LANDSCAPE FORMATION AND SOIL GENESIS IN VOLCANIC PARENT MATERIALS IN HUMID TROPICAL LOWLANDS OF COSTA RICA

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LANDSCAPE FORMATION AND SOIL GENESIS IN VOLCANIC PARENT MATERIALS IN HUMID TROPICAL LOWLANDS OF COSTA RICA

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STELLINGEN

Het Atlantische laagland van Costa Rica is dynamisch in meer dan alleen geologisch opzicht: overstromingen en aarbevingen bekonkurreren de voorpaginas van kranten met landbezettingen en de opkomst of ondergang van landbouwaktiviteiten.

Landchapsvorming in het Atlantiche laagland van Costa Rica wordt bepaald door korte perioden waarin grote hoeveelheden sediment worden afgezet, gevolgd door lange perioden van betrekkelijke rust met geringe erosie.

- dit proefschrift

Bij een regenval van meer dan 3000 mm per jaar en en gemiddelde temperatuur van 25 °C, kan zich uit andesitisch vulkanisch zand binnen 2000 jaar een bodem ontwikkelen die geschikt is voor veelzijdig gebruik.

- dit proefschrift

De mineralogie van bodems onstaan uit andesitisch vulkanisch zand of lava in de Atlantische zone van Costa Rica verschilt niet van die in de meeste andere vulkanische bodems onder vergelijkbare omgevingsfactoren.

- dit proefschrift
- *Lowe, D.J. 1986. Controls on the rates of weathering and clay mineral genesis in airfall tephras: a review and New Zealand case study. p. 265-330. In S.M. Colman and D.P. Dethier (ed.) Rates of chemical weathering of rocks and minerals. Academic Press, Orlando, FL.*
- *Quantin, P. 1992. Les sols de l'archipel volcanique des Nouvelles-Hébrides (Vanuatu), étude de la pédogenese-initiale en milieu tropical. PhD thesis, ORSTOM, Paris, France.*

De vraag of het naast elkaar voorkomen van de mineralen gibbsiet en halloysiet in een zelfde bodemprofiel een gevolg is van verschillende klimaten waaronder bodemvorming zich heeft afgespeeld of van verschillen in verweringsmilieu binnen het profiel t.g.v. verschillende vochtstromingen, kan niet eenduidig worden beantwoord uit deze studie.

- dit proefschrift
- *Jongmans, A.G. 1994. Aspects of mineral transformation during weathering of volcanic material, the microscopic and submicroscopic level. PhD thesis, Agricultural Univ., Wageningen, the Netherlands.*

Zonder de grote sedimentaanvoer als gevolg van aktief vulkanisme had de kustlijn van het Atlantiche laagland van Costa Rica waarschijnlijk niet de rechte vorm gehad die nu de kustlijn kenmerkt.

- dit proefschrift

Het gebruik van een massabalans gebaseerd op inmobiliteit van titaan of een ander stabiel element in de bodem is omgeven met vele onzekerheden; niettemin geeft het een verhelderend beeld omtrent verwering en nieuwvorming in een bodem.

- dit proefschrift
- *Chadwick, O.A., G.H. Brimhall, and D.M. Hendricks. 1990. From a black box to a grey box - a mass balance interpretation of pedogenesis. Geomorphology 3:369-390.*

Gezien de enorme sociale en ekonomische konsekwenties die vulkaanuitbarsting met zich mee kunnen brengen en het belang van kennis over de eruptiegeschiedenis om uitbarstingen te kunnen voorspellen, zijn de meeste Costaricaanse vulkanen verbazingwekkend slecht bestudeerd.

Door de aanwezigheid van macroporien in gronden op vulkanische moedermaterialen in het Atlantische laagland van Costa Rica zijn voor simulatiemodellen van de vochthuishouding voor gewassen naast metingen van vochtspanningen in het -5.0 tot -60 kPa bereik, ook metingen van verzadigde doorlatendheid onder veldomstandigheden nodig.

- *Leummens, H., J. Bouma, and H.W.G. Booltink. 1995. Interpreting differences among hydraulic parameters for different soil series by functional characterization. Soil Sci.Soc.Am.J. 59:344-351.*

Het simpelweg op verschillende hoogtes grond in een zakje stoppen, de grondmonsters analyseren en er een wetenschappelijk verhaal over schrijven is heel iets anders dan het bestuderen van een klimosekwentie waarin moedermateriaal en leeftijd van de bodem zoveel mogelijk gelijk dienen te zijn.

- *Grieve, I.C., J. Proctor, and S.A. Cousins. 1990. Soil variation with altitude on volcan Barva, Costa Rica. Catena 17:525-534.*

Wie dagen lang in een door muggen vergeven tropisch regenwoud vertoeft, kan er begrip voor opbrengen dat eigenaren van zulke bossen ze liever omhakken en er weiland van maken.

Het overtuigen van boeren om duurzaam te boeren zal pas echt lukken wanneer er op duurzame wijze ook "goed geboerd" wordt.

Stellingen behorend bij het proefschrift "Landscape formation and soil genesis in volcanic parent materials in humid tropical lowlands of Costa Rica", André Nieuwenhuyse, Wageningen, 29 mei 1996.

CONTENTS

CHAPTER 1. INTRODUCTION

In 1986 the Agricultural University Wageningen set up a research project in the humid tropical Atlantic lowland of Costa Rica with the objective to study rational use of natural resources. Parts of this project deal with the study of geomorphologic processes which may benefit or constrain land use, and of soil, one of the resources itself.

Topography of the Atlantic lowland is mainly flat, although part of the active Irazú and Turrialba volcanoes, as well as of the Tertiary Talamanca cordillera were included in the study area. Due to the low population density and the availability of land suited for cultivation in cooler and dryer parts of the country, during a long time the permanently hot and humid climate made the area little attractive for colonization. Large scale occupation started only at the end of the 19th century, and accelerated after 1960. Today, significant forest areas remain only in protected areas.

During the first phase of the programme, a geomorphological and soil inventory was carried out. Landscape changes were observed to occur relatively rapidly and often appeared to be related to volcanic or tectonic activity, as well as to sea level changes. It was felt that further research on this topic was scientifically interesting and necessary within the scope of the research programme. Especially knowledge on the magnitude and recurrence interval of the events was important to evaluate sustainability of land use. Important aspects of sustainable land use are also the time necessary to form a well developed soil from a fresh parent material, and the amount of nutrients released by weathering.

Except in the southeastern part, soils formed on volcanic parent materials occur throughout the study area. Such soils were recently placed at the order level in the Soil Taxonomy classification system as Andisols (Soil Survey Staff, 1992). Their formation starts by rapid weathering of an assemblage of easily weatherable minerals usually present in volcanic ejecta, such as volcanic glass, olivine, plagioclase and pyroxene. At the same time, secondary weathering products are typically short-range order minerals such as allophane and imogolite, and Al- and Fe-humus complexes. Due to the presence of these materials these soils possess unique physical and chemical properties. In this thesis, aspects of soil formation on volcanic parent materials are studied.

Earlier, two related dissertations were published. Veldkamp (1993) studied organic carbon contents of soils developed on volcanic parent materials as a function of land use and time after deforestation. He concluded that when forest was converted to low-productive pasture soil organic carbon decreased with about 30 to 60 Mg C ha^{-1} during the first 20 years after deforestation. The presence of significant Al-organic matter complexes, which stabilize strongly soil organic carbon, was thought to be the main reason for the relatively low decline in soil organic matter after forest clearing, compared to other regions. Furthermore, these losses could be reduced by more than 50% when introducing high-productive grass species. Jongmans (1994) studied some of the phenomena found in profiles investigated in the present study using micromorphological techniques. He focused on the presence of 2:1 layer silicates in soil formed on andesitic sand under well drained conditions, a weathering environment where their presence is difficult to explain. The presence of these minerals was found to be a consequence of inheritance from the parent material. Furthermore, formation of clay coatings in volcanic soils in the study area was studied. Such coatings appeared to be a common feature in these soils, especially in deeper soil horizons. They result from co-precipitation of Al and Si, liberated upon weathering of primary minerals. Differences in environmental conditions at a micro-scale leads to formation of different coatings within the same soil horizon. Desilication gradually changes chemical and mineralogical composition of the coatings. In the study area, gibbsite seems to be the final weathering stage. Biological activity often was observed to destroy them towards the soil surface.

This thesis focuses on the relationships between geomorphology and rates of soil formation, with some emphasis on the lower alluvial plains and coastal areas. The structure of this thesis is as follows: the second chapter gives a description of the study area, in which geological setting, climate, land use and the most important soils of the area are discussed briefly. Chapters 3 and 4 highlight the influence of volcanism on respectively coastal and alluvial plain formation. Chapters 5 and 6 describe in detail soil formation in a Holocene beach ridge chronosequence on andesitic sand. A more quantitative approach of soil formation is presented in chapter 7, in which elemental gains and losses induced by soil formation are estimated for several soil profiles formed on volcanic parent material. The final chapter is a synthesis, in which I discuss the results of this study in the framework of the specific geological and environmental factors in the area.

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CHAPTER 2. GENERAL INFORMATION ABOUT GEOLOGY, CLIMATE, LAND USE, AND SOILS OF THE STUDY AREA

2.1. Geology and geomorphology

Geologically three main units can be distinguished: (1) the lowlands bordering the Caribbean Sea, which form the main part of the studied area and are filled up with alluvial and marine sediments, (2) the young Central cordillera composed of active volcanoes, of which the northern slopes of the Irazú and Turrialba volcanoes were studied, and (3) the northern part of the Talamanca cordillera in which old sedimentary and volcanic rocks occur, limiting the area in the southeast (Figure 1). Elevation varies from sea level to about 2500 m on the slopes of the Central cordillera and about 700 m in the Talamanca cordillera.

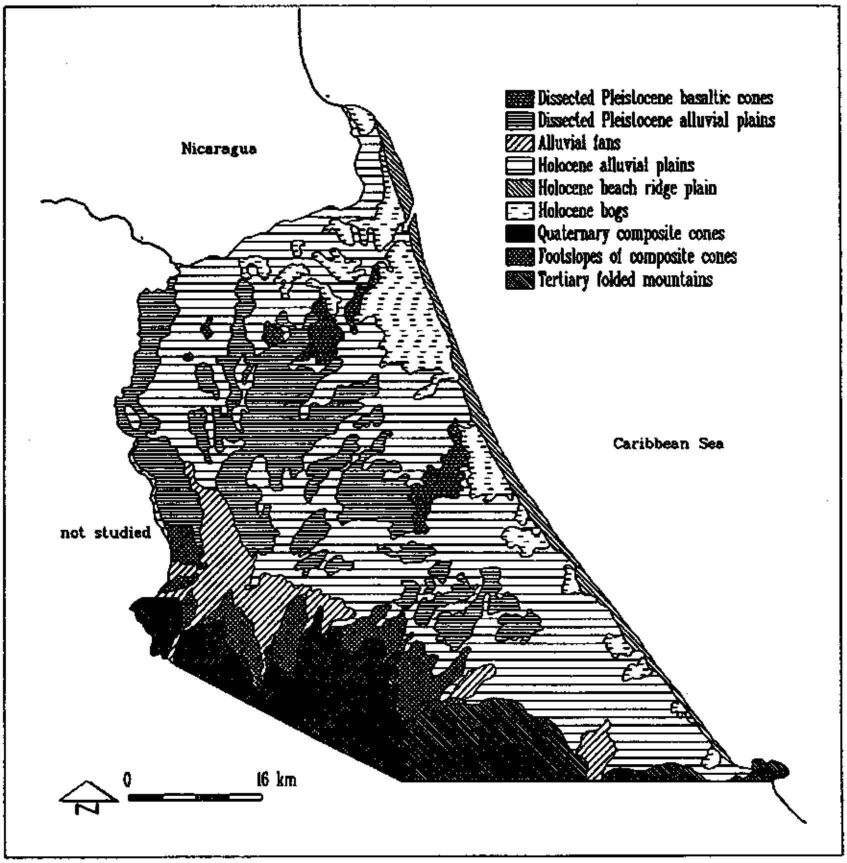

Figure 1: Main geomorphological units of the Atlantic Zone of Costa Rica.

The Caribbean Lowland. The extensive lowland forms the southeastern part of the Nicaragua depression, which stretches over a length of about 600 km from the Gulf of Fonseca northwest of Nicaragua, to the southeastern Caribbean coast of Costa Rica (see Figure 1 in chapter 3). It is a back-arc basin formed by crustal thinning as a consequence of the subduction of the Cocos plate beneath the Caribbean plate (Weyl, 1980; Seyfried et al., 1991).

