.20	0.32	5.8	2.2
Average Dust ⁵	43.57	9.00	1.81	7.94	1.15	1.52	3.60	0.53	6.03	1.8
PALEOSOLS										
F-2	361.88	73.45	19.54	42.77	12.29	16.50	52.78	8.44	11.30	11.59
LD-1	336.67	60.48	15.16	33.21	9.74	17.67	46.35	7.05	12.12	6.67
PR-3	95.90	4.70	9.09	21.28	6.80	4.53	26.16	2.33	7.92	3.48
Z-10	241.70	49.58	12.73	31.64	9.25	13.58	43.05	6.77	9.78	3.04
Z-3	224.62	48.10	11.84	32.17	8.86	12.60	41.03	5.96	12.32	4.03
SG-1	203.80	44.01	11.02	25.45	7.57	15.98	35.31	5.84	8.88	2.88
FS-1	180.24	39.58	9.50	24.93	8.04	15.06	36.70	5.62	8.92	3.22
LD-5	1,407.91	172.19	53.82	128.48	20.85	13.24	29.34	1.83	15.38	166.05
Q-10	229.79	54.50	14.52	29.21	9.44	18.08	60.46	10.39	5.43	2.04
SS66-NON										
SOM-2	74.18	56.41	14.25	34.30	10.49	20.15	51.21	7.67	8.45	1.88
SHR-2	231.57	6.65	12.75	30.25	8.78	13.03	32.12	4.95	10.65	3.35
PG-1	210.22	6.29	19.17	35.19	13.78	27.62	70.83	9.58	11.42	1.40
RBD-5	155.84	11.51	11.61	31.68	9.54	7.95	33.59	0.00	11.96	2.52
ORD-1	332.61	74.44	20.11	27.40	13.50	21.59	61.84	8.84	10.65	3.27
GRR-1	407.67	84.92	23.18	38.53	15.39	28.17	74.34	10.24	12.06	4.46
GRR-3	373.52	76.97	20.91	33.77	13.45	21.90	58.02	8.69	11.02	3.63
PY-3	56.51	52.12	12.33	35.51	9.80	13.82	40.62	5.60	26.21	10.04
IQL	86.70	7.92	13.97	27.25	8.05	9.06	10.32	26.17	22.40	4.53
IRE-1	222.76	44.48	10.34	34.79	8.22	11.75	38.95	6.43	15.17	5.60

Table 7b. Corrected, major, trace and REE data by INAA

rrected, major, trace and REE data by INAA	Sample	Chondrite ¹	Upper Volcanics ³	UCC ⁴	Average Dust ⁵	MH-5	MH-10	9-HW	MH-7	MH-8	6-HW	Min:	Mean:	Max:	SUBMARINE CLAYS	HSCS/1	HSSP-01	HUN-1	Mean:	VOLCANIC SANDS WBR/VOI
	Nd	0.711	88.2	26	43.57	57.60	268.22	287.06	286.85	176.11	70.19	56.51	263.20	1,407.91		139.50	242.82	149.67	177.33	98 273
	Sm	0.231	14.7	4.5	9.00	54.04	54.62	56.99	57.60	35.16	50.01	4.70	51.07	172.19		26.46	43.75	28.33	32.85	10 7 47
	Eu	0.087	4.09	0.88	1.81	13.73	14.51	15.46	15.66	8.78	14.52	8.78	15.94	53.82		6.12	10.16	6.74	7.67	120.13
	Gd	0.306	10.3	3.8	7.94	29.99	31.34	31.42	29.53	25.09	28.59	21.28	34.95	128.48		21.54	31.05	29.99	27.53	337.00
	$\mathbf{T}\mathbf{b}$	0.058	1.35	0.64	1.15	2.33	10.97	12.22	12.36	6.43	1.74	1.74	10.00	20.85		4.45	6.78	5.76	5.66	78 VC
	Ho	0.085	1.29	0.80	1.52	14.64	17.91	17.95	22.83	11.34	7.24	4.53	15.77	28.17		6.96	12.01	6.88	8.62	10 73
	Yb	0.248	2.12	2.20	3.60	6.53	34.25	36.39	44.03	32.80	39.06	6.53	41.44	74.34		16.31	25.15	21.45	20.97	16 36
	Lu	0.038	0.29	0.32	0.53	1.64	3.05	4.91	5.63	5.07	0.00	0.00	6.51	26.17		2.55	3.81	3.20	3.19	00.0
	Ηf	0.179	7.89	5.8	6.03	11.26	11.62	9.80	7.43	11.22	7.24	5.43	11.62	26.21		11.33	11.84	13.45	12.20	17 22
	Та	0.026	10.1	2.2	1.8	3.84	4.12	3.68	2.11	3.78	2.72	1.40	10.56	166.05		3.79	6.11	4.90	4.94	572 2/2

