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GENESIS OF MARINE TERRACE SOILS, BARBADOS, WEST INDIES: EVIDENCE FROM MINERALOGY AND GEOCHEMISTRY

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ABSTRACT

Well-developed, clay-rich soils dominated by interstratified kaolinite–smectite are found on the uplifted coral reef terraces on the island of Barbados. The reef limestone is unlikely to have been the soil parent material however, because it is 98 per cent $CaCO_3$ and geomorphic evidence argues against the 20 m of reef solution required to produce the soils by this process. The mineralogy of the sand, silt, and clay fractions of the soils, and trace element geochemistry, suggest that aeolian materials carried on the trade winds from Africa, volcanic ash from the island of St. Vincent, and quartz from Tertiary bedrock on the island itself are the parent materials for the soils.

INTRODUCTION

Clay-rich soils, often attaining great thicknesses, or thick bauxite deposits have been found on limestone terrains on many Caribbean and western Atlantic islands, including Barbados, Jamaica, Haiti, the Dominican Republic, the Bahamas, and Bermuda. The origin of such soils has been debated for many years because much of the limestone is of high purity. Four modes of genesis have been suggested for such soils and bauxites: (1) Accumulation of insoluble residues from the underlying carbonate rock (Schmedeman, 1948, 1950; Hartman, 1955; Hill, 1955; Hose, 1963; Clarke, 1966; Ahmad et al., 1966; Sinclair, 1967; Ahmad and Jones, 1969a; Blackburn and Taylor, 1969; (2) Deposition on lower limestone surfaces of fluvially-transported clays from topographically-higher positions (Zans, 1959; Chubb, 1963); (3) Weathering of volcanic ash which has fallen on the limestone surfaces (Goldich and Bergquist, 1947, 1948; Burns, 1961; Waterman, 1962; Comer, 1974; Comer et al., 1980); and (4) Weathering of fine-grained aeolian sediments which have origins in distant regions (Syers et al., 1969; Bricker and Mackenzie, 1970; Prospero et al., 1981). On Barbados, all four mechanisms have been used at one time or another to explain the origin of clay-rich soils found on tectonically-uplifted coral limestone terraces of Quaternary age. In this paper, we present new mineralogical and geochemical data from Barbados soils and hypothesized parent materials in order to evaluate the four possible modes of origin described above.

STUDY AREA AND METHODS

Barbados is located approximately 145 km east of the Lesser Antilles island chain in the southern Caribbean Sea (Figure 1). The island apparently represents the emergent portion of the Barbados submarine ridge and exposes the crestal zone of a broad accretionary prism of the Lesser Antilles forearc (Speed and Larue, 1982). In
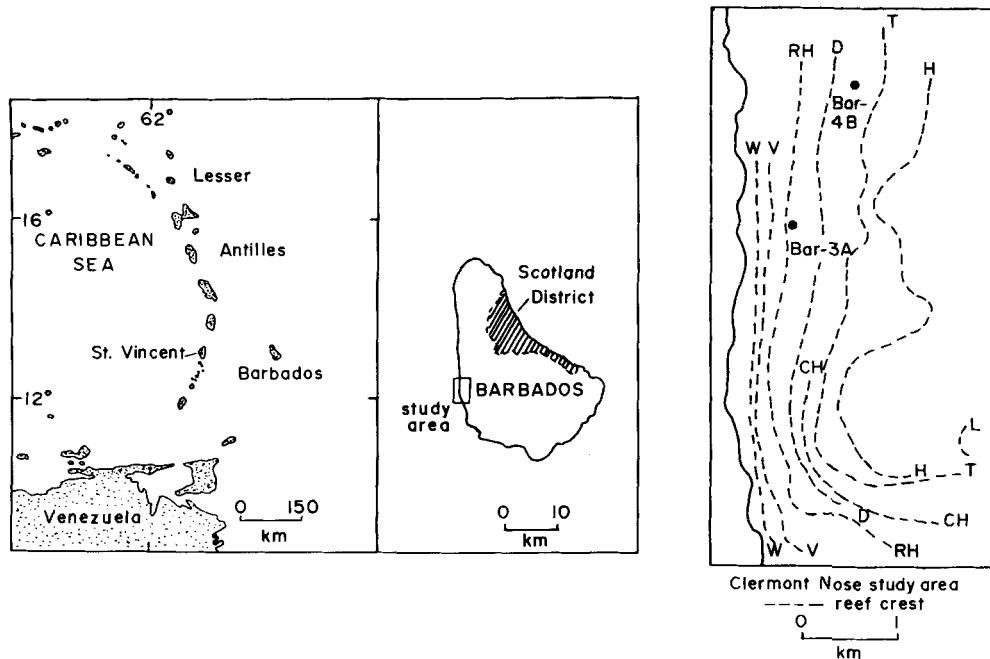
Figure 1. Location of study area in the Clermont Nose area of Barbados. Reef terrace names and ages are from Bender et al. (1979) as follows: W = Worthing (82 ka); V = Ventnor (105 ka); RH = Rendezvous Hill (125 ka); D = Durants (~ 180 ka); CH = Cave Hill (~ 200 ka); T = Thorpe (~ 220 ka); H = Husbands (~ 320 ka); L =·Lodge (~ ~ 360 ka)

contrast to the Lesser Antilles chain which is largely volcanic, Barbados is composed of Tertiary sedimentary rocks overlain by Quaternary coral reef limestones. Pre-Quaternary rocks on the island are (1) a basal complex of quartzose turbidites, hemipelagites, and melange and (2) a younger series of nappes composed mostly of calcareous pelagite (Larue et al., 1985). Overlying the Tertiary rocks is a series of coral reef terraces, each with distinctive forereef, crest, and backreef facies (Mesolella et al., 1969). Barbados has been tectonically uplifted in Quaternary time at a rate of about 0·3 m/1000 y based on studies of the coral reef terraces by Mesolella et al. (1969), James et al. (1971), and Bender et al. (1979).These workers mapped the terraces in detail and provided uranium-series dates of coral which indicate ages of terraces ranging from ~ 60,000 y BP to ~ 640,000 y BP. Still higher terraces that have been mapped are beyond the range of U-series dating.