Since the late Cretaceous, the basin has been filled up by dominantly marine sediments, while the Quaternary deposits are composed mainly of fluvio-volcanic sediments derived from the cordilleras. In some places, the combined thickness of the Tertiary and Quaternary material in the Limón Basin is about 8 to 10 km (Weyl, 1980; Seyfried et al., 1991). Maximal thickness of Quaternary deposits in the northeastern coastal area is about 500 m (pers.comm. RECOPE), which indicates a maximum average subsidence of 0.25 mm yr^{-1}. However, probably only part of the basin subsides at this rate, while the surrounding mountainous parts are being uplifted. The complex tectonic behaviour was illustrated during a strong earthquake in August 1991 (Richter 7.5), when the area around the city of Limón suffered a sudden uplift from 50 to 140 cm, while the alluvial plains northwest of the city subsided up to 40 cm (La Nación daily newspaper, 17-07-1991).

In the northeastern part of the basin some small, strongly dissected remains of basaltic cones rise up to maximally 300 m above the plain (Figure 1). They are the product of small fissure eruptions and are composed of olivine basalt (Sprechmann, 1984). One of these basalts was dated at 1.2 ± 0.4 million years (Bellon and Tournon, 1978), indicating that the hills are of early Pleistocene age.

Scattered throughout the area small hills with flat tops occur at 10 m (in the eastern part) to 25 m (towards the footslopes of the Central cordillera) above the actual river floodplains. They are thought to be the remains of an older Pleistocene terrace level. Sedimentary structures indicate that they consist mainly of fluvial deposits, although toward the Central cordillera also mudflow deposits contributed to their formation (Van Ruitenbeek, 1992). ^{14}C dating revealed that these deposits are >50,000 years old (GrN-18196, GrN-18594). Soils and saprolite which cover the deposits are somewhat less weathered than those on the basaltic hills, so they are estimated to be of middle or late Pleistocene age. Given their high position in the landscape, sediments are thought to have been deposited during a former high sea-level, possibly during the Eemian/Sangamonian interglacial ($\pm$125 ka ago).

Poorly drained depressions occur frequently occur between these hills. A 5 m thick peat deposit in one of them, 10 km east of the Río Jiménez village, was dated at the base at 4410 ^{14}C-years (GrN-17320). Height differences in the landscape during the glacial period and early Holocene (when the sea-level was much lower) must have been much more pronounced, and poorly drained areas were probably rare.

Holocene alluvial fan and plain deposits make up about 55% of the study area. As indicated by ^{14}C dating, most, if not all, of the alluvial plain deposits are young and probably have been deposited during the last 6000 years when the rapid sea-level rise after the last glacial period, slowed down. The oldest ^{14}C ages obtained were in a stratified clay loam deposit in which a moderately deep soil had developed. Organic remains underlying the soil were dated at 5310 ± 50 ^{14}C yrs BP (GrN19438), indicating that the soil must be younger. More developed soils on similar deposits have not been observed by the author. Inundations occur regularly in the alluvial fan areas and in the alluvial plain, especially at <20 m above sea level.

North of the city of Limón up to the Nicaraguan border the coastline is almost straight. The beach is bordered by a sandy beach ridge plain up to 3 km wide and <6000 years old, while further landward extensive Holocene peat swamps overlie marine sand deposits, probably deposited as beach ridges or in a coastal lagoon or deltaic environment (Battistini and Bergoeing, 1984; chapter 3 of this thesis). The coastline must have fluctuated considerably during the recent geological past. De Jong (1994) presents data on a brackish environment about 17 km inland of the actual coastline. Shells in these deposits at 4 to 6 m below actual sea level, were dated at 34,000 ^{14}C years BP (GrN-19968). Water well drilling (SNE unpublished data) revealed that similar shell deposits occur in a much wider environment, strongly suggesting that the coastline was indeed different from the actual one.

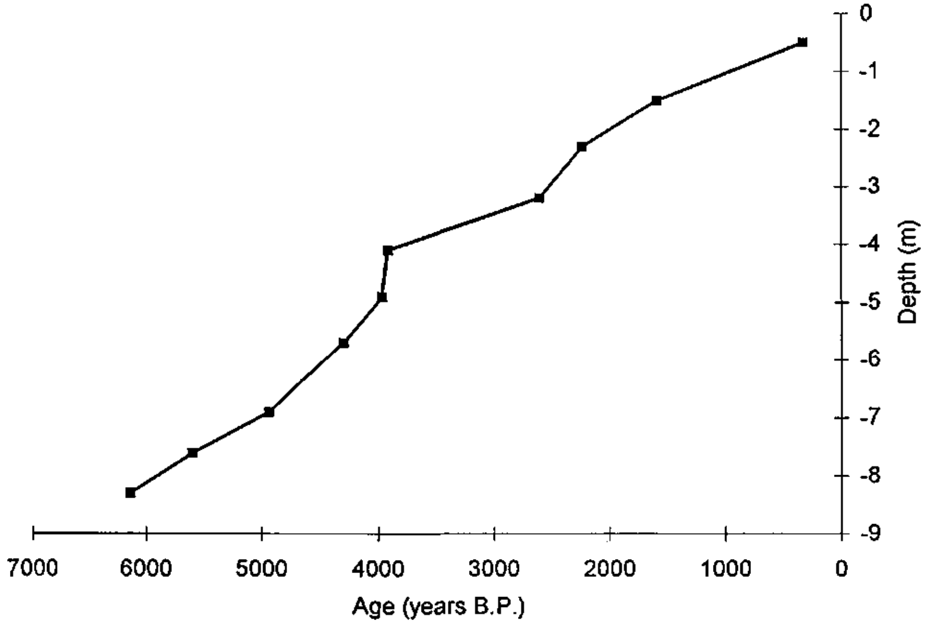

Figure 2: **^{14}C dating of peat deposits at the transition with a basaltic hill "Cerro del Coronel" (De Jong, 1994).**

Peat growth started when the rapid sea-level rise during the early Holocene slowed down about 6000 years ago. The thickest (about 10 m) peat deposit in the area was found at about 11 km from the actual coastline and its transition with sand of probably marine origin was dated at 6285 ^{14}C years BP (GrN 17566, see also chapter 3). At 5 km from the coastline the transition of a 2.3 m thick peat deposit with the underlying marine sand was dated at 3460 ^{14}C years BP (GrN 17566), and still closer to the coastline (1.2 km) the base of a 65 cm thick peat deposit was dated at 1065 ^{14}C years old (GrN 17567). De Jong (1994) dated the transition between this peat deposit and one of the basaltic hills in the area. His data (Figure 2) indicate that peat growth proceeded more or less at the same rate as the Holocene sea-level rise (Van der Plassche, 1982; Fairbanks, 1989).

The Central cordillera. This cordillera is composed of the irregularly formed active stratovolcano complexes Platanar - Porvenir (2267 m), Poás (2722 m), Barva (2906 m), Irazú (3423 m) and Turrialba (3328 m). The northern flanks of the Turrialba and Irazú are the only parts of the cordilleras included in the study area. The edifices were built during the Pleistocene on older ignimbrite- and lava flows (Weyl, 1980). In between the Barva and Irazú, the remnants of a large volcanic complex denominated Zurquí are located. A lava of this edifice was dated at 0.5 million years (Bellon and Tournon, 1978), indicating that, in spite of its advanced stage of erosion, rocks are relatively young. Historic records indicate activity of Poás, Irazú and Turrialba, while information of the other two volcanoes is incomplete and not reliable (Alvarado, 1989). Nevertheless, ^{14}C dating and stratigraphy of ash deposits indicate that also these volcanoes were active during the Holocene (E. Fernández, pers.comm.). Although the dominant northeastern winds deposit ash mainly at the southern and western side of the volcanoes (Melson et al, 1985), also on the northern flanks thick ash deposits were found at altitudes >600 m. Most of the recent eruptions involve pyroclastic material, while lava flows are scarce and mostly limited to the crater zone (Barquero, 1977; Prosser and Carr, 1987, Reagan, 1987). However, several large lava flows younger than 20,000 years have been identified and dominate part of the volcano flanks. Within the study area at least two of such flows from the Turrialba volcano are found: one of them about 5 km southeast of Guácimo was dated at $<18{,}190$ ^{14}C yrs BP (GrN18942). The second and largest of them was estimated by Reagan (1987) to be about 2000 years old. The transition between the Central cordillera and the lowland is gradual, and extensive alluvial fans have been formed (Kesel and Lowe, 1987).

The Talamanca cordillera. The largest and highest mountain range of Costa Rica is the Talamanca cordillera, which stretches southeast of the Central cordillera and reaches into Panama. The geological structure of the cordillera is rather complex, being built up of folded Tertiary sedimentary rocks, with intercalated volcanic and Middle Miocene plutonic rocks (Seyfried et al., 1991).

During the Lower Tertiary an ocean basin extended throughout the region of the present Talamanca cordillera in northwestern direction into Nicaragua (Rivier, 1973). In this basin large masses of sediments were deposited. Until the Lower Miocene mainly marine sediments were deposited. During the Middle Miocene sedimentation in shallow water dominated, indicating beginning uplift of the area (Seyfried et al., 1991). During the Upper Miocene, deposition of marine sediments was limited to small areas such as around the actual city of Limón, while in other parts deposition of terrestrial material derived from the cordillera dominated. Uplift was accompanied by vigorous andesitic volcanism and intrusions. K-Ar age determinations of the plutonic rocks vary between 8 and 19 million years (Bellon and Tournon, 1978; Appel, 1990). However, no intrusions have been identified in the studied part of the cordillera. The most recent volcanic activity in the Talamanca cordillera probably occurred during the Pliocene, as evidenced by the remains of some volcanic edifices. Volcaniclastic conglomerates, mudflows, and sandstones of the Suretka Formation lie as erosion products on the northern and eastern slopes, and are generally believed to be the youngest rocks of the Talamanca cordillera. However, in the northwestern part of the cordillera volcanic rocks are found on top of the Suretka formation. Because no eruption centre could be established, they may be related to early volcanism of the Central cordillera and were subsequently uplifted (see also Wielemaker and Vogel, 1993).

The morphology reflects resistance of the various formations against erosion: soft greywackes and silt and clay stones form low relief areas with smooth slopes, in which frequent landslides occur, while sandstones or conglomerates form escarpments with steeper slopes. Major rivers as the Pacuare and Chirripó are deeply incised. The transition between the Talamanca cordillera and the lowland is abrupt, with rather small alluvial fan development.

2.2. Climate

2.2.1. General aspects

Due to the close proximity to the sea at any point, the weather in Costa Rica is determined mainly by atmospheric perturbations that originate on either the Caribbean Sea or the Pacific ocean. Three large scale systems are of importance (Portig, 1976; Herrera, 1985):

1: North - south movements of the Intertropical Convergence Zone (ITCZ), the zone in which northern and southern trade winds meet, generating large cloud systems. As a consequence, regions influenced by the ITCZ experience high rainfall (McIntosh and Thom, 1983). The influence of the ITCZ on the weather in Costa Rica consists of weak southwestern winds on the Pacific side that generate rainfall. From December till April, the ITCZ is located south of the Equator and does not influence weather in Costa Rica. First rains fall in the southwestern part of the country in April, and reach the northwestern part of the Pacific lowland at the end of May. Late November to early December, the ITCZ begins to loose its influence, initiating the dry season at the Pacific side of the country.

2: During the northern winter Central America is affected by strong, cool northeastern trade winds. From December until March these winds occasionally transport cold (polar) air masses towards Central America. The mountain systems which separate the Caribbean from the Pacific side of the country retain these air masses, causing rainfall during one or several days at the Caribbean side, while the Pacific side of the country remains dry. When the trade winds do not transport these cold cloudy air masses, they generate dry weather also in the northern and Caribbean lowlands. From May till November northeastern trade winds are warmer and weak, and only affect the Caribbean side of the country where they generate rainfall. However, temporary replacements of high pressure systems in the northern part of the Gulf of Mexico may cause stronger northeastern trade winds, thus generating more than average rainfall in the Caribbean side of the country and drier than average weather in the northern Pacific areas.

3: Instable low pressure belts which travel from the eastern part of the Caribbean Sea in western direction. They are caused by tropical cyclones which occur in the Caribbean Sea from June till October, and cause heavy rainfall for 1 to 3 days mainly in the Caribbean part of Costa Rica.

Besides these large-scale systems, locally also small-scale air movements as marine or mountain winds influence the weather.

2.2.2. Climate of the Atlantic Zone

The climate of the Caribbean lowland is hot and humid throughout the year. Although from December to April dry spells of up to several weeks may occur and may cause water shortage for pasture and crops, invasions of cold air from northern temperate and polar regions during this period produce moderate to heavy rainfall in the whole region during one or several days.