REE Data from Taylor and McLennan, 1985 - Table 11.1
Major data presented in %, all other results in ppm
Muhs et al, 2012

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	Age	Exp.	SiO ₂	TiO ₂	Al_2O_3	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K_2O	P_2O_5	LOI
Age	1.00												
Exp.	0.86	1.00											
SiO ₂	-0.80	-0.69	1.00										
TiO ₂	0.16	-0.05	-0.19	1.00									
Al_2O_3	0.69	0.53	-0.66	0.14	1.00								
Fe_2O_3	0.64	0.47	-0.62	0.52	0.83	1.00							
MnO	-0.28	-0.36	-0.12	0.10	0.09	0.10	1.00						
MgO	-0.44	-0.31	0.40	-0.34	-0.83	-0.85	-0.10	1.00					
CaO	-0.59	-0.53	0.35	-0.27	-0.61	-0.60	0.14	0.68	1.00				
Na ₂ O	-0.24	-0.24	0.13	-0.23	-0.57	-0.48	-0.12	0.51	0.65	1.00			
K ₂ O	-0.01	-0.07	-0.26	-0.18	0.11	0.08	0.55	-0.14	0.04	0.11	1.00		
P_2O_5	0.44	0.56	-0.61	0.10	0.30	0.25	0.00	-0.29	-0.51	-0.32	0.13	1.00	
LOI	-0.54	-0.49	0.31	-0.33	-0.59	-0.60	0.12	0.69	0.99	0.66	0.05	-0.49	1.00

Table 8a. r-values for paleosol carbonate-free major elements (no MH-8 or LD-5)

Table 8b. R²-values for paleosol carbonate-free major elements (no MH-8 or LD-5)

			1			5		,					
	Age	Exp.	SiO ₂	TiO ₂	Al_2O_3	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P_2O_5	LOI
Age	1.00												
Exp.	0.74												
SiO_2	0.63	0.48	1.00										
TiO ₂	0.03	0.00	0.04	1.00									
Al_2O_3	0.47	0.28	0.44	0.02	1.00								
Fe ₂ O ₃	0.41	0.22	0.38	0.27	0.69	1.00							
MnO	0.08	0.13	0.01	0.01	0.01	0.01	1.00						
MgO	0.19	0.09	0.16	0.12	0.69	0.73	0.01	1.00					
CaO	0.35	0.28	0.12	0.07	0.37	0.36	0.02	0.46	1.00				
Na ₂ O	0.06	0.06	0.02	0.05	0.33	0.23	0.01	0.26	0.43	1.00			
K ₂ O	0.00	0.00	0.07	0.03	0.01	0.01	0.30	0.02	0.00	0.01	1.00		
P_2O_5	0.19	0.31	0.38	0.01	0.09	0.06	0.00	0.09	0.26	0.10	0.02	1.00	
LOI	0.29	0.24	0.10	0.11	0.35	0.36	0.01	0.47	0.97	0.44	0.00	0.24	1.00