The present climate of Barbados is of the trade-wind littoral type, characterized by moderate seasonal changes in temperature (mean annual temperature ranges from 24–28°C), but with a distinctive dry season which lasts from about December to May. Mean annual precipitation ranges from about 110 cm on the coasts to over 212 cm in the Scotland District (Figure 1) which is the area of highest elevation on the island (Rouse and Watts, 1966).

Soils developed on Quaternary marine terraces were sampled on the west coast of Barbados, in the Clermont Nose area as mapped by Bender et al. (1979). We described and sampled 11 soil profiles from hand-dug pits on coral terraces ranging in age from ~ 125,000 y to ~ 1,000,000 y (Muhs et al., 1986). Soils on the Rendezvous Hill and Durants terraces (Figure 1) which have uranium-series dates of ~ 125,000 and ~ 180,000 y BP respectively (Mesolella et al., 1969; Bender et al., 1979) were selected for detailed mineralogical and chemical analyses in this study. Soils were described according to the new terminology of the U.S. Soil Conservation Service (Guthrie and Witty, 1982) and classified according to the Soil Survey Staff (1975). The soil on the Rendezvous Hill terrace is in an area near the boundary between a Black Soil delineation and Grey–Brown Soil delineation as mapped by Vernon and Carroll (1965). The soil on the Durants terrace is in a delineation mapped as a Grey–Brown Soil, sandy variant. These locations were chosen because they are on flat-lying (0–2 per cent slopes) sites near reef crests where erosion has been minimal and the soils apparently have not been cultivated.

Particle size analyses were done by the pipette method (Day, 1965) and bulk density was done by the clod method (Blake, 1965). Organic matter was determined by the Walkley–Black method and $CaCO_3$ content was determined by CO_2 evolution with a Chittick apparatus (Dreimanis, 1962). Cation exchange capacity was determined by ammonium acetate saturation and titration and pH was measured on 1:1 soil–water pastes using a glass electrode. Total secondary iron (both crystalline and amorphous) was extracted by the dithionite–citrate–bicarbonate method; quantities were measured colorimetrically. Clay was isolated by sedimentation after destruction of carbonates with $NaC_2H_3O_2 \cdot 3H_2O$ and organic matter with H_2O_2, and dispersion with Na-hexametaphosphate. Clay samples were saturated with Mg and mounted on glass slides for X-ray diffraction analysis. The clay samples were X-rayed three times: in an air-dry condition, after solvation with ethylene glycol, and after heat treatment (550°C for two hours). Silts were separated from sands by wet sieving and from clays by repeated sedimentation and siphoning. The silt-size fractions were then ground to a fine powder and X-rayed as random mounts. The sand-size fractions were investigated by X-ray diffraction of ground powders and by optical inspection with a petrographic microscope using the very fine sand fraction. Major element analysis was done by wavelength-dispersive X-ray fluorescence spectroscopy according to Taggart et al. (1982). Concentrations of the trace elements Rb, Sr, Y, Zr, and Nb were determined by energy-dispersive X-ray fluorescence spectroscopy; concentrations of Th were determined by isotope-dilution alpha spectrometry (Rosholt, 1984, 1985).

SOIL MORPHOLOGY AND MINERALOGY

The soils we sampled have morphologies similar to those described by Vernon and Carroll (1965) and Ahmad and Jones (1969b). The profile on the ~ 125,000 y old Rendezvous Hill terrace (Bar-3A) has little evidence of horizonation, exhibits strong angular blocky structure, and has slickensides (Table I). Thus, we classify this soil as a Vertisol (Typic Chromudert), assuming that it develops desiccation cracks later in the dry season (no cracks were visible at the time of sampling). The soil on the ~ 180,000 y old Durants terrace (Bar-4B) has strongly-expressed horizons due to colour differences, but does not have clay films (Table I). Hence, we classify this soil as a Mollisol (Typic Hapludoll), but note that some pedoturbation processes are at work because of pressure faces on ped surfaces and the lack of textural change as a function of depth (Table I). Bar-4B is thinner than Bar-3A even though it is on an older geomorphological surface; hence it has probably been partially eroded.

Particle size data confirm the observations of earlier workers that Barbados terrace soils are high in clay (Table II). The lower clay content of Bar-4B compared to Bar-3A is probably due to greater admixtures of

Table I. Morphological properties of Barbados soils

Profile	Horizon	Depth (cm)	Colour (moist)	Texture	Structure	Coatings	Boundary
Bar-3A	A1	0–12	10YR 3/2	clay	1,fm,sbk	occasional pressure faces	gs
	A2	12–33	10YR 3/2	clay	2,fm,sbk	common pressure faces	gs
	A3	33–40	10YR 3/2	clay	3,fms,abk	common slickensides	cw
	C	40–70	2.5Y 5/2	clay	3,c,abk	common slickensides, occasional clay films	
	R	70+	weathered coral limestone				
Bar-4B	A	0–35	7.5Y 3/2	sandy clay	2,cm,abk	occasional pressure faces	cs
	Bw1	35–48	5YR 3/4	sandy clay	2, cm, abk	common pressure faces	gs
	Bw2	48–60	7.5YR 5/6	sandy clay	1,cm,sbk	—	ai
	R	60+	indurated beach sediments & coral limestone				

Abbreviations: 1 = weak; 2 = moderate; 3 = strong; f = fine; m = medium; c = coarse; sbk = subangular block; abk = angular block; gs = gradual smooth; cw = clear wavy; cs = clean smooth; ai = abrupt irregular.