Most of the area receives a mean annual rainfall of about 3500 - 4000 mm, although yearly variation is large and may cause differences as much as 100% between one year and another. Mean annual rainfall diminishes along the coast from 5700 mm in the extreme northeast (station Barra del Colorado) to 3500 mm in the city of Limón. From the coastal area toward the west rainfall decreases to values of about 3600 mm (station Mola) at some 25 km from the Caribbean Sea. Rainfall increases with altitude at the transition of the lowland with the cordilleras, with values of 4450 mm at 250 m (station Los Diamantes in Guápiles), and about 7100 mm on the middle slopes of the cordilleras (station La Montura, 5 km southwest of the studied area at 1000 m) (Figure 3). In the parts of the Central cordillera above 2000 m rainfall is much lower, as indicated by the annual rainfall of about 2600 mm at station Irazú at 3400 m elevation.

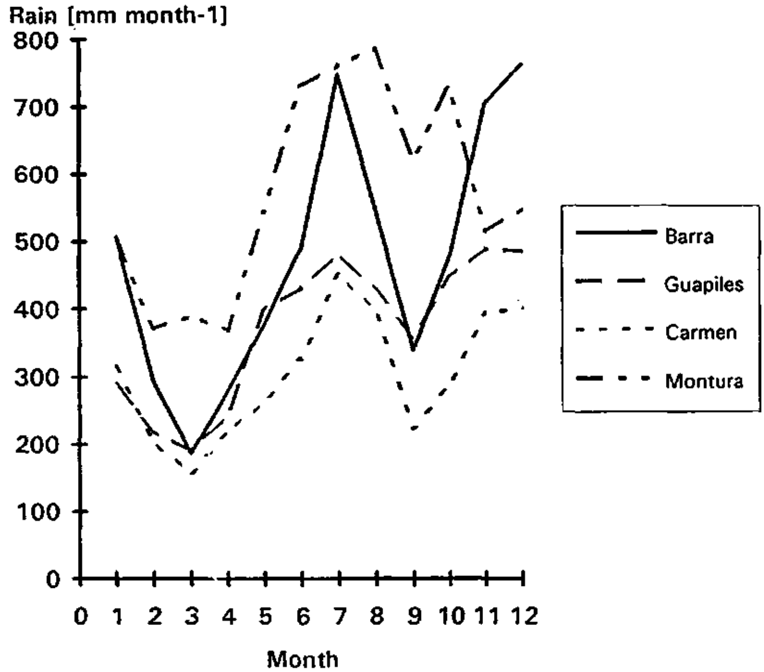

Figure 3: Rainfall distribution throughout the year for selected weather stations.

Vahrson (1991) calculated for a number of weather stations rainfall intensities with a return period of 10 years for events of 5, 10, 15, 30, and 60 minutes (Table 1). Compared to other parts of the country (data from San José and Liberia), rainfall intensities are rather low. As indicated by the slightly higher values for the Los Diamantes station, highest intensities probably occur on the lower and middle slopes of the cordilleras. Data for the weather station Irazú indicate that at altitudes >2000 m intensities are much lower than elsewhere.

Table 1: Calculated rainfall intensities with a return period of 10 years for various stations in the study area after Vahrson (1991).

station	altitude	rainfall intensity for an event lasting:				
		5 min	10 min	15 min	30 min	60 min
	(m)			(mm.h^{-1})		
Stations in study area:						
Mola	70	209	170	160	119	77
Los Diamantes	250	265	176	160	124	92
Carmen	15	210	145	135	106	83
Siquirres	70	226	164	144	117	84
Limón	5	241	163	130	108	79
Bataan	15	224	161	118	107	82
Stations outside study area:						
Irazú volcano[1]	3400	122	92	81	54	39
San José	1172	281	192	173	115	88
Liberia	85	402	272	223	166	103

[1] illustrative for altitudes >2000 m.

Mean annual temperature of the lowland is about 26°C, with differences between the warmest (April - June) and coolest months (December - January) of about 2 °C. Temperature decreases with altitude at a rate of 0.52 °C per 100 m (Herrera, 1985).

Relative air humidity is high, with average daily values of about 85 to 90% throughout the year. Even during dry spells, air humidity at noon never falls below 60%. Wind speed is low, and usually does not exceed 0.8 to 1.6 m s^{-1} in the interior (stations Los Diamantes, Carmen) and 1.9 to 2.8 m s^{-1} along the Caribbean coast (stations Barra del Colorado, Limón) (Zarate, 1978). Occasionally, during thunderstorms, local gusts of wind may damage crops, especially in banana plantations. No historic records exist which mention the occurrence of hurricanes on the Atlantic coast of Costa Rica.

Potential Penman evapotranspiration was calculated to be about 3.5 to 4.5 mm day^{-1} in the lowland (station Carmen) and decreases with altitude to about 2.5 to 3.5 mm day^{-1} on the footslopes of the cordilleras (station Los Diamantes at 250 m) (Castro, 1985). On the more cloudy and cooler higher slopes potential evapotranspiration is probably much lower.

Day length in the area varies from approximately 11.5 hours the 21th of December to 12.5 hours the 21th of June, related to its position between 10° and 11° north latitude.

2.2.3. Past climates

Palynological studies in Costa Rica and Panama indicate that climate changed during the Quaternary. During the last glacial maximum about 18,000 years ago, average temperatures at higher altitudes were an estimated 4 - 8 °C lower than at the present, and the forest line may have been located at an altitude of about 2000 m, down from its present 3300 m (Martin, 1964; Bush and Colinvaux, 1990; Hooghiemstra et al., 1992). It is likely, therefore, that also lowlands were cooler than today. Bartlett and Barghoorn (1973) think that during glacial periods rainfall, just as today, varied widely in amount and distribution at short distances, but that the Caribbean coast of Panama had a dryer climate or a more seasonal rainfall distribution. Therefore, it is possible that also the Costa Rican Caribbean lowland had a drier climate than today.

2.3. Land use

Widespread evidence of pre-Columbian settlements is found on well drained sites throughout the area (Willey, 1971). Recent colonization of practically virgin forest land started about 300 years ago along the Matina and San Juan rivers, and about 150 years ago in the footslopes of the cordilleras when a railroad was constructed. During the past 40 years conversion of forest into pasture and cropland accelerated to such extent that actually in the lowland outside the protected areas only swamp forest and patches of partially logged forest remain. Mountain slopes of both cordilleras still maintain a considerable forest cover. Pasture is the main land use (200,000 ha), followed by bananas with 45,000 ha (Belder, 1994). Range land is mainly used extensively for beef production, while large-scale banana plantations are managed intensively, with application of large amounts of fertilizer and pesticides. Although the total cropped area is considerable, crops other than banana often occupy only small fields within farms. The most important crops of the study area are briefly discussed below.

Until 1989 maize was one of the most important crops for many small farmers in the area because of its subsidized price. When subsidy was cancelled, the crop lost importance rapidly, and it now is grown mostly for home consumption or fresh cobs. In the study area land under maize cultivation decreased from about 18,000 ha in 1987 to about 1000 ha in 1994 (Abarca, 1995, pers.comm.).

Palm-heart is considered one of the promising crops for export, and suits the environmental conditions of the study area. At present an estimated 1500 to 2000 ha are used for palm-heart production.

Probably from the start of colonization, cassava and other root crops have been an important food crop for people of the area. With improving infrastructure, also the importance of root crops as a cash crop increased. Export of cassava alone increased from 3200 tons in 1978 to 12,000 tons in 1988 (SEPSA, 1989). This undoubtedly also increased the total area cultivated, for which, however, no data are available. Other root and tuber crops show similar tendencies.

Plantain has always been a food crop as well, of which commercialization increased as infrastructure improved, like cassava. The Costa Rican plantain production is concentrated in the Limón province, but the most important areas are south of the study area.

During the past 10 years several thousands of hectares of tree plantations have been established. Most plantations use exotic species such as teak, melina and eucalyptus trees, while native species are used less frequently. Furthermore, a considerable area of privately owned forest which has been logged is being managed for periodical timber production.

Actually, rice is only sawn locally for home-consumption. Until 1985 approximately, however, about 4000 to 7000 ha were grown in the Matina area. Apparently, high air humidity caused high pest control costs which made rice cultivation uncompetitive compared to the dryer Pacific lowlands.

2.4 DESCRIPTION OF SOILS

Using data stored in the GIS at the Atlantic Zone Programme (Wielemaker and Vogel, 1993), soils of the area have been subdivided into 8 groups, based on differences in age, parent material and genesis pathways (Figure 4):

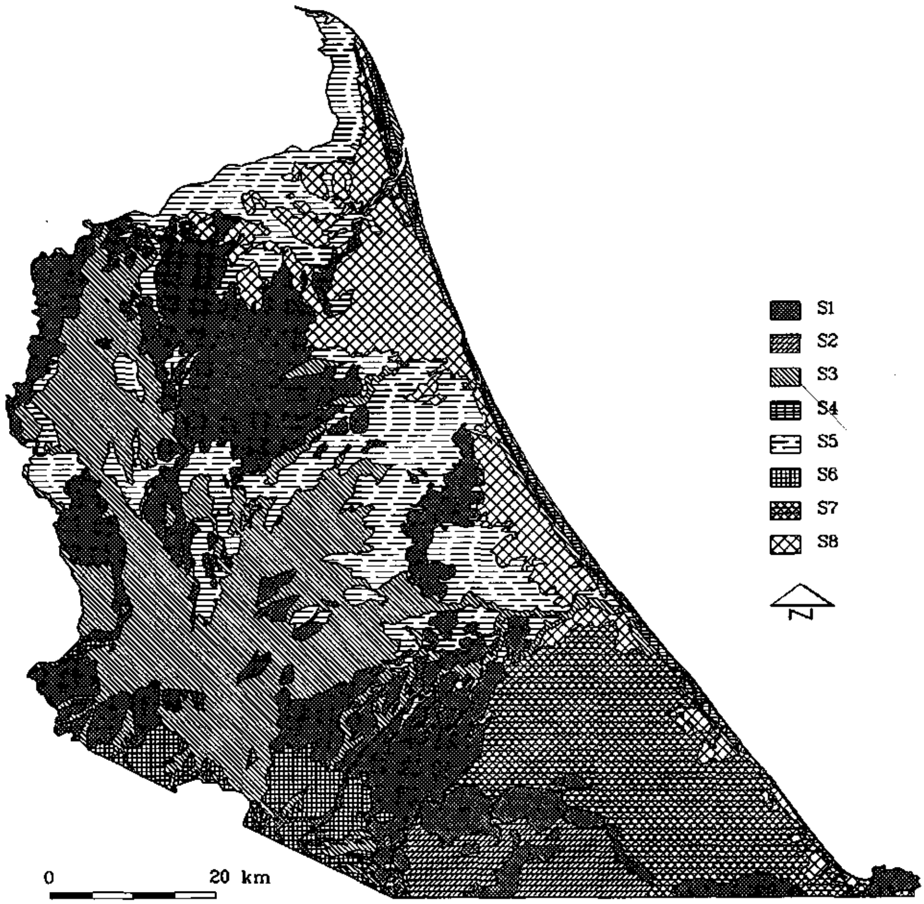

Figure 4: Distribution of mayor soil groups of the study area (for an explanation of the numbers S1 to S8, see table 2).

Table 2: Major soil groups of the Atlantic zone of Costa Rica as presented in figure 4.

S1:	Old, strongly weathered, well-drained clayey soils
S2:	Old, moderately to imperfectly drained clayey soils, developed on sedimentary rock in the Talamanca cordillera
S3:	Young soils developed on well-drained, sandy sediments derived from the Central cordillera
S4:	Young soils developed on poorly to imperfectly drained, sandy sediments derived from the Central cordillera
S5:	Young soils developed on fine textured sediments derived from the Central cordillera
S6:	Young soils developed on sediments from the Talamanca cordillera
S7:	Soils developed on ash deposits under extremely humid conditions
S8:	Holocene peat soils

In the following paragraphs these groups are briefly discussed. Soils were classified according Soil Survey Staff (1992).

2.4.1. Old, strongly weathered, well-drained clayey soils (Haploperox, Oxic Humitropepts).

Soils of this subgroup are found throughout the area on the dissected remnants of Pleistocene terraces, on the early Pleistocene basaltic hills in the northeast, on old fluvio-volcanic deposits, mud- and lava flows in the Central cordillera, and on Tertiary volcanic rock and conglomerates in the Talamanca cordillera.

Slopes vary between flat and steep, and stones may be present on basaltic hills and in both cordilleras. A thin (< 15 cm) A horizon is underlain by a thick homogeneous B horizon. Below a depth of 1 to 2 m the B horizon gradually changes into a CB horizon which is several meters thick, which was altered chemically and mineralogically, but has recognizable sedimentary or parent rock structures. In few cases unaltered parent material could be studied. Parent material in the dissected terrace remnants resembled that of young alluvial soils, while old lava flows in the Central cordillera rocks were similar to younger lava flows.