	$^{\mathrm{Pb}}$																				1.00
	Sn																			1.00	0.41
	Nb																		1.00	-0.17	0.25
	Υ																	1.00	0.66	-0.03	0.42
	\mathbf{Sr}																1.00	-0.19	-0.33	0.61	0.70
	Rb															1.00	0.96	0.02	-0.15	0.59	0.83
	Ga														1.00	0.85	0.83	-0.09	-0.05	0.38	0.74
	Zn													1.00	-0.06	0.13	-0.06	0.69	0.43	-0.05	0.43
	Cu												1.00	0.15	0.88	0.92	0.87	0.02	-0.05	0.32	0.81
	Ni											1.00	0.22	0.38	0.09	0.28	0.16	0.22	0.33	0.39	0.30
	Nd										1.00	0.15	0.11	-0.38	0.10	0.30	0.40	-0.34	-0.33	0.55	-0.04
	Ce									1.00	0.97	0.14	0.23	-0.43	0.22	0.40	0.49	-0.42	-0.37	0.50	0.02
ments	La								1.00	0.08	-0.02	-0.11	0.42	-0.06	0.33	0.30	0.25	0.09	0.05	0.03	0.12
ace ele	Ba							1.00	0.76	0.24	0.06	-0.14	0.71	-0.21	0.63	0.56	0.57	-0.22	-0.25	-0.04	0.30
e-free tr	Zr						1.00	0.72	0.34	0.66	0.54	-0.06	0.65	-0.40	0.61	0.67	0.79	-0.60	-0.55	0.34	0.26
urbonate	Co					1.00	0.55	0.26	-0.10	0.39	0.31	0.27	0.68	-0.04	0.75	0.78	0.82	-0.22	-0.17	0.51	0.61
sols c	\mathbf{Cr}				1.00	-0.13	0.05	0.10	0.20	-0.01	0.00	-0.02	-0.14	-0.20	-0.22	-0.22	-0.14	0.00	0.28	-0.08	-0.19
or palec	Λ			1.00	-0.13	0.49	0.35	0.38	0.14	0.10	0.02	0.24	0.39	-0.14	0.59	0.43	0.41	-0.26	-0.22	0.30	0.22
alues fo	\mathbf{Sc}		1.00	-0.26	0.03	-0.23	-0.25	-0.23	-0.06	-0.28	-0.25	-0.17	-0.27	0.12	-0.33	-0.21	-0.23	0.19	-0.06	-0.07	-0.11
9a. r-v	CI	1.00	-0.26	0.36	-0.22	0.47	0.37	-0.04	-0.11	0.61	0.63	0.62	0.30	-0.13	0.29	0.54	0.54	-0.20	-0.24	0.79	0.30
Table		CI	\mathbf{Sc}	>	Cr	Co	Zr	Ba	La	Ce	Νd	ïZ	Cu	\mathbf{Zn}	Ga	Rb	\mathbf{Sr}	Υ	qN	Sn	Рb

	$^{\mathrm{Pb}}$																				1.00
	Sn																			1.00	0.16
	βŊ																		1.00	0.03	0.06
	Υ																	1.00	0.44	0.00	0.17
	\mathbf{Sr}																1.00	0.04	0.11	0.37	0.49
	Rb															1.00	0.92	0.00	0.02	0.35	0.68
	Ga														1.00	0.73	0.68	0.01	0.00	0.14	0.54
	Zn													1.00	0.00	0.02	0.00	0.48	0.19	0.00	0.19
	Cu												1.00	0.02	0.78	0.84	0.75	0.00	0.00	0.10	0.65
	N:											1.00	0.05	0.14	0.01	0.08	0.03	0.05	0.11	0.15	0.09
	Nd										1.00	0.02	0.01	0.14	0.01	0.09	0.16	0.12	0.11	0.30	0.00
	Ce									1.00	0.95	0.02	0.05	0.19	0.05	0.16	0.24	0.18	0.14	0.25	0.00
ements	La								1.00	0.01	0.00	0.01	0.18	0.00	0.11	0.09	0.06	0.01	0.00	0.00	0.02
trace el	Ba							1.00	0.57	0.06	0.00	0.02	0.50	0.04	0.40	0.31	0.33	0.05	0.06	0.00	0.09
te-free	Zr						1.00	0.52	0.12	0.43	0.29	0.00	0.42	0.16	0.37	0.44	0.63	0.36	0.30	0.11	0.07
carbona	Co					1.00	0.30	0.07	0.01	0.15	0.10	0.07	0.46	0.00	0.57	0.61	0.67	0.05	0.03	0.26	0.38
cosols o	\mathbf{Cr}				1.00	0.02	0.00	0.01	0.04	0.00	0.00	0.00	0.02	0.04	0.05	0.05	0.02	0.00	0.08	0.01	0.04
for pale	Λ			1.00	0.02	0.24	0.12	0.15	0.02	0.01	0.00	0.06	0.15	0.02	0.34	0.18	0.17	0.07	0.05	0.09	0.05
values	\mathbf{Sc}		1.00	0.07	0.00	0.05	0.06	0.05	0.00	0.08	0.06	0.03	0.07	0.01	0.11	0.05	0.05	0.04	0.00	0.00	0.01
b. R ² -,	CI	1.00	0.07	0.13	0.05	0.22	0.14	0.00	0.01	0.38	0.39	0.38	0.09	0.02	0.08	0.29	0.29	0.04	0.06	0.62	0.09
Table 9		C1	Sc	>	\mathbf{Cr}	Co	Zr	Ba	La	Ce	ΡŊ	ïZ	Cu	Zn	Ga	Rb	Sr	Υ	qN	Sn	Pb