Table II. Physical and chemical properties of Barbados soils

Profile	Horizon	Organic Depth (cm)	Matter (%)	Sand* (%)	Silt* (%)	Clay* (%)	Bulk density (g cm^{-3})	$CaCO_3$ (%)	pH (1:1)	CEC† (meq/100 g)	Fe_d‡ (%)
Bar-3A	A1	0–6	5·3	21·9	13·9	64·2	1·55	4·6	7·30	45·1	0·91
	A1	6–12	4·0	22·2	11·6	66·2	1·55	6·2	7·75	44·2	1·78
	A2	12–22	2·4	21·6	11·5	66·9	1·51	5·1	7·35	45·0	1·61
	A2	22–33	2·4	20·7	12·0	67·3	1·47	4·4	7·55	39·0	1·24
	A3	33–40	2·0	20·6	13·0	66·4	1·65	5·2	7·80	44·8	1·23
	C	40–50	1·6	12·9	5·7	81·4	1·59	22·0	7·95	40·3	0·67
	C	50–60	1·3	10·6	1·5	87·9	1·77	36·0	8·05	36·6	0·39
	C	60–70	1·3	4·0	1·9	94·1	1·71	41·8	8·05	22·7	0·25
Bar-4B	A	0–8	2·7	51·8	5·7	43·3	1·58	4·6	7·20	29·5	1·55
	A	8–17	2·0	51·9	5·3	42·8	1·64	5·0	7·75	25·1	1·19
	A	17–26	1·1	52·3	4·4	43·3	1·69	5·6	7·80	23·7	1·18
	A	26–35	1·3	52·9	4·2	42·9	1·44	3·3	7·80	22·4	1·24
	Bw1	35–41	1·0	49·7	8·5	41·8	1·63	1·5	7·85	16·7	1·09
	Bw1	41–48	1·1	54·4	3·7	41·9	1·59	11·3	7·85	20·6	1·83
	Bw2	48–54	1·1	59·8	0·4	39·8	1·63	56·3	7·90	12·7	0·54
	Bw2	54–60	1·0	48·1	2·0	49·9	1·50	67·7	7·95	9·3	0·46

* Expressed on an organic matter and $CaCO_3$-free basis.
† Cation exchange capacity.
‡ Dithionite-extractable iron.

beach or lagoonal sand derived from the subjacent reef. The vertic properties of Bar-3A which are not found in Bar-4B probably result from the higher clay content in the former soil (∼ 66 per cent vs. ∼ 43 per cent).

Clay minerals in all horizons of both soils are largely randomly interstratified kaolinite–smectite with trace amounts of quartz in some horizons (Table III). The X-ray diffraction pattern of this mineral (Figure 2) is
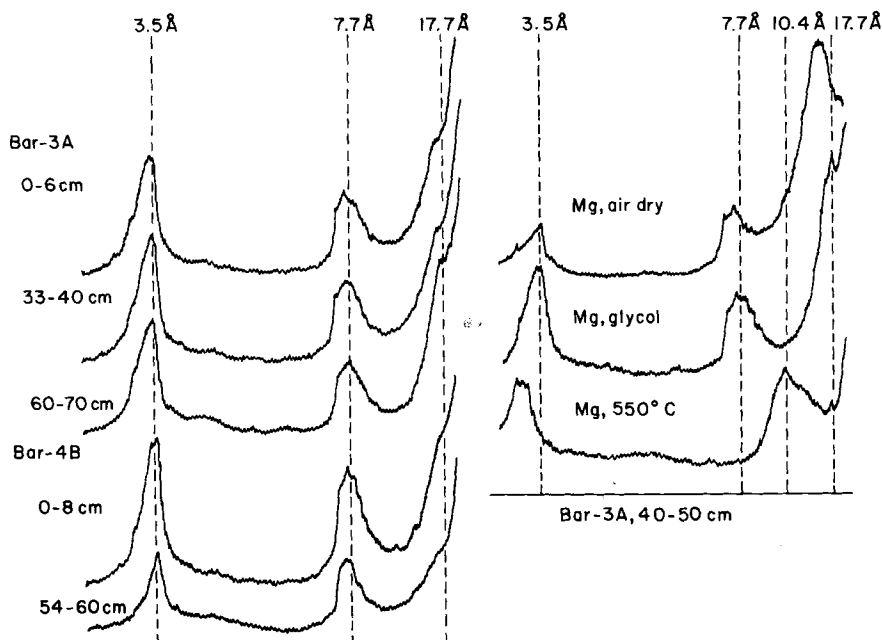

Figure 2. Representative X-ray diffractograms of clays from Barbados soils showing variability as a function of depth in the profile and various treatments

characterized by a peak centring on about 7·4–7·7 Å, but with a sharp dropoff on the high angle side and a gradual tail on the low angle side (Cradwick and Wilson, 1972). A second peak is present at about 3·5–3·6 Å. Upon heating to 550°C for two hours, a peak appears at about 10 Å at its highest point, but with a tail to the low-angle side. Yerima et al. (1985) also noted these patterns for interstratified kaolinite–smectite, and the patterns fit calculated peak positions for this mineral (Reynolds, 1980). Data on the cation exchange capacities (CEC) of the soils support this mineralogical identification, since CEC values of 9–45 meq/100 g are higher than would be expected for a pure kaolinite-rich soil of comparable clay content (e.g., < 10 meq/100 g) (Table II). The X-ray pattern of halloysite is easily confused with that of interstratified kaolinite–smectite, but the measured cation exchange capacities are too high for halloysite, which typically has values of 5–10 meq/100 g (Grim, 1968). The significantly lower cation exchange capacities for Bar-4B are probably due to the lower clay contents in this profile (Table II). Ahmad and Jones (1969b) also found interstratified kaolinite–smectite in Barbados soils. Based on Reynold's (1980) calculated values of kaolinite content as a function of peak positions, these interstratified clays are 75–90 per cent kaolinite (Table III).