Texture is clayey, and organic matter content varies from about 5 to 10% in the A horizon to less than 1% in the B horizon. pH-H_2O varies from 4 to 5, and exchangeable acidity is usually more than 2 cmol(+) kg. The sum of exchangeable bases is low, and varies from 1-3 cmol(+) kg^{-1} in the A horizon to < 1 cmol(+) kg^{-1} in the B horizon.

Mineralogy.

Holdridge et al. (1971) describe three soil profiles on the low basaltic hills bordering the Río Colorado. Mineralogy is described as dominantly halloysite/kaolinite and gibbsite, with minor amounts or traces of montmorillonite. Verburg (1991, pers.comm.) studied thin sections of soils on these hills and observed that the $> 10\ \mu m$ fraction was composed mainly of gibbsite and iron concretions.

Analyses of soils on Pleistocene terrace remnants and on old lava flows from the Central cordillera indicated that the sand and coarse silt fractions are composed mainly of gibbsite and iron oxide concretions, and small primary opaque minerals. Weatherable minerals are almost absent in the fine-earth fraction of the upper meters of these soils. However, especially near the Central cordillera some fresh primary minerals are often present in the upper 50 cm due to recent addition of volcanic ash, while soils formed on lavas, mudflows, or conglomerates may still contain coarse primary mineral fragments throughout the profile.

X-ray diffraction analyses of Mg-saturated, field-moist clay separates showed 0.7 nm peaks in the upper part of most profiles and 1.0 nm peaks at greater depth. After air-drying these 1.0 nm peaks shifted to 0.7 nm, indicating the presence of 1.0 nm halloysite. Transmission electron microscope images of soils on Pleistocene terraces revealed the presence of tubular halloysite at all depths, although halloysite in the upper 40 cm or so had a more crumpled appearance. Besides kaolin minerals, gibbsite and goethite are present in variable amounts.

Total kaolin and gibbsite content for some profiles was estimated using TGA. Both minerals can make up maximally 40% of the total soil mass. Dithionite extractable Fe often amounts 5 to 8% in the upper meter of dissected terrace remains, indicating that crystalline iron (hydr)oxides can make up an important part of the soil mass.

Sometimes, in cemented parts of the CB horizon Al_o values were found to be extremely high (e.g. > 10%), indicating that the cementing agent is short-range order material. Regularly, low 1.4 nm peaks were observed in diffractograms. The peaks disappeared by heating to 300 °C, indicating interlayered vermiculite/montmorillonite. The presence of these minerals in the upper part of the profile may be explained by recent additions of volcanic ash, although they also were observed in the parent rock and may have been incorporated in the soil by disintegration of rocks. Finally, especially in the saprolite considerable amounts of cristobalite may be present.

In soils developed on more permeable rocks of the Talamanca cordillera, the dominant clay mineral is kaolinite, while gibbsite is only present in small amounts (Zunnenberg, 1990). Whether this is due to their older age and dominant kaolinite formation in past climates, or to factors as internal drainage and parent material composition, remains to be studied.

2.4.2. Moderately to imperfectly drained clayey soils, developed on sedimentary rock in the Talamanca cordillera (Aeric Tropaquepts, Aquic Humitropepts)

Sedimentary silt and claystone in the Talamanca cordillera are often calcareous and have a dense, impermeable structure. Most of these rocks have been altered hydrothermally or are strongly weathered, so the original mineralogy often cannot be recognized. Soils formed on these parent materials differ from those on other rocks in the area. Soils have a thin A horizon underlain by a B horizon which often shows mottling within 1 m depth. Texture is mostly clay loam to clay. Apparently, the dense rock structure impedes drainage to such a point that groundwater level may be within 1 m of the surface on the top of hills, and profiles show mottling at shallow depth. Soils are often very acid (pH-H_2O <4) and exchangeable acidity may be as high as 20 cmol(+) kg^{-1} soil. Organic matter content is high (up to 10%) in the A horizon and usually less than 1% in the B.

Mineralogy.

Holdridge et al. (1971) describe a moderately to imperfectly drained soil profile on limestone slightly south of the town of Siquirres in the Talamanca area. In spite of its high topographical position, the soil was imperfectly drained with mottles from 70 cm downward and had a pH-H_2O of 3.9 at this depth. I therefore assume that it belongs to this soil subgroup. Mineralogy is described as dominantly halloysite (50 to 60%), with some allophane and vermiculite (20 to 25% each). Since only XRD was carried out, the presence of allophane is at least questionable.

Very acid (pH-H_2O <4.2) soil profiles with a moderate to poor internal drainage developed on shales in the lower (<300 m) slopes of the Talamanca cordillera were studied by Zunnenberg (1990) and the author (Nieuwenhuyse, 1994). Although profiles were 20 km south of the study area, they belong to the same soil subgroup. Results indicate that the dominant clay mineral is smectite, with minor amounts of kaolinite. The X-ray diffraction patterns suggested that interstratification of smectite and kaolinite may occur, although this certainly requires further study. Contrary to the findings of Holdridge et al. (1971), no indications for the presence of short-range order materials or halloysite were found. Acid ammonium oxalate extractable Al values were <0.6%, and extractable Si was absent.

2.4.3. Young soils developed on sandy sediments derived from the Central cordillera (Udands, Tropopsamments, Andic Dystropepts).

Sand is probably the dominant component of the sediments brought down from the mountains and is deposited throughout the lowland. With increasing distance to the source areas, fluvial sand deposits become less dominant. Soils belonging to this group are found mainly on the alluvial fans and on mixed alluvial and mudflow deposits at altitudes of 50 to

300 m, as well as on choked river channels and adjacent overbank deposits throughout the alluvial plains. Along the Caribbean coast a sandy beach ridge plain occurs.

Drainage varies from moderately well to excessive. Soils have developed on >1 m sandy sediments. They usually have a thick (30 to 100 cm), brownish black loamy A horizon. In spite of the high activity of burrowing soil fauna, on alluvial fans sometimes different sedimentary deposits can be recognized in these thick A horizons, indicating that the soil parent material has been deposited during various events. The yellowish brown B horizon has a loamy to loamy sand texture and is 30 to 70 cm thick.

Many soils in the footslopes and on alluvial fans contain an important amount of gravel, cobble, and stones. Due to the depositional pattern typical of humid tropical alluvial fans (Kesel and Lowe, 1987), the coarse fragments are usually concentrated in elevated elongate ridges of 10 - 30 m wide, with 1 to 2 m lower parts between them, in which fine material may have been deposited during later floods. Accumulation of fine material is thought to be enhanced by surface runoff related erosion, which removes part of the <2 mm material from the ridges to the adjacent lower parts, causing the formation of an almost stone-free upper 40 to 80 cm of the profile. In such places the A horizon may grade directly into a C horizon. The absence of a B horizon seems to be due to poor drainage, since in such soils groundwater is present for a considerable part of the year at the transition A to C horizon. Apparently, biological activity homogenizes the soil thoroughly down to the ground water table. Alternation of both soil types is abrupt, and they may be found next to each other at distances <20 m.

Frequently, C horizon material is cemented to some degree, probably due to illuviation of amorphous Si-Al gels or iron compounds. Such cemented C horizons may hamper drainage and lower the production capacity of these soils.

In the central part of the alluvial plain, soils of this subgroup have developed on >1 m thick medium to coarse sand deposits, which were deposited as sheet sand or channel fills (see also chapter 4). In this subgroup also soils on levee and point-bar deposits have been included. In many of these soils a clear C horizon is often absent, and most are well to moderately well drained. Only the youngest soils are AC profiles, while in layered profiles the C horizon may be absent, and only buried A and B horizons occur.

Mineralogy.

Sand and coarser material is mainly andesitic. The sand grains consist of rock fragments composed of plagioclase, pyroxene and opaque mineral phenocrysts in a matrix in which volcanic glass usually is present (Van Seeters, 1993; chapter 6). Individual glass grains are scarce, and total glass content of the parent material is usually less than 20%, by volume. Individual plagioclase and pyroxene grains are common, and minor amount of opaques, hornblende, olivine, (felsic) rock fragments and sand-sized clay bodies are usually present. Such clay bodies were investigated in detail by Jongmans et al. (1994). They are composed of both 1:1 and 2:1 layer silicates and are thought to have formed by hydrothermal transformations in the volcanoes, and were subsequently eroded and transported.

In most well-drained soils of this subgroup, high Al_o, Fe_o, and Si_o values indicate that dominant weathering products are short-range order materials as allophane and Al- and Fe-humus complexes (Shoji et al., 1993).

X-ray diffractograms of clay separates of most of these soils (including soils of < 500 years old) showed only small peaks at 0.7 and 1.4 nm, indicating a dominance of amorphous materials. The kaolinite/halloysite and 2:1 layer silicates like smectite, vermiculite, and chlorite, probably came from disintegrated sand-sized clay bodies (Jongmans et al., 1994). In somewhat older soils, sometimes low 0.485 nm peaks indicate the presence of gibbsite. The soils on levees and point bar deposits have only weak Andic properties. This is thought to be due to the young age of the soils, which is also indicated by an often clear sedimentary layering.

2.4.4. Young soils developed on poorly to imperfectly drained, sandy sediments derived from the Central cordillera (Tropaquepts)

Soils of the this subgroup are found throughout the alluvial plain and occur more frequently in areas at elevations < 50 m above sea level. They have formed in imperfectly to poorly drained sand deposits, often with thin layers of fine-textured material. Many of the alternating sandy and loamy sediment layers are thought to be deposited as crevasse splays and floodplain deposits, while thicker poorly drained sand deposits are believed to be related to choked river systems (chapter 3), or they occur in the beach ridge plain. Soils of this subgroup are mostly found spot-wise in lower areas, alternating with soils of the previous group, which occupy geographically higher positions.

<u>Mineralogy.</u>

Poor drainage is thought to have slowed down weathering, preventing formation of clay. This is most clearly shown in the coastal plain. Higher parts (beach ridges) with well-developed Andisols on them (chapter 5 and 6) alternate with poorly drained lower areas (swales), with little evidence of soil formation except for slight alteration of sand grains and incorporation of organic matter. Clay mineralogy of poorly drained sandy soils is more variable than that of well-drained sandy soils. Often they contain clay minerals similar to those present in fine river sediments (smectite, vermiculite, and 0.7 nm kaolinite/halloysite), especially when sandy sedimentary layers alternate with fine textured layers. The clay in these fine textured layers is probably mixed through the soil by biological activity. However, 1.0 nm halloysite tends to dominate these sandy soils, while it is hardly found in fine-textured river sediments. In some poorly drained soils developed on thick sand deposits halloysite is the only clay mineral present. It was probably formed by *in-situ* weathering of volcanic sand in poorly drained environments, as described for New Zealand soils by Parfitt et al. (1983) and Parfitt and Wilson (1985).

2.4.5. Young soils developed on fine-textured sediments from the Central cordillera (Tropaquepts, Dystropepts, Eutropepts)

Moderately to poorly drained soils developed in fine-textured sediments dominate the lower parts of the alluvial plains. The poorest drained soils occur at elevations <20 m. Most soils of this subgroup are less developed and are probably young. Especially in the lower parts of the alluvial plains, sedimentation of fine-textured material is still active. Abandoning of river channels has made various of the backswamps inactive and has led to incision, and consequently to improvement of drainage.

Horizon differentiation in the somewhat better drained soils is moderate. Soils have a thin (5 to 15 cm thick) brown A horizon, underlain by a yellowish brown (often mottled) cambic B horizon which grades into a mottled CB horizon (or buried soil). Poorly drained soils have a weaker horizon differentiation and have grey AC profiles.

Texture varies from loamy to clayey, and profiles often show textural differences between or within soil horizons due to sedimentary layering. Structure development is related to drainage: under moderate to imperfect drainage structure is angular or subangular blocky, while the permanently waterlogged soils and deeper horizons have weak structures or are structureless. The common but short dry periods (Paragraph 2.1.2) apparently permit some degree of structure development: many of the poorly drained clay soils show a weak angular blocky structure down to a considerable depth (>50 cm). Ongoing sedimentation or incision of rivers may gradually improve drainage, permitting physical ripening and faunal activity, and consequently structure development.

Organic matter content is about 5 to 10% in the A horizon and decreases with depth to 1% or less at depths >50 cm. Soils are often slightly acid, pH values vary from about 5.8 to 6.5, although some of the most developed, moderately well drained clayey soils may have pH values of 5.0 to 5.5 in the topsoil. CEC values are usually >20 cmol(+) kg^{-1} soil, and exchangeable bases vary from 5 to 20 cmol(+) kg^{-1} soil.

Mineralogy.