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Table 10	Madalina	- f	-1	-1	1	
I able IV.	Modeling	of major	element	changes	aue to	weathering
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	Shallow	Volcanic	Amount	Gain	Gain
Parameter	Volcanics	Sands	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	37.70	4.92	0.52	-37.18	-98.61
TiO ₂	5.02	47.03	5.02	0.00	0.00
Al ₂ O ₃	11.40	1.86	0.20	-11.20	-98.26
Fe ₂ O ₃	13.60	8.19	0.87	-12.73	-93.57
MnO	0.23	0.19	0.02	-0.21	-91.32
MgO	6.89	0.99	0.11	-6.78	-98.47
CaO	13.90	32.39	3.46	-10.44	-75.13
Na ₂ O	3.00	0.33	0.03	-2.97	-98.84
K ₂ O	2.06	0.02	0.00	-2.06	-99.90
P_2O_5	1.08	0.13	0.01	-1.07	-98.73
LOI	5.31	0.90	0.10	-5.21	-98.19
Totals:	100.19	96.93	10.67	-89.84	

10a. Weathering of shallow volcanic basement to volcanic sands via TiO₂

10b. Weathering of shallow volcanic basement to volcanic sands via Al_2O_3

	Shallow	Average	Amount	Gain	Gain
Parameter	Volcanics	Paleosol	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	37.70	31.39	0.102	-37.60	-99.73
TiO ₂	5.02	3.11	0.010	-5.01	-99.80
Al_2O_3	11.40	35.10	11.400	0.00	0.00
Fe ₂ O ₃	13.60	16.40	0.053	-13.55	-99.61
MnO	0.23	0.18	0.001	-0.23	-99.75
MgO	6.89	5.77	0.019	-6.87	-99.73
CaO	13.90	0.00	0.000	-13.90	-100.00
Na ₂ O	3.00	2.82	0.009	-2.99	-99.69
K ₂ O	2.06	1.09	0.004	-2.06	-99.83
P_2O_5	1.08	4.13	0.013	-1.07	-98.76
LOI	5.31	0.00	0.000	-5.31	-100.00
Totals:	100.19	<i>99</i> . <i>99</i>	0.32	-88.58	

	African	Average	Amount	Gain	Gain
Parameter	Dust	Paleosol	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	60.14	31.39	10.75	-49.39	-82.13
TiO ₂	0.72	3.11	1.06	0.34	47.91
Al ₂ O ₃	12.02	35.10	12.02	0.00	0.00
Fe ₂ O ₃	5.82	16.40	5.61	-0.21	-3.53
MnO	0.06	0.18	0.06	0.00	4.50
MgO	2.29	5.77	1.98	-0.31	-13.71
CaO	5.24	0.00	0.00	-5.24	-100.00
Na ₂ O	0.61	2.82	0.97	0.36	58.52
K ₂ O	1.71	1.09	0.37	-1.34	-78.10
P_2O_5	0.24	4.13	1.41	1.17	489.46
LOI	8.64	0.00	0.00	-8.64	-100.00
Totals:	97.49	100.00	34.24	-63.25	