X-ray diffraction analyses of the silt fractions of Barbados soils confirm the findings of Ahmad and Jones (1969b) that this particle size is dominated by quartz with lesser amounts of plagioclase (Table III). Examination of very fine sand-sized particles under a petrographic microscope and X-ray diffraction analyses of whole-sand samples indicate an assemblage dominated by quartz, with lesser amounts of pyroxenes, magnetite, and plagioclase. Fine-grained haematite coats some of the sand grains.

EVALUATION OF POSSIBLE PARENT MATERIALS

Coral residuum

Insoluble residue from the coral reef limestone was thought by Harrison and Anderson (1928), Vernon and Carroll (1965), and Ahmad and Jones (1969b) to be the main parent material for Barbados terrace soils. It is clear that reef carbonate has made some contribution to the soils, as there is a fairly high $CaCO_3$ content in the lower parts of both soils (Table II). Concentrations of CaO and to a lesser extent, Sr parallel the distribution of carbonate (Tables IV and V). Examination of the carbonate components of these lower horizons indicates that much of the $CaCO_3$ is detrital coral fragments rather than secondary $CaCO_3$, although secondary calcretes have been observed elsewhere on Barbados (James, 1972; Matthews, 1974; Harrison, 1977). These coral fragments are highly leached, however, because Sr concentrations, corrected for the non-carbonate fraction, are much lower than those found in unweathered Barbados corals (Cross and Cross, 1983). Some of the carbonate fragments near the tops of the profiles may have been added from modern beaches by aeolian processes.

A more important question is the contribution of the coral reef limestone to the non-carbonate component of the soils. The amount of carbonate dissolution required to produce the observed soil profiles entirely from

Table III. Mineralogical data for Barbados soils

Profile	Horizon	Depth (cm)	Sand* (0·05–2 mm)	Silt* (0·002–0·05 mm)	Clay* (< 0·002 mm)	Kaolinite† (%)
Bar-3A	A1	0–12	Q, p, pf, m	Q, PF	K–S,q	85
	A2	12–33	Q, p, pf, m	Q, PF	K–S,q	80
	A3	33–40	Q, p, m	Q, pf	K–S,q	80
	C	40–70	q, pm,	Q, pf	K–S,q	75–80
Bar-4B	A	0–35	Q, p, m	Q, pf	K–S,q	85
	Bw1	35–48	Q, p, m	Q, pf	K–S	90
	Bw2	48–60	Q, p, m	Q, pf	K–S	90

* Abbreviations: K–S = interstratified kaolinite–smectite; Qq = quartz; PF and pf = plagioclase; p = pyroxenes, m = magnetite. Upper case indicates abundant; lower case indicates moderate amounts or rare.
† Estimated from calculated peak positions of interstratified kaolinite–smectite in Table 4.5F of Reynolds (1980).

Table IV. Major element composition (wt.%) of Barbados soils (whole-soil samples), ashes from St. Vincent, and dust sampled on Barbados

Location	Profile	Horizon	Depth (cm)	SiO_2	Al_2O_3	Fe_2O_3*	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	MnO	LOI†
Barbados	Bar-3A	A1	0–6	49·5	15·8	6·97	0·87	4·23	0·23	0·17	0·82	0·23	0·22	21·30
		A1	6–12	50·1	16·0	6·98	0·87	5·31	0·25	0·15	0·82	0·20	0·22	19·71
		A2	12–22	52·0	16·5	7·10	0·82	4·09	0·23	0·15	0·85	0·17	0·22	17·66
		A2	22–33	51·1	16·8	7·29	0·77	4·35	0·23	0·14	0·87	0·15	0·24	17·64
		A3	33–40	52·9	16·9	7·33	0·76	3·70	0·22	0·14	0·87	0·15	0·24	16·51
		C	40–50	40·8	15·9	6·35	0·82	13·0	0·14	0·10	0·70	0·08	0·14	22·79
		C	50–60	35·1	13·8	5·34	0·75	18·3	0·12	0·09	0·59	0·06	0·07	26·08
		C	60–70	31·0	13·7	5·16	0·77	22·1	0·13	0·09	0·55	0·06	0·05	26·40
Barbados	Bar-4B	A	0–8	64·4	12·2	5·23	0·54	3·32	0·25	0·16	0·61	0·13	0·16	12·82
		A	8–17	67·0	11·6	4·98	0·52	3·36	0·24	0·14	0·58	0·13	0·17	11·11
		A	17–26	66·6	11·0	4·68	0·49	3·68	0·21	0·12	0·55	0·11	0·16	12·18
		A	26–35	68·1	11·4	4·83	0·43	2·56	0·22	0·13	0·58	0·10	0·14	11·06
		Bw1	35–41	66·8	13·3	5·31	0·36	1·45	0·16	0·15	0·60	0·07	0·10	11·72
		Bw1	41–48	65·0	12·4	4·88	0·37	3·93	0·15	0·15	0·53	0·07	0·11	12·28
		Bw2	48–54	37·8	7·2	2·64	0·53	24·7	0·14	0·13	0·26	0·10	0·03	26·39
		Bw2	54–60	30·4	5·33	1·95	0·60	32·3	0·16	0·12	0·21	0·11	<0·03	29·6
St. Vincent	1902 unweathered ignimbrite			54·8	18·5	9·00	3·81	8·91	3·36	0·56	0·93	0·14	0·19	0·14
	unweathered Holocene ignimbrite			54·4	18·7	8·79	3·86	3·83	3·21	0·59	0·95	0·14	0·18	0·59
	unweathered Pleistocene scoria			51·1	17·1	10·4	5·76	10·8	2·50	0·37	1·09	0·10	0·21	0·86
Barbados	dust samples‡			47·2	14·5	6·60	nd	nd	nd	nd	1·01	0·43	0·08	nd

* Total Fe as Fe_2O_3.
† Loss-on-ignition at 900°C
‡ SiO_2, Al_2O_3, Fe_2O_3, TiO_2, P_2O_5, and MnO are mean concentrations of 12 dust samples analysed by Glaccum (1978).