A poorly drained soil profile described by Holdridge et al. (1971) along the Río Colorado contained mainly montmorillonite and kaolinite, with traces of gibbsite. A second profile at the footslope of low basaltic hills contained mainly halloysite and gibbsite, and some montmorillonite. Since none of the other investigated soils of this subgroup contained significant amounts of halloysite or gibbsite, its abundance in this profile is remarkable. However, this may be due to the proximity of the basaltic hills, in which both minerals are abundant and from which they may have been eroded and deposited at the site.

X-ray diffraction analysis of the clay fractions of soils formed on sediments derived from the Central cordillera indicate that the deposits are a mixture of 0.7 nm kaolinite or halloysite, 2:1 layer silicates (smectite, chlorite, and mica/illite or interstratifications), and some 1.0 nm halloysite, which must have been formed in different, contrasting weathering environments. Short-range order materials and gibbsite were not found in these soils.

2.4.6. Soil developed on sediments from the Talamanca cordillera (Eutropepts)

Sediments derived from the Talamanca cordillera differ in their mineralogical composition from sediments which originate in the Central cordillera. Composition of gravel in the bedload of the Matina river is mainly hydrothermally altered rock, sedimentary rock, slightly altered volcanic rock and few granites. Sand is composed mainly of strongly weathered rock fragments, with minor clay pseudomorphs, feldspar, and volcanic rock fragments. Many of the sedimentary rock fragments and hydrothermally altered fragments are calcareous, which is reflected in fresh silt and sand sized sediment. Analysis of fresh silt and sand sized river sediment indicate a $CaCO_3$ content between 0 and 2.6%, with an average of about 0.8% (Van Seeters, 1993). As a consequence, soils which develop on these sediments differ from those elsewhere in the Atlantic lowland.

Each year extensive areas are flooded and receive fresh sediments, so many soils in this southeastern area are young. This was shown in 1991 and 1992, when sediment supply increased tremendously after a strong earthquake in the Talamanca cordillera. During inundations following heavy rainstorms after the earthquake, up to 400 m wide areas along the Reventazón, Pacuare, Matina, and minor rivers were covered with up to 50 cm thick fresh sediment layers (Van Seeters, 1993). Soils generally have a thin (5 to 10 cm) A horizon overlying a cambic B horizon, which is often mottled in its lower part.

Organic matter content is about 4 to 10% in the upper 10 cm of the soil and less than 1% at depths >30 cm. Compared to soils developed on sediments from the Central cordillera these contents are low. pH and CEC values of these soils are high, as are contents of exchangeable Ca and Mg.

Mineralogy.

No XRD analysis was done for soils of this subgroup, but some indications about mineralogy were obtained indirectly. The mineralogy of the parent material differs from that of the other soils of the Limón basin by the absence of easy weatherable volcanic minerals. Consequently, soils of this area do not contain X-ray amorphous materials, which is reflected by low P retention ($<50\%$), and low contents of Al_o ($<0.5\%$). High CEC values at low organic matter contents indicate dominant presence of 2:1 layer silicates as smectite and vermiculite, which was also observed in samples from fresh river sediments (Van Seeters, 1993).

XRD patterns of clay suspensions from alluvial soils on sediments derived from the Talamanca cordillera south of the study area (Nieuwenhuyse, 1994) showed large smectite peaks and small kaolinite peaks. In recently deposited river sediment, usually some free $CaCO_3$

is present. $CaCO_3$ dissolves rapidly under the prevalent climatic conditions, and usually no reaction with HCl was observed in somewhat older soils (Van Seeters, 1993).

2.4.7. Soils developed on ash deposits under extremely humid conditions (Hydrudands)

Thick ash accumulations occur on the northern slopes of Turrialba and Irazú volcanoes from about 600 m upward. I suppose that this ash originally was rather fine-textured and contained more volcanic glass than the fluvio-volcanic deposits at lower elevations. Consequently, it weathered more easily. The different parent material, the extremely high rainfall and leaching, and the lower temperature has led to formation of soils which differ from others in the study area.

From 600 to 1500 m, field texture is usually silty clay loam or silt loam. Soils are deep AB profiles, and usually no clear ash layers can be distinguished. Above 1500 m, soils are sandier and a clear ash layering is present. Furthermore, on a 2000-years-old lava flow (Reagan, 1987), no clearly recognizable ash deposits were present at 500 to 1500 m above sea level. Apparently, in the past 2000 years ash deposition at these elevations was minimal. Therefore, it seems likely that soils above 1500 m are younger than the deep soils developed on ash at lower elevations, and that recent eruptions deposited significant amounts of ash only near the summit. Older, probably more voluminous, eruptions must have been responsible for the lower ash beds.

Organic mater content in the upper 10 to 30 cm of soils between 600 and 1500 m altitude is usually >15%, and may be as high as 30%. Organic matter is often still as high as 5 to 10% at depths of 50 to 100 cm. Topsoils are acid (pH-H_2O between 4.5 and 5.5), and pH increases with depth to values of about 6. CEC is very high (over 40 cmol(+) kg^{-1}), and base saturation low (<10%).

Mineralogy.

These soils have very high contents of Al_o (values over 6% are common), as well as P- retention values (>95%). Short-range order materials and metal humus complexes clearly make up most of the clay fraction. Allophane contents could not be calculated for a lack of data on acid ammonium oxalate extractable Si. The only available XRD analyses are by Colmet Daage et al. (1970) and ISRIC (pers.comm. Kauffman, 1994) for soils on the eastern slopes of the Turrialba volcano under slightly dryer conditions. They found some gibbsite, quartz and cristobalite (possibly caused by opal phytoliths), as well as traces of interstratified chlorite-montmorillonite and of kaolinite and/or dehydrated halloysite, confirming that short-range order material dominates the clay fraction.

2.4.8. Peat soils (Tropohemists)

Peat grows in permanently waterlogged sites where sediment supply is low or absent. Especially in the most distal part of watersheds behind beach ridges along the Caribbean Sea, and in small valleys enclosed in the dissected Pleistocene terrace remnants, peat deposits are found.

Most peats in the area are periodically flooded by rivers, and receive additions of some sediment (peat fens). The centre of some of the most extensive peat areas, however, are never flooded. Vegetation on such peat bogs is exclusively fed by rain water. Peat soils on fens and bogs differ significantly in their chemical properties, as illustrated by data in Table 3:

Table 3: Selected properties of centre (bog) and edge (fen) of peat deposit, Tortuguero National Park.

Depth	Loss on	pH		exchangeable cations			
	ignition	H_2O	KCl	Ca	Mg	K	Na
(cm)	(%)			(cmol⁺/ kg peat)			
<u>bog</u>							
0-20	95	4.3	3.9	1.3	4.8	1.5	1.2
50-60	97	4.5	4.0	1.9	1.8	1.0	0.8
<u>fen</u>							
0-20	83	5.2	5.0	9.8	12.8	0.5	0.9
50-60	81	5.2	5.2	10.3	14.3	0.5	0.7

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CHAPTER 3. VOLCANIC ORIGIN OF HOLOCENE BEACH RIDGES ALONG THE CARIBBEAN COAST OF COSTA RICA.

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Volcanic origin of Holocene beach ridges along the Caribbean coast of Costa Rica

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Abstract

The formation of beach ridges along the Caribbean Coast of Costa Rica, ranging in age between about 100 and 5000 years, appears to be related to discontinuous sand supply by volcanic eruptions in the Costa Rican Central Cordillera. This is indicated by consistent and statistical significant differences in chemical composition between sediments of individual beach ridges, which cannot be explained by textural differences. The differences are thought to be caused by variations in magma composition of different eruptions in the volcanic hinterland. SiO_2 contents of the ridges vary between about 52% and 59%, other elements show variations corresponding to magmatic differentiation. Petrographically this is expressed in variations in the amount of andesitic rock fragments and pyroxene grains. Chemical composition of beach ridge sediments is alike in composition to the erupted products. However, in spite of the removal of part of the mobile elements and mixing with weathered sediments, original differences between the erupted products appear to remain reflected in the resulting beach ridge sediments.

1. Introduction

Beach ridges may be defined as continuous linear mounds of mainly sandy sediment near the high-water line, heaped up by waves (Reineck and Singh, 1980). A beach ridge plain may be composed of a series of clearly developed ridges and swales, or may take the form of plateau-like barrier tops with small ridges and almost no swales.

Holocene sandy beach ridge plains are found in many places of the world, e.g. Nayarit coast, Mexico (Curray et al., 1969), and Kamchatka coast, Soviet Union (Zenkovich, 1967). Most of these plains were built on tectonically rather stable coasts when the rapid rise of the sea level, which started at the end of the last glacial age, began to slow down about 5000 ^{14}C years ago (e.g. Walcott, 1972), and beach ridges could extend seaward (Van Straaten, 1965).

Presence of several beach ridges behind each other would imply a certain cyclicity in ridge-forming coastal processes or a discontinuous supply of sediment, or a combination of both. Sand may be supplied by (1) landward movement of sediment stored on the continental shelf by wave action (e.g. Curray et al., 1969; Beets et al., 1992), or by (2) rivers discharging in the sea and subsequent longshore drift and coastal currents which transport the sand parallel to the coastline (e.g. Zenkovich, 1967). A minor source of sand (Komar, 1976), which locally may be important, is coastal erosion (e.g. Van Straaten, 1965).

A Holocene sandy beach ridge plain occupies almost the entire coast of the humid tropical Caribbean lowland of Costa Rica and continues

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northward into Nicaragua and southward into Panama. It is interrupted only in the south by a few outcrops of uplifted Tertiary rocks and Quaternary coral reefs (Battistini and Bergoeing, 1984). We studied geomorphology and geochemistry of the Costa Rican northern part of this plain for about 75 km between Punta Castilla and Jalova lagoon.

In many of the world's beach sediments mineralogical and geochemical differences between individual ridges are not or hardly discernable (e.g. Savage et al., 1988). The andesitic composition of the Costa Rican beach ridge plain, however, permits distinction of small differences between sediments of individual ridges, or sets of ridges. Within the framework of a soil development study, we found that differences in geochemical composition within one ridge caused by chemical weathering were less pronounced than differences between the succeeding ridges. We therefore decided to study these geochemical differences in more detail. In this paper we argue that discontinuous sand supply to rivers caused by volcanic eruptions in the hinterland strongly influenced formation and composition of beach ridges along the northern Caribbean coast of Costa Rica.

2. Setting

The studied beach ridge plain is located in the Limón basin, which forms the southeastern continuation of the Nicaragua Depression (Fig. 1). It is a backarc basin of tectonic origin, formed by the subduction of the Cocos plate beneath the Caribbean plate and dates back to the Early Tertiary (Weyl, 1980; Burke et al., 1984).

The Quaternary infill of the northeastern part of the Limón Basin is composed mainly of fluvio-volcanic sediments derived from the active Central American arc, attaining a thickness of up to 500 m in the coastal section (RECOPE, pers. commun., 1991). This indicates an maximal average subsidence of 0.25 mm/yr. During a strong earthquake in April 1991 (Richter 7.4), part of the basin south of the study area suffered a sudden subsidence up to 1 m (Malavassi, pers. commun., 1992). However, no subsidence could be observed in the study area.

The main rivers supplying sediment to the studied part of the shore are the San Juan (most of its water reaching sea through the Colorado river), Tortuguero and Parismina rivers (Fig. 2). The Tortuguero and Parismina river rise in the Costa Rican Central Cordillera, in which four active volcanoes are located. Although the San Juan basin drains geologically different areas, we assume that sediment contribution from the Tertiary and older rocks in its drainage basin is relatively small and mainly fine-textured due to their advanced stage of chemical weathering. Sandy sediments which are derived from the active volcanic arc in Nicaragua and northern Costa Rica are deposited in Lake Managua and Lake Nicaragua. Thus, we assume that most of the sand reaching the coast is derived from the Central Cordillera. In addition, also the active Arenal volcano drains directly into the San Juan river (Fig. 1). Sediment supply in these rivers is highly episodic. Strombolian, Vulcanian and scarce Plinian eruptions cause ash fall mainly on the higher parts of the volcanoes. During the 1963 to 1965 eruptions of the Irazú volcano the ash was found to be eroded almost as fast as to be deposited, and accumulation occurred only during the dry season. Consequently, rivers draining the summit area had extremely high sediment loads, not only during floods, but also during normal discharges (ICE, 1964; Waldron, 1967). We therefore suppose that much of the erupted products reached the sea rapidly, while a smaller part was deposited on alluvial fans and plains.