10c. Weathering of African dust to paleosol via TiO_2

10d. Weathering of African dust to paleosol via Al_2O_3

	African	Average	Amount	Gain	Gain
Parameter	Dust	Paleosol	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	60.14	31.39	7.27	-52.87	-87.92
TiO ₂	0.72	3.11	0.72	0.00	0.00
Al ₂ O ₃	12.02	35.10	8.13	-3.89	-32.39
Fe ₂ O ₃	5.82	16.40	3.80	-2.02	-34.78
MnO	0.06	0.18	0.04	-0.02	-29.35
MgO	2.29	5.77	1.34	-0.95	-41.66
CaO	5.24	0.00	0.00	-5.24	-100.00
Na ₂ O	0.61	2.82	0.65	0.04	7.17
K ₂ O	1.71	1.09	0.25	-1.46	-85.19
P_2O_5	0.24	4.13	0.96	0.72	298.52
LOI	8.64	0.00	0.00	-8.64	-100.00
Totals:	97.49	100.00	23.15	-74.34	

Table 10.	Modeling	of majo	r element	changes	due to	weathering
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	African	Submarine	Amount	Gain	Gain
Parameter	Dust	Clay	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	60.14	44.45	16.18	-43.96	-73.09
TiO ₂	0.72	2.44	0.84	0.12	16.08
Al ₂ O ₃	12.02	33.02	12.02	0.00	0.00
Fe ₂ O ₃	5.82	10.98	3.76	-2.06	-35.39
MnO	0.06	0.03	0.01	-0.05	-81.25
MgO	2.29	2.69	0.92	-1.37	-59.83
CaO	5.24	0.00	0.00	-5.24	-100.00
Na ₂ O	0.61	2.32	0.80	0.19	30.47
K ₂ O	1.71	1.63	0.56	-1.15	-67.35
P_2O_5	0.24	2.43	0.83	0.59	246.76
LOI	8.64	0.00	0.00	-8.64	-100.00
Totals:	97.49	100.00	36.40	-61.57	

10e. Weathering of African dust subamrine clay via TiO_2

10f. Weathering of African dust to submarine clay via Al_2O_3

	African	Submarine	Amount	Gain	Gain
Parameter	Dust	Clay	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	60.14	44.45	13.11	-47.03	-78.20
TiO ₂	0.72	2.44	0.72	0.00	0.00
Al ₂ O ₃	12.02	33.02	7.64	-4.38	-36.40
Fe ₂ O ₃	5.82	10.98	2.54	-3.28	-56.32
MnO	0.06	0.03	0.01	-0.05	-87.32
MgO	2.29	2.69	0.62	-1.67	-72.84
CaO	5.24	0.00	0.00	-5.24	-100.00
Na ₂ O	0.61	2.32	0.54	-0.07	-11.79
K ₂ O	1.71	1.63	0.38	-1.33	-77.93
P_2O_5	0.24	2.43	0.56	0.32	134.44
LOI	8.64	0.00	0.00	-8.64	-100.00
Totals:	97.49	100.00	29.50	-71.36	

		African	Amount	Gain	Gain
Parameter	UCC	Dust	Remaining (g)	or Loss (g)	or Loss (%)
SiO ₂	65.89	60.14	41.76	-24.13	-36.62
TiO ₂	0.50	0.72	0.50	0.00	0.00
Al ₂ O ₃	15.17	12.02	8.35	-6.82	-44.98
Fe ₂ O ₃	4.49	5.82	4.04	-0.45	-9.99
MnO	0.07	0.06	0.04	-0.03	-40.48
MgO	2.20	2.29	1.59	-0.61	-27.71
CaO	4.19	5.24	3.64	-0.55	-13.15
Na ₂ O	3.89	0.61	0.42	-3.47	-89.11
K ₂ O	3.39	1.71	1.19	-2.20	-64.97
P_2O_5	0.20	0.24	0.17	-0.03	-16.67
LOI	0.01	8.64	6.00	5.99	59900
Totals:	100.00	97.49	69.44	-32.30	