Table V. Trace element composition (ppm) of Barbados soils (whole-soil samples), ashes from St. Vincent and dust collected on Barbados

Location	Profile	Horizon	Depth (cm)	Rb	Sr	Y	Zr	Nb	Th
Barbados	Bar-3A	A1	0–6	22·3	220	29·6	122	6·3	3·62
		A1	6–12	17·6	254	27·6	115	5·3	3·68
		A2	12–22	19·5	223	28·6	119	7·6	3·72
		A2	22–33	19·1	203	28·2	119	9·4	3·77
		A3	33–40	17·3	180	29·9	125	7·7	3·94
		C	40–50	11·4	236	27·8	102	6·7	3·33
		C	50–60	10·8	224	22·5	85	7·6	2·76
		C	60–70	10·0	297	26·8	78	9·3	2·76
Barbados	Bar-4B	A	0–8	12·9	137	24·2	118	6·7	3·01
		A	8–17	12·3	144	25·6	106	8·4	2·94
		A	17–26	14·7	141	23·9	105	7·7	3·02
		A	26–35	17·2	110	22·2	117	6·8	3·06
		Bw1	35–41	13·0	64	19·1	117	6·7	3·37
		Bw1	41–48	6·9	143	25·3	98	5·9	3·25
		Bw2	48–54	4·7	729	12·5	62	4·6	2·00
		Bw2	54–60	7·2	886	6·6	42	2·9	1·64
St. Vincent	1902 ignimbrite			6·4	185	18·4	59	3·2	0·61
	Holocene ignimbrite			7·0	214	17·7	85	4·2	0·64
	Pleistocene scoria			2·8	205	17·2	89	5·3	0·44
Barbados	Dust samples*			nd	nd	40	167	nd	12·41

* Y and Zr data are mean concentrations of 12 dust samples analysed by Glaccum (1978); Th reported is mean concentration of 14 dust samples analysed by Rydell and Prospero (1972).

the limestone can be calculated from data on texture, carbonate content, bulk density and insoluble residue in coral. Pittman (1974) measured the porosity in 52 samples of aragonitic *Acropora palmata* corals on Barbados; his data indicate a mean porosity of $\sim$ 31 per cent. Because aragonite has a density of 2·93 g cm^{-3}, this yields an average bulk density of 2·02 g cm^{-3} for unaltered, aragonitic reef corals. Ahmad and Jones (1969b) measured insoluble residues in coral limestone of Barbados that ranged from 0·7–2·5 weight-per cent. Day (1983) found insoluble residues as high as 10 per cent, but less than 2 per cent in most samples. Differences in reported values may be the result of varying amounts of contamination by overlying soils (Comer, 1974). If we assume a typical value of 2 per cent insoluble residue with a bulk density of 2·02 g cm^{-3} in the reef limestone underlying the profiles, this yields 0·04 g of non-carbonate material in 1·0 cm^3 of reef limestone. The mass of non-carbonate material in the two soil profiles is calculated from the horizon thicknesses, bulk densities, and carbonate contents (Table II) using the profile summation method (Muhs, 1982). The results indicate that Bar-3A has $\sim$ 92 g of non-carbonate material in a 1 cm^2 cross-section column of soil extending from the top of the profile to the contact with the reef limestone. Bar-4B, 60 cm thick, has $\sim$ 79 g of non-carbonate material in a 1 cm^2 soil column. Thus, if Bar-3A were derived entirely from the accumulation of insoluble residues, $\sim$ 23 m of reef dissolution would be required with no subsequent erosion of the residuum. For Bar-4B, $\sim$ 20 m of reef dissolution would be required. Using these estimates of the amount of surface lowering and the uranium-series dates of $\sim$ 125,000 y and $\sim$ 180,000 y for the Rendezvous Hill and Durants reefs, respectively, we calculate long-term average reef lowering rates of 0·1–0·2 mm y^{-1}. Spencer (1985) reports reef lowering rates at inland sites on Grand Cayman Island (which has a climate similar to that of Barbados) of 0·06–0·69 mm y^{-1}. These data suggest that Barbados soils could in fact have formed as residual products of reef dissolution. However, two factors cause us to reject this hypothesis. First, Spencer's (1985) measurements were made over a period of less than two years; even if his data are accurate, they are probably not representative of reef dissolution rates over the entire late Quaternary. Second, such enormous amounts of carbonate removal by solution should be displayed as highly degraded terraces and karst in the geomorphology of the island. The terraces are not

degraded but rather have well expressed crests, backreef zones, and forereef slopes. Local relief (i.e., difference in crest elevations) between the geomorphologically well-expressed Rendezvous Hill and Durants terraces is only on the order of ~ 6 m (Bender et al., 1979), far less than the amount of hypothesized carbonate removal. In addition, the uppermost horizontal reef facies changes are still well exposed in roadcuts (Mesolella, 1967; Mesolella et al., 1970). Karst is well developed only on the older (> 460,000 y) terraces of the island (Harrison, 1977; Hopley, 1982; Day, 1983). Finally, the more arid coastal locations are characterized by subsurface accumulation of secondary carbonate, rather than karst (James, 1972; Harrison, 1977).