3. The coastal plain of northeastern Costa Rica

3.1. The shoreface

The offshore has a gentle gradient; the 20 m isobath is more than 5 km from the coastline. Tidal range is 0.3 to 0.6 m. No information on wave height, speed or frequency is available. During more than 60% of the time wind speed is less than 3 m s^{-1}, while wind speeds of over 7.5 m s^{-1} occur during less than 2% of the time

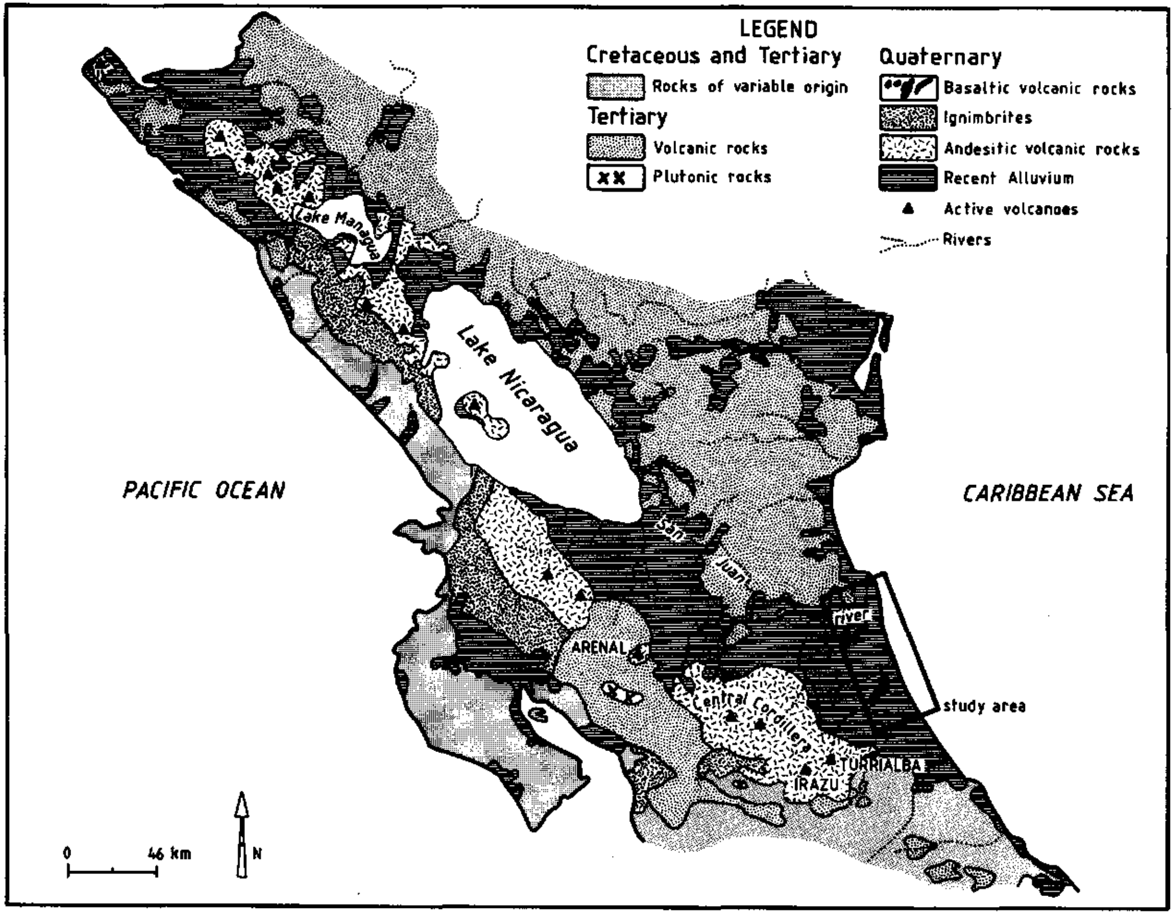

Fig. 1. Geological setting of the study area. Simplified redrawing from "Mapa Geológico Preliminar de Nicaragua" (IGN, 1973) and "Mapa Geológico de Costa Rica" (Ministerio de Industria, Energía y Minas, 1982).

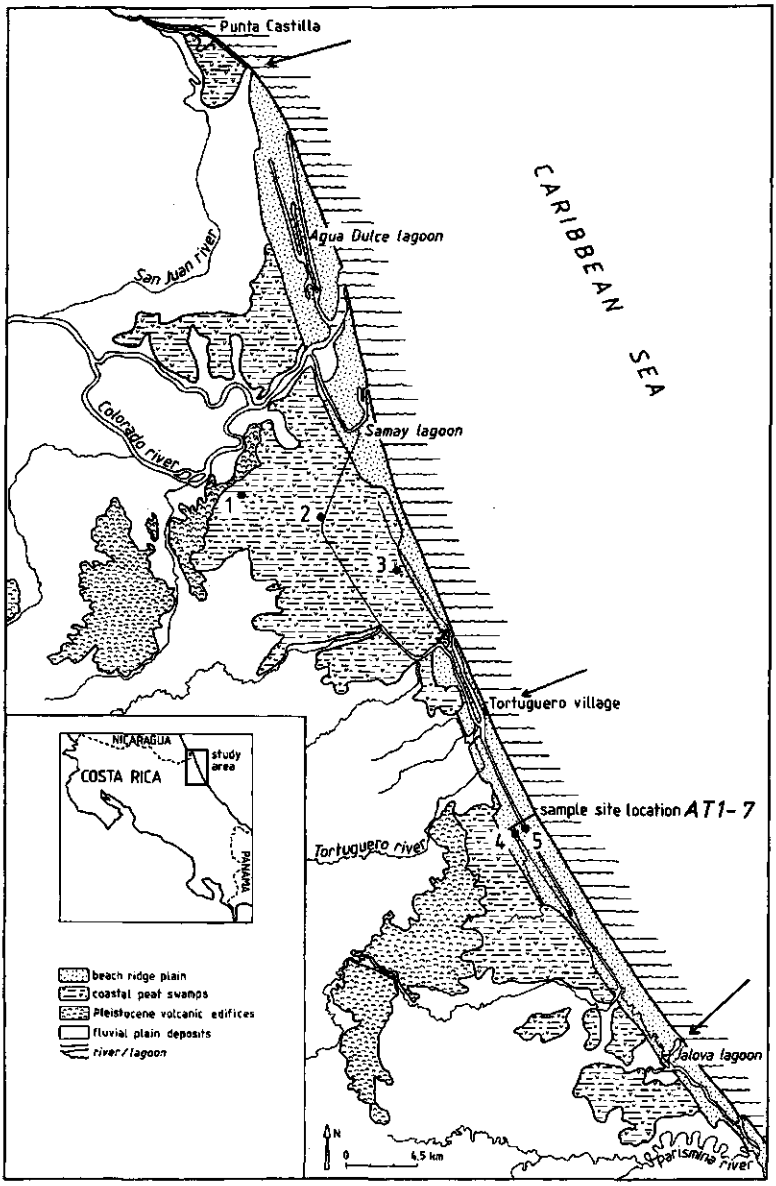

Fig. 2. Geomorphology of the study area, showing also ^{14}C sample site locations (*1–4*) as well as point *5* referred to in text. Arrows mark positions of major river outlets about 125 years ago.

(Zarate, 1978). Storms or hurricanes have not occurred on the Costa Rican Caribbean coast in historic times. Prevailing northern and northeastern winds (Zarate, 1978) produce longshore drift in mainly southern direction. According to local fishermen, coastal currents run from northwest to southeast, enhancing the effect of longshore drift. The beach is generally 20 to 40 m wide measured between the low water line and the beginning of vegetation.

3.2. Morphology and age of the beach ridges

The Holocene sea-level rise, together with the ongoing subsidence of the Limón basin, led to deposition of ever higher beach ridges at the sea side, and burying of inland ridges by forest peat or fluvial sediments (Figs. 2 and 3).

Width of the visible beach ridge plain varies between 0.3 and 3 km. Ridges are nearly straight and some extend for over 9 km, although their length is variable and difficult to measure due to the dense vegetation cover. Often, a clear beach ridge pattern cannot be recognized, neither in the field nor on aerial photographs (Fig. 4). We do not know whether this is due to post-deposition processes (erosion, soil formation) or if deposition of clearly recognizable beach ridges and swales alternates with deposition of a strandplain. Most recent ridges are at 3 m above sea level, while more inland ridges are lower (Fig. 3).

Thickness of the sandy barrier sediments in the transect is at least 10 m. In drill holes down to that depth using a hand operated bailer boring set no abrupt textural changes were found, although slight coarsening of sediments with depth was encountered.

In none of the drill holes shells or other datable material was found. Therefore, we estimated the age of the beach ridges by ^{14}C dating on peat samples taken with a Dachnovsky sampler in the coastal area. An age of 6285 ± 45 yrs B.P. (GrN 17321) was obtained at the base of a 9.8 m-thick peat deposit resting on sand which we interpreted to be of marine origin (point 1, Fig. 2), about 11 km inland at 1–2 m above sea level. A sample

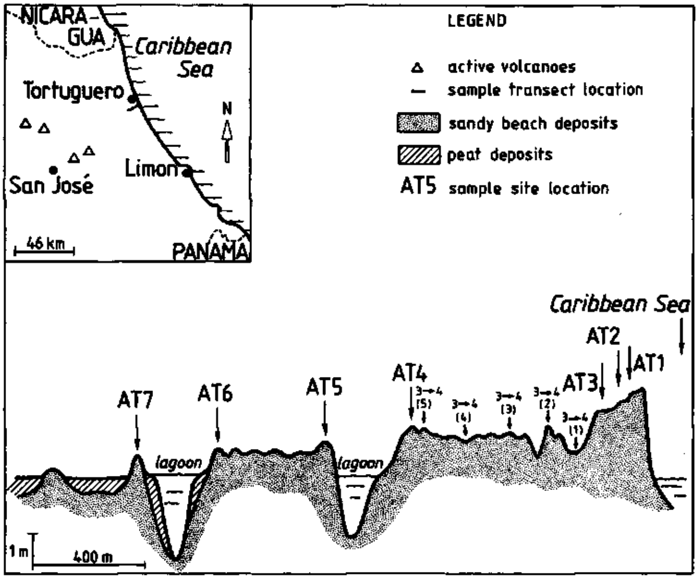

Fig. 3. Cross-section of studied transect and location of sample sites.

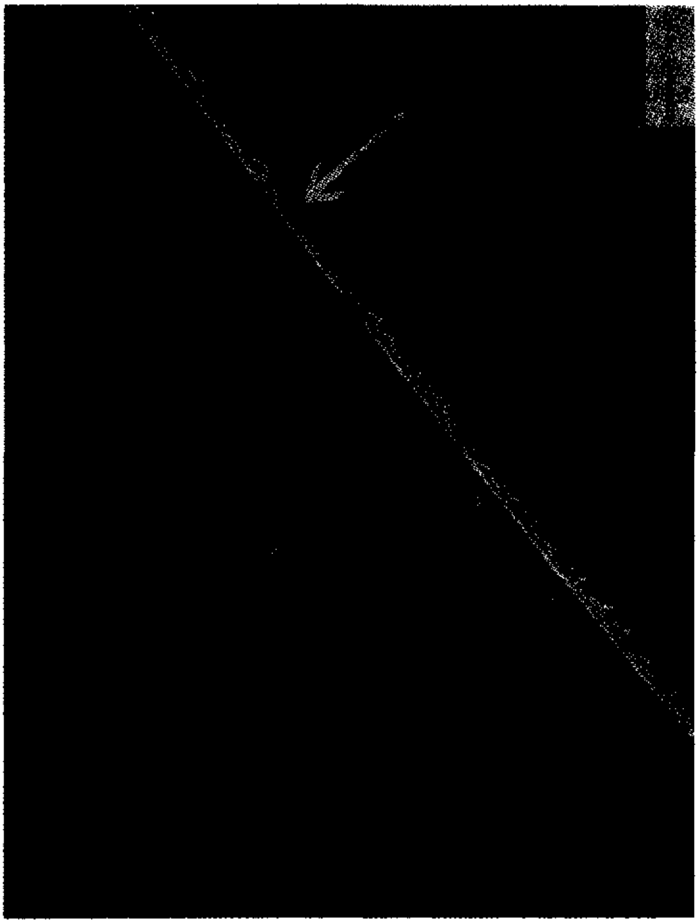

Fig. 4. Aerial photograph taken 17-03-1981 showing the studied transect (indicated by arrow). Scale 1:35,000.

taken at 1 m above this transition was dated at 5760±60 yrs B.P. (GrN 17322). Nowhere in the coastal plain thicker or older peat deposits were found. At smaller distances to the sea, sand was found closer to the surface: at 5 km from the sea the transition sand–peat was at about 1.5 m below sea level (point 2), giving an age of 3460±60 yrs B.P. (GrN 17566), while at 1150 m from the coastline sand at about sea level was covered with 65 cm of peat (point 3). Its base was dated at 1065±45 yrs B.P. (GrN 17567). Given the depths at which these samples were taken and the fact that because of the decrease in rate of sea-level rise around 6000 yr B.P. (Van der Plassche, 1982; Fairbanks, 1989) sediment deposition gradually could cope with sea-level rise, we estimate that a prograding barrier system was established from about 5000 yr B.P. onward. Ridges found at the surface are therefore probably less than 4000 years old. At the east side of the lagoon which is located in between the AT6 and AT7 sites (point 4) we dated the transition peat–beach ridge sand at about 1 m below sea level at 1830 ± 75 yrs B.P. (GrN 17318). At the west side of this lagoon, an organic mud layer at the transition with barrier sand gave a ^{14}C age of 1980 ± 40 yrs B.P. (GrN 17319). This means that the AT6 ridge must have been present at that time. We suppose that about 2000 yrs B.P. environmental conditions of this site changed, enabling peat growth to start. This change may have been caused by the formation of the lagoon separating AT4 and 5 (point 5), implying that these ridges may be slightly older, or about 2000 years old. However, it cannot be excluded that also AT6 has about this age. Based on soil development on similar, well dated fluvial sand deposits in the Limón basin (A. Nieuwenhuyse, unpubl. data), we estimate that the AT3 site is less than 500 years old and AT1 and 2 less than 200 years. The progressively older age of the ridges going inland is also evident from increasing soil development with depth (Nieuwenhuyse et al., 1993).