10g. Weathering of UCC to African dust via TiO_2

APPENDIX A

Submarine Core Logs and Seismic Profiles



Core log from Vollbrecht, 1996



Seismic profile from Vollbrecht, 1996



Core log from Vollbrecht, 1996



Seismic profile from Vollbrecht, 1996

APPENDIX B

X-ray Diffraction Data Sheets



Harrington Sound Submarine Core Sample HSCS-1 – XRD Data

Hungry Bay Submarine Core Sample Hun-1 – XRD Data

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Castle Harbour Geosol Sample MH-8 – XRD Data



Shore Hills Geosol Sample SHR-1 – XRD Data

APPENDIX C

Data Quality Assessment

Precision	of Instrum	nental Ne	utron Acti	vation An	alysis				
Sample	RGM-1		RGM-1	RGM-1	RGM-1	RGM-1	Precision		
Batch	Published	St.Dev.	Batch	Batch	Batch	Batch	Mean	St.Dev.	Error (%)
Element			252	256	262 + 264	264 + 265			
Na20*	4.07	0.15	4.18	4.07	4.38	4.27	4.22	0.14	3.79
Sc	4.40	0.30	4.50	4.24	4.79	4.46	4.50	0.23	2.21
Cr	3.70	1.20	3.52	3.30	3.39	2.97	3.29	0.24	11.01
FeO*	1.77	0.01	0.59	1.58	1.79	1.71	1.42	0.56	19.80
Co	2.00	0.20	1.43	1.94	2.18	2.02	1.89	0.32	5.48
Rb	149.00	8.00	144.62	138.78	149.16	139.45	143.00	4.87	4.03
Cs	9.60	0.60	9.86	9.02	10.51	9.18	9.64	0.68	0.43
La	24.00	1.10	23.25	21.48	24.77	23.05	23.14	1.35	3.59
Ce	47.00	4.00	48.49	28.06	51.18	3.04	32.69	22.30	30.44
Nd	19.00	1.00	19.80	19.20	25.41	20.76	21.29	2.82	12.07
Sm	4.30	0.30	2.16	3.77	4.45	4.06	3.61	1.01	16.07
Eu	0.66	0.08	0.44	0.59	0.65	0.60	0.57	0.09	13.72
Gd	3.70	0.40	9.01	8.89	9.75	9.47	9.28	0.40	150.80
\mathbf{Tb}	0.66	0.06	0.44	0.62	0.53	0.50	0.52	0.08	20.81
Ho	0.95	0.22	1.30	1.25	1.43	1.44	1.36	0.09	42.77
Yb	2.60	0.30	2.59	3.40	2.72	2.56	2.82	0.39	8.37
Lu	0.41	0.03	0.40	0.12	0.45	0.42	0.35	0.15	15.17
Ηf	6.20	0.30	6.48	6.13	6.79	6.59	6.50	0.28	4.78
Ta	0.95	0.10	0.85	0.98	1.29	1.15	1.07	0.19	12.12
Th/Pa	15.10	1.30	14.94	14.23	15.60	14.72	14.87	0.57	1.52
U/Np	5.80	0.50	6.36	5.55	4.99	5.51	5.60	0.56	3.44
Note:	* - denote:	s concentr	ation prese	inted as pe	rcentage, a	ll others in	parts per mi	illion (ppm)	
	- denotes	% error >	20%		I		I	1	





Na₂O INAA vs XRF



FeO Fe2O3 INAA vs XRF





Cr INAA vs XRF



APPENDIX D

Carbonate Data Correction Method

CaO Correction Method

The paleosol, soil and submarine clay samples analyzed for this investigation all contained varying concentrations of calcium carbonate, in the form of lithic fragments or reprecipitated calcite cement. In order to make direct elemental comparisons it was necessary to renormalize the data to remove the CaO using the original CaO concentrations and Loss On Ignition (LOI) data. The data were corrected as follows:

- 1. The molecular weight (MW) of Ca = 40.07, O=15.99, and C=12.01.
- 2. Therefore, the MW of CaO = 56.03 (40.04 + 15.99) and the molecular weight of CO₂ = $44.01 (12.01 + (2 \times 16))$.
- In addition, CaO represents ~56.03% of CaCO₃ and CO2 represents ~43.97% of CaCO₃ by the following:

$$CaCO_3 = MW_{CaO} \div (MW_{CaO} + MW_{CO2})$$

- 4. As a result, the ratio of $CO_2/CaO = \sim 0.785\%$.
- The CaO values determined by XRF were multiplied by the CO₂/CaO ratio to obtain "LOI due to CaCO₃".
- 6. "Excess LOI" was determined by subtracting "LOI due to CaCO₃" from LOI.
- 7. A "Carbonate Free Factor" was calculated by the following equation:

Carbonate Free Factor = $100 \div (100 - CaO - LOI due to CaCO_3)$

8. A "Carbonate and Volatile Free Factor" was calculated as follows:

Carbonate and Volatile Free Factor = $100 \div (100 - LOI - CaO)$

- 9. A "Normalization Factor" was then calculated by dividing 100 by the sum of the major elements, not including CaO.
- 10. The elemental concentrations were then corrected by multiplying the original laboratory values by the Normalization Factor.

			ļ	T otal W	/eight at	L03	ss at
	Crucible -	Crucible & Sample	= Sample	525°C	$1,100^{\circ}C$	525°C	$1,100^{6}$
Sample	(g)	(g)	(g)	(g)	(g)	(%)	(%)
F-2							
LD-1	9.36	12.22	2.86	11.96	11.39	9.09	29.02
PR-3	8.72	10.88	2.16	10.59	10.32	13.43	25.93
Z-10	24.61	30.29	5.68	30.18	27.84	1.94	43.13
Z-3	24.66	25.55	0.89	25.45	25.21	11.24	38.20
SG-1							
FS-1	22.40	26.52	4.12	26.42	25.40	2.43	27.18
LD-5	8.46	11.85	3.39	11.75	10.35	2.95	44.25
Q-10	9.02	11.19	2.17	10.72	10.67	21.66	23.96
SS66-NON	24.42	29.22	4.80	29.02	26.96	4.17	47.08
SOM-2	8.62	11.64	3.02	11.48	10.15	5.30	49.34
SHR-2	23.02	28.70	5.68	28.42	26.32	4.93	41.90
PG-1	24.80	25.86	1.06	25.85	25.39	0.94	44.34
RBD-5							
ORD-1	27.58	30.48	2.90	30.30	29.24	6.21	42.76
GRR-1	9.79	13.03	3.24	12.67	11.55	11.11	45.68
GRR-3	8.76	12.56	3.80	12.17	11.04	10.26	40.00
PY-3							
IQL	23.65	26.00	2.35	25.67	25.17	14.04	35.32
IRE-1	8.54	11.55	3.01	11.17	10.27	12.62	42.52
MH-5	9.46	12.23	2.77	11.72	11.23	18.41	36.10
MH-10	9.55	12.55	3.00	11.78	11.67	25.67	29.33
9-HIM	8.47	10.54	2.07	10.03	9.98	24.64	27.05
MH-7	8.82	10.87	2.05	10.47	10.41	19.51	22.44
MH-8	8.96	12.56	3.60	11.93	11.76	17.50	22.22
6-HIM	10.52	13.73	3.21	13.23	12.69	15.58	32.40
HSCS/1	10.43	14.26	3.83	13.75	13.49	13.32	20.10
HSSP	26.21	27.13	0.92	26.93	26.81	21.74	34.78
HUN-1	23.72	27.27	3.55	26.95	26.72	9.01	15.49
WBB/VOL	23.41	25.75	2.34	25.75	25.62	0.00	5.56