One could argue that our soils represent accumulations of insoluble residues that were derived from reef dissolution over a large area of terrace surface. If this were true, one would expect thick (70–100 cm) soils in only a few low-lying areas and little or no soil on most of the rest of the reef surface. Field observations by ourselves and Vernon and Carroll (1965) do not support this, as the terraces have gentle slopes of 3–5 per cent over most all of their surfaces, and soil thicknesses are relatively uniform on the terraces. The soils we sampled are of average thickness and do not represent 'cumulic' landscape positions. Thus, we conclude from our calculations and the geomorphic evidence that carbonate dissolution and accumulation of insoluble residues play relatively minor roles in the genesis of Barbados soils.

Tertiary bedrock

Harrison and Anderson (1928) suggested that upper terrace soils were derived largely from the insoluble residues of the reef limestone, but that as uplift and stream dissection proceeded, the underlying Tertiary rocks contributed sediment to the living reef offshore via transportation in fluvial systems. We can test Harrison and Anderson's (1928) hypothesis by examining the sediments found on modern Barbados beaches and offshore areas and the mineralogy of the soils themselves. Data from Randall (1968), Macintyre (1970), and our own observations indicate that most modern beach and offshore sediments are largely (usually > 90 per cent) composed of $CaCO_3$ derived from fossil or modern coral reefs and other calcareous organisms. Thus, little sediment derived from Tertiary bedrock units is contributed to the offshore area via modern fluvial systems, even though the drainage network is well developed (Fermor, 1972). Macintyre (1970) found the non-carbonate fraction of modern offshore sands to be composed of quartz, plagioclase, glass, and heavy minerals such as olivine, hypersthene, augite, hornblende, and magnetite. He interpreted the plagioclase, glass, and heavy minerals as indicative of a volcanic source and concluded that these minerals were derived from the island of St. Vincent to the west. The most likely source for the quartz is the Tertiary quartzose turbidites of the basal complex in the Scotland District on Barbados (Speed and Larue, 1982; Larue and Speed, 1983; Larue et al., 1985). We did find quartz as the main constituent in the sand fractions of the soils we sampled (Table III). Chemical data also support this observation. Most soils show a depletion of SiO_2 relative to immobile constituents (Al_2O_3, Fe_2O_3, and TiO_2) compared to their presumed parent materials (see review in Birkeland, 1984). We plotted ratios of weight-per cent SiO_2 to sums of Al_2O_3, Fe_2O_3, and TiO_2 (commonly referred to as R_2O_3) for the soils (whole-soil samples) and two of the hypothesized parent materials (Figure 3). The high ratios clearly show the importance of the quartz-rich sand fractions of both soils. Bar-3A, with ~ 20 per cent sand in most horizons, has ratios not much different from the hypothesized parent materials and Bar-4B, with its higher sand content (Table II), shows significantly higher ratios than those in the ash and dust. We suggest that this quartz is derived from the Tertiary basal complex, as quartz is absent in St. Vincent ash-fall sands (Harrison et al., 1980) and quartz that is found in airborne dust is largely silt-sized particles (Glaccum, 1978; Glaccum and Prospero, 1980). Thus, Tertiary quartzose turbidites may contribute perhaps 20 per cent of the non-carbonate fraction of Bar-3A but perhaps as much as 50 per cent of the non-carbonate fraction of Bar-4B, despite the fact that modern reef environments contain only minor amounts of quartz.

Airborne dust fall

Harrison (1977) suggested that a soil parent material on Barbados might be airborne dust, although he provided no tests of this hypothesis. Airborne dust has been collected on Barbados for a number of years and there now exists an extensive literature on the subject summarized by Prospero (1981) and Prospero and Nees (1986). Analysis of satellite imagery suggests that the dust is eroded from soils in arid zones of west Africa and is
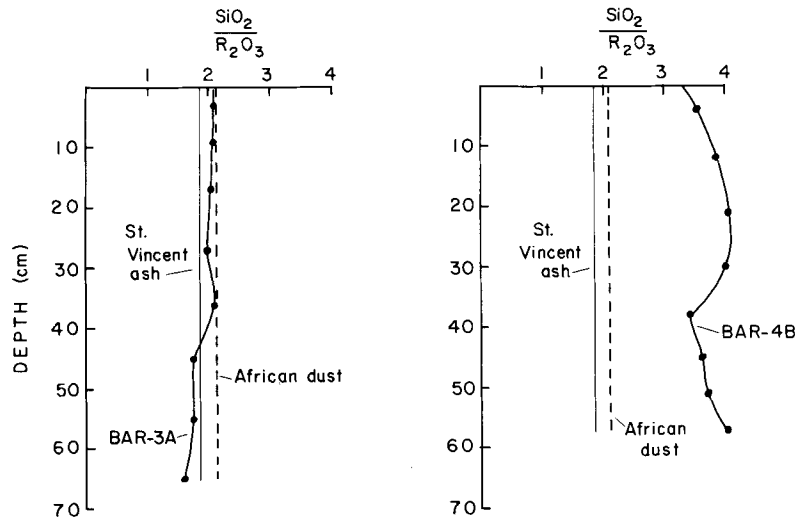
Figure 3. Plots of weight-per cent ratios of SiO_2 to R_2O_3 ($Al_2O_3 + Fe_2O_3 + TiO_2$) in Barbados soils (whole-soil samples) as functions of depth compared with similar ratios in St. Vincent ash and African dust collected on Barbados. Calculated from data in Table IV

carried across the Atlantic on the northeast trade-winds. Estimates of deposition rates in the Caribbean region range from 0·4–0·8 g cm^{-2} 1000 y^{-1} (Prospero and Carlson, 1972; Glaccum, 1978) and may have been higher during glacial periods (Kolla et al., 1979). These rates are lower by a factor of 3 to 10 than dustfall rates for the southwestern United States summarized by Muhs (1983).