3.3. Recent coastal developments

Using aerial photographs, we studied coastal changes between 1952 and 1989, with an estimated precision of measurements of about 30 m. Around the outlet of the Colorado river the coast prograded 200–300 m, indicating that actually this river carries large volumes of sand into sea, which subsequently is deposited as beach ridges by wave action. Part of the sand may also be derived from coastal erosion in the area just south of Punta Castilla. Coastal recession in this section was estimated to be maximally 300 m during the last 35 years. During field work we observed that beach ridges with well developed soil on them were being eroded by waves, implying that during a considerable time span this part of the coast must have been prograding or stable. It also indicates that the volume of sand carried into the sea by the San Juan river actually is too small to compensate for erosion. We think that erosion in this section has been going on since the 19th century when, according to old reports (in: Gonzalez, 1910), a sudden change in the river course caused the major outlet of the San Juan river to shift from Punta Castilla to its present position. Dynamiting of the river course upstream in 1905 enhanced this effect (Céspedes, 1967). Between Samay lagoon and Jalova lagoon no changes of the coastline could be measured. In this sector an equilibrium seems to exist between coastal erosion and sedimentation. The absence of changes around the outlet of the Tortuguero river may be explained by a change in the size of its drainage basin. According to an older report, the Tortuguero basin included a large area of rainy mountain slopes on which much of the ash erupted by Turrialba and to a lesser extend Irazú volcano is deposited (Gonzalez, 1910). On aerial photos from 1952 and 1960 the Tortuguero river is still connected with the mountainous areas through an apparently intermittent branch. In 1970 a flood choked the connecting branch with sediment and since then the Tortuguero river rises as a small brook in the footslopes of the Central Cordillera. Based on measurements of the drainage basin and rainfall data we estimate that discharge of the Tortuguero river has reduced by about 60%. The most important change must have occurred between 1860 and 1952, and this is thought to be the reason for the absence of changes on the coastline observed on aerial photos.

These observations illustrate the importance of sediment input by rivers for the shaping of sandy coastal plains: depending on the volume of sediment input dominating processes at the coast may vary from coastal erosion to beach ridge formation. They also stress the importance of the location of river outlets: close to the Colorado outlet high sediment input leads to deposition of many short ridges, while at greater distances (= lower sediment input) deposition of new ridges is less frequent, but ridges are longer. The fact that ridges between Tortuguero and Jalova lagoon are rather long indicate that they were deposited at rather great distances from a river mouth, probably the Tortuguero river.

4. Methods

To study geochemical and grain size variability between the different ridges, we used the same 2.5 km transect perpendicular to the coastline south of Tortuguero village (Figs. 2 and 4) as sampled by Nieuwenhuyse et al. (1993, 1994) for soil development studies. Except for two abandoned river channels (lagoons) which actually drain the swamps and never inundate the beach ridge plain, fluvial influence is absent in this area. Seven tops of ridges, designated AT1–AT7 (Fig. 3) were sampled for further study. Many more ridge tops can be distinguished in the transect, but these were not investigated. Height differences were measured using a theodolite.

Along each selected ridge, parallel to the coastline, 10 samples were taken 5 m apart in order to obtain an overall characterization. Although the modern backshore sediments show layering, this feature was absent in any of the pits dug at the sampling sites. This is thought to be caused by nesting sea turtles, which each year arrive in large numbers, and thoroughly mix the sediment of the youngest ridge to a depth of about 90 cm. This was considered the maximum depth at which samples should be taken. Furthermore, post-depositional chemical changes should be minimal. Therefore, samples of the AT1 to AT3 ridges were taken at a depth of 40 to 50 cm, while samples of AT4 to AT7 were taken at a depth of 80 to 90 cm. The samples of AT6 and AT7 contained small amounts of soil material. Some individual samples were taken in between the AT3 and AT4 ridges.

Geochemistry of the last (1963–1965) eruption of Irazú volcano was studied in 6 bulk samples of different ash layers around the crater. From the literature, some data about chemical composition of the last Irazú eruption and all data on geochemistry of eruptions of Turrialba volcano were taken.

The bulk chemical composition was determined by X-ray fluorescence. After ignition at 900°C to determine contents of crystalline water and organic material, glass disks were obtained by melting aliquot of the samples with lithium tetraborate and analyzed on a Philips XRF assembly. Major elements were recalculated to a volatile-free basis. Sand texture was determined by sieving over 2000, 425, 250, 212, 150 and 53 μm.

Thin sections had been prepared previously at one site per ridge from various depths in combination with the study on soil formation. Since soil formation was found not to have a profound influence on mineralogical composition these data have been included in this study. Only data from the topsoils of the four older profiles were not included, since weathering was found to lead to comminution of grains (Nieuwenhuyse et al., 1994). Petrographical composition was determined by point counting of 300–600 grains in thin sections, using standard optical techniques. Rock fragments were classified as a separate category, and not the individual phenocrysts inside them. It was not tried to normalize for possible effects of grain size distribution on the mineralogic composition of the sands by means of fractionation.

5. Results

The beach ridge sediments consist mainly of sand with a grain size varying between 53 and 425 μm (Table 1); less than 30 g kg^{-1} were composed of >2 mm particles. M.50 value varied between 150 and 250 μm. Weathering may have influenced slightly particle size distribution in the two oldest ridges: at 80–90 cm depth, about 15% in AT6 and 18% in AT7 was composed of silt and clay-sized material. Petrographically the ridge sediments consisted of single mineral grains and rock fragments (Table 2). The single mineral grains were mainly plagioclase, augite and hypersthene. Also few green and brown hornblende grains and traces of biotite, olivine and opaques were found. The rock fragments were mainly andesitic, composed of plagioclase, pyroxene and small magnetite phenocrysts and a matrix in which volcanic glass may be present. Frequently pre-deposition weathering features were observed. Microcrystalline felsic rock fragments were few, composed of arkose and quartzarenite. Finally, common spherical sand sized bodies were found, composed of layer silicates which are thought to have formed by hydrothermal processes in the volcanoes (Jongmans et al., 1994).

Differences in chemical composition between ridges or sets of ridges found in the previously mentioned study on soil development were confirmed by the results of this study (Table 3). The chemical composition of the samples from recent Irazú ash is given in Table 4, together with litera-

Table 1
Mean textural composition of the sand fraction of the beach ridges sediments (particle size classes in micrometers)

Site	No. of samples	Weight (%)				
		2000–425	250–425	212–250	150–212	53–150
AT1	10	2.2^{a}	43.1^{a}	24.9^{a}	24.7^{a}	5.2^{a}
AT2	9	0.9^{ab}	32.7^{b}	27.5^{b}	30.1^{b}	8.8^{b}
AT3	10	0.5^{c}	22.4^{c}	25.1^{a}	37.1^{c}	15.0^{c}
AT4	9	0.6^{cb}	20.9^{cd}	27.8^{ab}	39.4^{cde}	11.4^{d}
AT5	7	0.4^{cd}	21.5^{c}	27.3^{ab}	38.5^{ce}	12.4^{d}
AT6[1]	10	0.5^{c}	16.4^{d}	24.3^{a}	43.0^{d}	15.8^{cd}
AT7[1]	10	0.2^{d}	10.3^{e}	17.3^{c}	39.9^{cd}	32.3^{e}

[1]Slightly influenced by soil formation.
Different letter superscripts in a column indicate significant differences by *t*-test for variability at 1% level.

Table 2
Mean petrographical composition of the >20 μm fraction of beach ridge sediments. Minor components are not listed

Site	No. of samples	Volume (%)			
		Andesitic rock frag.	Pyroxene min. grains	Plagioclase min. grains	Clay bodies
AT1	3	48^{abc}	24^{abcd}	21^{ab}	4^{a}
AT2	3	46^{ac}	18^{a}	27^{b}	5^{a}
AT3	4	48^{a}	18^{a}	26^{a}	4^{a}
AT4	5	58^{b}	6^{b}	22^{b}	9^{bc}
AT5	5	59^{b}	5^{b}	23^{b}	9^{b}
AT6	5	58^{b}	11^{c}	18^{b}	8^{bc}
AT7	5	42^{c}	26^{d}	21^{b}	5^{ac}

Different letter superscripts in a column indicate significant differences by *t*-test for variability at 5% level.

ture data on composition of material erupted by the Irazú and Turrialba volcanoes.

6. Compositional differences between beach ridges

If a number of succeeding beach ridges were derived from a similar source, in general one would not expect significant geochemical differences between them. In this study, we found that SiO_2, K_2O and Na_2O are consistently lower in AT1, 2, 3 and 7 than in AT4, 5 and 6, while MgO, CaO and TiO_2 values are lower in AT4, 5 and 6 (Table 3). AT7 shows considerably higher TiO2 values than all other ridges. A parallel trend can be seen in the mineralogical composition, with relatively low amounts of pyroxene grains and relatively high amounts of andesitic rock fragments in AT4, 5 and 6. It seems obvious that the differences shown by chemical and mineralogical data are related and that, for example, high MgO contents may be due to high pyroxene contents.

It is unlikely that during the last 5000 years the contribution of the different source areas has varied considerably and that, for instance, the amount of sand derived from the Tertiary volcanic rocks oscillated greatly during this period. Neither do textural differences explain the observed geochemical variation. Profiles may chemically be identical, while texturally they are different. For example, sediments of the AT1 and 3 ridges are chemically almost identical, while texturally they differ significantly (Tables 1 and 3).

The differences can neither be ascribed to pro-

Table 3
Chemical composition of beach ridge sediments

	Mass fraction (%)											
Site	AT1[1]	AT2[1]	AT3[1]	3→4(1)[2]	3→4(2)[2]	3→4(3)[2]	3→4(4)[2]	3→4(5)[2]	AT4[1]	AT5[1]	AT6[1]	AT7[1]
SiO_2	53.92^{a}	54.65^{a}	54.50^{a}	54.02	55.19	56.23	57.91	57.00	57.36^{b}	58.13^{c}	56.84^{db}	52.43^{e}
TiO_2	1.06^{acd}	1.02^{a}	1.04^{ac}	1.21	1.05	0.88	0.94	1.00	0.92^{b}	0.97^{c}	1.15^{d}	1.57^{e}
Al_2O_3	16.87^{a}	17.82^{a}	17.93^{a}	16.88	18.23	20.09	21.48	20.49	21.05^{b}	21.02^{b}	18.91^{c}	17.72^{ac}
Fe_2O_3[3]	9.59^{a}	8.82^{a}	8.93^{a}	9.66	8.70	6.87	7.23	7.15	7.06^{b}	7.61^{c}	9.27^{a}	12.18^{d}
MnO	0.18^{a}	0.16^{ad}	0.16^{ad}	0.15	0.13	0.10	0.07	0.11	0.10^{b}	0.13^{cd}	0.15^{d}	0.16^{ad}
MgO	6.55^{ad}	5.76^{a}	5.91^{a}	6.40	5.37	3.95	2.68	4.00	3.22^{b}	3.34^{b}	4.99^{c}	6.86^{d}
CaO	8.24^{a}	7.91^{b}	7.88^{b}	7.95	7.30	7.34	4.89	5.89	5.62^{c}	4.35^{d}	4.61^{d}	6.19^{e}
Na_2O	2.32^{a}	2.47^{a}	2.46^{a}	2.35	2.53	2.88	2.86	2.89	2.84^{b}	2.76^{b}	2.48^{a}	1.73^{c}
K_2O	1.06^{ad}	1.15^{a}	1.15^{a}	1.12	1.25	1.40	1.70	1.42	1.57^{b}	1.52^{b}	1.41^{c}	0.99^{d}
P_2O_5	0.19^{a}	0.19^{ad}	0.19^{a}	0.23	0.24	0.20	0.18	0.21	0.20^{b}	0.16^{c}	0.15^{c}	0.14^{d}

[1]Presented values are means of 10 samples.
[2]Individual samples taken in between AT3 and AT4 ridges.
[3]Sum of Fe_2O_3 and FeO, expressed as Fe_2O_3.
Different letter superscripts in a row indicate significant differences by *t*-test for variability at 1% level.