It is difficult to assess the possible importance of airborne dust as a parent material from mineralogical data alone. Data in Delany et al. (1967) and Glaccum and Prospero (1980) indicate that mica is the dominant mineral in < 2 µm dust samples derived from Africa that are transported to Barbados. We found no mica in any of our clay samples, nor did Ahmad and Jones (1969b) in the Barbados terrace soils that they analysed. It is possible that airborne mica deposited on Barbados has altered through subsequent weathering. If this were true, we might expect to see transitional products such as interstratified mica–smectite or mica–kaolinite, at least near the tops of the profiles. On the other hand, it is possible that mica alteration rates are greater than dust fall rates. There is some evidence for possible airborne dust additions in the form of quartz, present in both the clay and the silt fractions (Table III). Volcanic ash from St. Vincent is andesitic and lacks quartz (Diller and Steiger, 1902; Hay, 1959, 1960; Harrison et al., 1980). The presence of clay and silt-sized quartz is evidence of either weathering to fine sizes of sand-sized quartz derived from Tertiary bedrock, or aeolian deposition from the African sources, because quartz dominates the silt fraction of the dust (Delany et al., 1967; Prospero et al., 1970; Glaccum and Prospero, 1980). Similarity of oxygen isotopic compositions of Barbados dust-derived quartz and quartz from soils on the Bahamas and the U.S. east coast suggests a common aeolian origin (Syers et al., 1969).

Volcanic ash fall

Harrison (1919), Vernon and Carroll (1965), and Ahmad and Jones (1969b) all suggested that volcanic ash falls from the island of St. Vincent (Figure 1) might have influenced soil genesis on Barbados to some degree, but none of these workers regarded this as the primary parent material. Eruptions from St. Vincent which deposited ash on Barbados are documented during historic time in 1784, 1812, 1902–1903, and 1979 (Diller and Steiger, 1902; Flett, 1902; Hay, 1960; Harrison et al., 1980).

The presence of interstratified kaolinite–smectite as the dominant clay mineral in Barbados soils (Figure 2) is suggestive of a volcanic ash parent material, as this clay mineral seems to be a common alteration product of volcanic materials in tropical climates. Kaolinite–smectite is reported in clays weathered from volcanic ash in Mexico (Schultz et al., 1971), in soil clays derived from basalt in Burundi (Herbillion et al., 1981) and in some El

Salvador Vertisols derived from Quaternary andesitic alluvium and volcanic tuff (Yerima *et al.*, 1985). All of these are tropical environments characterized by relatively high mean annual temperatures, although the amount of precipitation varies among study areas. Mineralogical data from the sand fractions of Barbados soils also support a volcanic origin (Table III). Pyroxene and magnetite (and to a lesser degree, plagioclase) are found in the sand-size fractions at all depths in both profiles. Plagioclase is also abundant in the silt fractions. There was no unequivocal evidence of sand-sized glass in the soils, although sand-sized glass from historical eruptions at St. Vincent is reported to have landed on Barbados (Harrison *et al.*, 1980). It is possible, however, that most of the glass which has fallen on Barbados in late Quaternary time has already altered to clay, as glass alteration rates seem to be rapid in humid, tropical environments (Hay, 1959, 1960; Davies *et al.*, 1979).

Evidence of soil parent materials from trace element geochemistry

The mineralogical data indicate that Barbados soils are polygenetic in that more than one parent material has contributed to pedogenesis. The sand fraction has evidence for contributions from Tertiary bedrock and to a lesser extent from volcanic ash. Quartz in the silt fraction may be derived from weathering of the Tertiary bedrock sands or from African dust. Based on the similarities between the clay mineralogy of Barbados soils we report here and the clay mineralogy of tropical soils derived from volcanic parent materials reported by other workers, there is a suggestion that Barbados soil clays may be derived from alteration of St. Vincent ashfalls. The possibility remains, however, that some of the interstratified kaolinite–smectite may be derived from alteration of aeolian mica from Africa.

In a further effort to distinguish between St. Vincent (volcanic) and African (dust) sources, we conducted analyses for the immobile trace elements Ti, Y, Zr, and Th in order to develop parent material signatures. The method is based on the assumption that these elements are immobile hydrolysates under the pH–Eh conditions found in soils and thus will remain close to the site of release upon weathering (Mason and Moore, 1982). Because of mass losses during weathering, it is not possible to compare concentrations of immobile trace elements in soils with concentrations in hypothesized parent materials. Therefore, we developed parent material signatures by the use of trace element ratios. The two ratios that are most distinctive for volcanic ash from St. Vincent and dust from Africa are Zr/Th and Ti/Y. Similar ratios for Barbados soils can be compared with the hypothesized parent materials because these ratios should not be affected by mass losses accompanying weathering.

Results of these calculations show that the two parent materials (volcanic ash and African dust) have ratios that do not overlap when the sample variability ($\pm 1\sigma$) is considered (Figures 4 and 5). For both soils at all
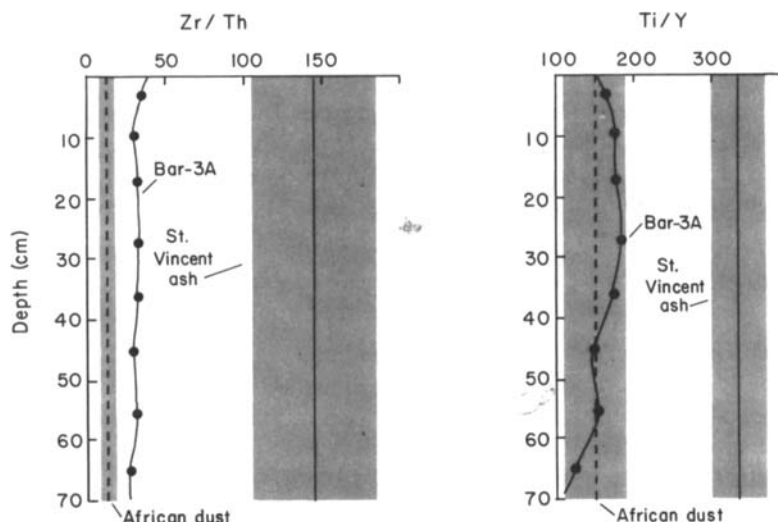

Figure 4. Plots of Zr/Th and Ti/Y as functions of depth in Bar-3A (whole-soil samples) compared to mean values for St. Vincent ash samples and Barbados dust samples. Shaded area represents variability in the ash and dust samples ($\pm 1 \sigma$). Calculated from data in Tables IV and V
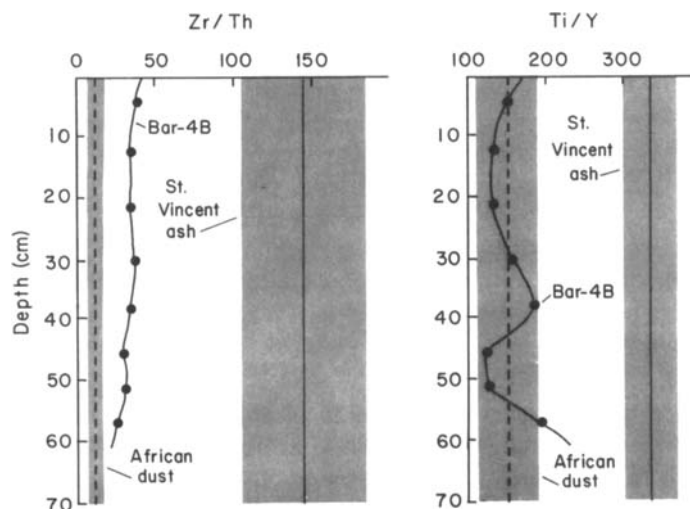
Figure 5. Plots of Zr/Th and Ti/Y as functions of depth in Bar-4B (whole-soil samples) compared to mean values for St. Vincent ash samples and Barbados dust samples. Shaded area represents variability in the ash and dust samples ($\pm 1 \sigma$). Calculated from data in Tables IV and V

depths Zr/Th ratios plot in between the two parent materials, but closer to the values for African dust. For Ti/Y, soil values are within the range of values for African dust and significantly lower than St. Vincent ash values. Analyses were conducted on whole-soil samples, but contributions from the coral reef fragments ($CaCO_3$) and the Tertiary bedrock (quartz) should not affect the ratios as Zr, Th, Ti, and Y are not found in reef $CaCO_3$ or quartz to any significant degree. We conclude from these data that both dust and ash may have contributed to the soil clays, but African dust may have been more significant as a parent material. Thus, much of the interstratified kaolinite–smectite has probably been derived from alteration of aeolian mica and a lesser amount has been derived from alteration of volcanic glass and plagioclase. The lack of either mica or glass in any of our soils, even near the tops of the profiles, indicates that rates of weathering of these materials exceed rates of dust fall or ash fall.

We can test whether or not our conclusion about the importance of dust as a parent material is reasonable by a calculation of the mass of clay in a 1 cm^2 soil column and comparison with the Barbados dust fallout rates cited earlier. Bar-3A, with its thicker profile, has probably experienced less erosion and a profile summation calculation using data in Table II yields $\sim 66\ g\,cm^{-2}$ of clay. The uranium-series date on this terrace is $\sim 125,000$ y (Mesolella et al., 1969), so with a dust flux of $0.4–0.8\ g\,cm^{-2}\ 1000\ y^{-1}$, $50–100\ g\,cm^{-2}$ of material should have accumulated in 125,000 y. Based on data in Kolla et al. (1979), the estimated dust fall rates are minima, so a significantly greater amount of material could have accumulated than the amount calculated. Because chemical weathering results in mass losses, we conclude that African dust could indeed have supplied most of the parent material for the soil clays.

CONCLUSIONS

These studies of Barbados soils permit several conclusions regarding soil genesis. Based on soil and parent material mineralogy and major and trace element geochemistry, we conclude that the parent materials for Barbados terrace soils are a combination of airborne dust from Africa, volcanic ash originating on the island of St. Vincent, and quartz derived from the Tertiary bedrock on the island itself. Of these, African dust appears to be the most important source for soil clays. The processes of dust fall and ash fall are gradual enough that weathering in the warm, humid environment of Barbados keeps ahead of accumulation of ash materials such as glass and aeolian mica. Airborne dust arriving from African dust storms on the trade winds may also contribute some clay and silt-sized quartz to the soils. Sand-sized quartz is probably delivered to growing reefs from the Tertiary basal complex via river systems; since quartz is relatively unreactive in subaerial environments it

persists through soil development after uplift of the reef. The main contribution of the coral reef itself to soil development seems to be detrital carbonate found in the lower parts of the profiles; calculations of soil mass and carbonate residue do not support a residual origin for the soil clays.

Our conclusions about parent materials for Barbados soils have important implications for soil genesis on other limestone terrains in the region. Because African dust has been observed in the Caribbean region in a latitudinal belt extending from Miami, Florida to northern South America (Prospero, 1981; Prospero et al., 1981), it is possible that this aeolian sediment is the parent material for many soils and bauxites found on relatively pure limestones. Mineralogical and geochemical studies of other Caribbean island soils can provide a rigorous test of this hypothesis.

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