Table 4
Mean chemical composition of eruption products of Turrialba and Irazú volcanoes

	Mass fraction (%)				
	Turrialba				Irazú
Unit[1]	1	2	4	3+5[2]	1
Age	125[3]	<500[4]	<1415[5]	±1980[5]	27[3]
No. spl.	7	4	5	11	9[6]
SiO_2	52.34^{a}	56.23^{ade}	59.74^{bde}	58.10^{bd}	54.35^{ce}
TiO_2	1.09^{ad}	0.88^{ac}	1.16^{ae}	0.87^{bca}	1.18^{d}
Al_2O_3	17.24^{a}	17.41^{a}	17.49^{ac}	17.21^{a}	15.94^{bc}
Fe_2O_3[7]	8.51^{a}	7.36^{bc}	6.39^{acd}	6.71^{bd}	7.93^{ac}
MnO	0.14^{a}	0.12^{b}	0.11^{bc}	0.12^{b}	0.13^{ac}
MgO	6.18^{a}	4.99^{ac}	4.52^{ad}	4.02^{bcd}	6.48^{a}
CaO	9.26^{a}	7.86^{b}	6.75^{bd}	6.95^{cd}	8.15^{b}
Na_2O	3.27^{a}	3.41^{ac}	3.36^{ad}	3.73^{bcd}	3.28^{a}
K_2O	1.43^{a}	1.75^{bac}	2.43^{bc}	2.27^{b}	1.89^{c}
P_2O_5	0.41^{a}	0.30^{b}	0.48^{ba}	0.32^{b}	0.36^{ba}

[1]Stratigraphic units (Reagan, 1987).
[2]Data from 4 samples of lava flow (unit 3) and 7 samples of pyroclastic unit (unit 5), probably of same age.
[3]Historic eruptions.
[4]Age estimation based on soil development (Reagan, 1987).
[5]Eruptions dated with ^{14}C (M.K. Reagan, pers. commun., 1991).
[6]Means for Ti, Mn and P are based on 8 samples.
[7]Sum of Fe_2O_3 and FeO, expressed as Fe_2O_3.
Different letter superscripts in a row indicate significant differences by *t*-test for variability at 1% level.

gressive weathering, as systematic trends with age are lacking and samples were taken from (almost) unweathered sediments. Moreover, K and Na are likely to show the same behavior as Ca and Mg with respect to weathering, whereas their trends are opposite in the beach ridge sediments.

During magmatic differentiation SiO_2, K_2O and Na_2O are enriched with respect to MgO and Fe_2O_3, especially. Therefore, a more likely explanation for the differences in geochemistry among the studied beach ridge sediments is that the composition of the primary volcanic sediment transported by the rivers to the sea has changed in time.

It is useful to compare the sediment composition with data on chemistry of the volcanoes supplying sediment to the coast (Table 4). Reagan (1987) distinguished one lava and four pyroclastic eruptions of the Turrialba volcano in the last 2000 ^{14}C years, which together with the Irazú volcano must have been the main sediment supplier to this part of the coast. It is evident from these data that compositional changes occurred between these eruptions. By far the largest of these eruptions has been dated about 2000 yrs B.P. The eruptive history of Irazú volcano is less well known, although since 1723 at least 6 small to moderate eruptions have been recorded (Barquero, 1977).

The most recent eruptions of both Turrialba (1864–1866) and Irazú (1963–1965) are characterized by low Si contents. Such low Si contents are also found in the AT1, 2 and 3 beach ridges, strongly suggesting that the sediments of these ridges mainly originate from these eruptions and were deposited less than 125 years ago (Fig. 5). The AT4 and 5 ridges are estimated to be about 2000 years old, and its sediment composition resembles composition of rocks of this age erupted by Turrialba volcano (Fig. 5).

Eruptive history of Turrialba before 2000 yr B.P. and chemical compositions of its products are poorly known. Since we assume that the AT6 and 7 ridges are older dan 2000 years, we did not try to correlate these ridges with eruptions. Neither can eruptions of the Turrialba volcano about 1400 and <500 yrs B.P. be related to any of the sampled ridges, due to lack of sampling of ridge sediments between the AT3 and 4 sampling sites. However, their chemical composition (Table 4) and that of

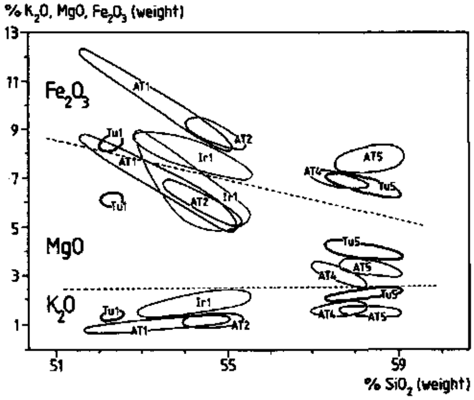

Fig. 5. Variation diagrams of *AT1*, *AT2*, *AT4* and *AT5* sediments, youngest Irazú eruption (*Ir1*) and Turrialba eruptions 125 (*Tu1*) and 2000 (*Tu5*) yrs B.P.

5 samples taken between the AT3 and 4 ridges (Table 3), do not indicate discrepancies with the above mentioned hypothesis.

Concentrations of the components SiO_2 and TiO_2 in the beach ridge sediments fall in about the same range as those of rocks of Turrialba and Irazú volcanoes, and the differences in these element contents between the beach ridges are in the same order of magnitude as the differences between rocks of different eruptions. Therefore, it seems likely that changes in primary sediment composition supplied to the coast have occurred indeed.

Cluster analysis of all major element data of beach ridges and eruptions indicates a relationship between chemical composition of beach ridges and eruptions. When 4 clusters are distinguished, one cluster is composed of the AT1, 2 and 3 ridge material together with the most recently erupted products of Turrialba and Irazú. AT4, 5 and 6, and the Turrialba eruption 2000 yrs B.P. form a separate cluster 2 (Fig. 6).

The beach ridge sediments are alike, but not identical in composition to the volcanic rocks, however. Some beach ridges are enriched in Al_2O_3 and Fe_2O_3, all are depleted in Na_2O and K_2O and to a lesser extent in CaO and MgO with respect to the volcanic rocks (Tables 3 and 4).

In addition to source composition, transport in

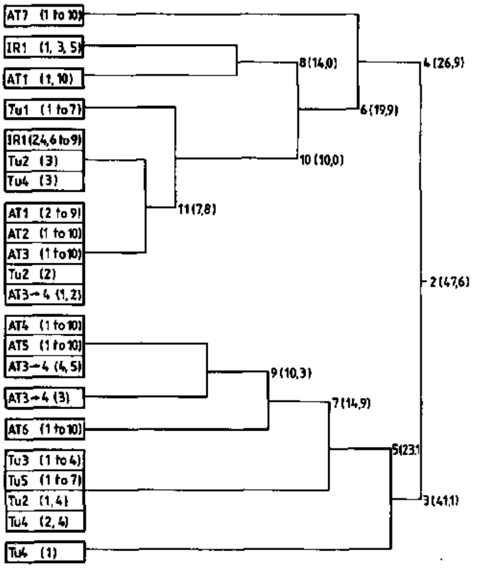

Fig. 6. Results of cluster analysis: dendrogram using average linking between groups for chemical composition of 110 samples of beach ridge sediments (*AT1–AT7*) and volcanic rocks (*Tu1–Tu5* and *Ir1*). Numbers between brackets refer to cases. Number of clusters are followed (between brackets) by squared Euclidean distance.

rivers and along the coast may have influenced sediment composition. During the various transportation pathways, sediment compositional modifications may have been produced by chemical weathering, e.g. in storage phases on mountain slopes and in the alluvial plains (Johnsson and Meade, 1990), as well as by the different ways of transport (Savage et al., 1988; Johnsson, 1990).

Chemical weathering of the beach ridge sediments by pedogenesis can be evaluated from Table 5: although chemical weathering during 2000 years (AT5) or more (AT7) has removed part of especially Ca, Na, K, and Al and Fe appear to be enriched, main differences in composition between beach ridges remain.

The greater amount of plagioclase and pyroxene grains in the AT1, 2, and 3 as well as the AT7 ridge as compared to AT4, 5 and 6 may be due to selective transport, but may also indicate different volume of the eruptions: the Plinian eruption of Turrialba about 2000 yr B.P. must have caused much more sediment transport toward the coast than the Strombolian eruptions during the last 500 years. The contribution of fresh grains must have been greater. Since rock fragments fall apart into their individual constituent during chemical and physical weathering, the amount of single minerals grains is thought to be higher in older, than in younger rock material. The low number of individual minerals in the ridges related to the 2000 yr B.P. event indicate that their residence time in rivers and along the coast was rather short and that mixing with older sediments was small.

We think that the presented mechanism of rapid erosion of volcanic ash and subsequent transport by normal fluvial action and deposition as beach ridge did not alter significantly the chemical composition of the sediments, even though the humid tropical climate and petrographic composition favor rapid chemical alteration of sediments.

7. Conclusions

(1) All surficial beach ridges studied were deposited within the last 4000 to 5000 years.

(2) The contribution to beach ridge formation of materials transported by rivers seems much more important than the contribution of materials already stored on the continental shelf. Input into the coastal area of large volumes of volcanic sand apparently compensated for, or even reversed the coastline regression one would expect when sea level rises about 1 m/1000 years.

(3) Coastal regression or progradation on the Caribbean coast of Costa Rica appears to have a direct relationship with supply of sand: relatively little sandy material leads to coastal erosion, while supply of relatively much material gives rise to progradation. Thus, beach ridge formation seems episodic and periods during which ridge formation prevails alternate with periods characterized by a stable or eroding coastline.

(4) The observed differences in chemical and mineralogical composition of sandy material arriving at the coast are thought to result from differences in eruption type and composition of the

Table 5
Chemical composition of selected soil profiles developed on beach ridges

Sample depth (cm)	Mass fraction (%)								
	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MgO	CaO	Na_2O	K_2O	P_2O_5
Profile AT2									
0–2	54.81	0.94	16.91	9.54	6.35	8.18	1.79	1.04	0.20
4–8	55.24	0.93	17.13	9.17	5.92	8.30	1.80	1.09	0.21
25–30	54.88	0.98	17.68	9.02	5.91	8.10	1.85	1.15	0.19
55–60	55.11	0.96	17.55	8.94	5.94	8.09	1.83	1.15	0.19
Profile AT5									
0–3	59.93	0.95	21.04	7.95	2.98	3.52	1.84	1.36	0.21
5–8	59.59	0.97	21.43	8.06	2.93	3.39	1.84	1.36	0.21
12–16	59.67	0.97	21.43	8.07	2.86	3.31	1.87	1.41	0.19
22–26	58.72	0.97	21.96	8.13	3.04	3.38	1.88	1.49	0.19
40–45	57.96	0.94	21.39	7.96	3.47	4.33	2.05	1.51	0.15
70–80	58.33	0.92	20.29	7.75	3.68	5.01	2.14	1.52	0.16
135–145	59.24	0.94	20.39	7.59	3.01	4.66	2.20	1.60	0.22
190–200	60.07	0.94	20.59	7.42	2.60	4.45	1.98	1.62	0.23
Profile AT7									
2–8	55.57	1.66	15.75	13.24	6.13	5.57	1.02	0.73	0.20
14–20	53.48	1.40	20.20	11.92	5.51	5.10	1.12	0.87	0.20
28–33	52.92	1.50	18.98	12.00	6.46	5.66	1.20	0.94	0.15
38–48	53.58	1.37	18.75	11.67	6.35	5.74	1.25	0.95	0.13
60–70	53.30	1.42	19.88	11.73	5.65	5.38	1.35	0.97	0.14
90–100	53.92	1.39	17.97	11.32	6.41	6.18	1.45	1.08	0.10
150–160	55.03	1.30	16.98	10.00	6.52	6.93	1.63	1.15	0.27
200–210	52.07	1.62	12.88	13.35	8.97	8.50	1.29	0.82	0.27

erupted volcanics, possibly modified by diverse manners of transport and weathering. Probably these differences have been more or less preserved during the subsequent formation of ridges, and the time period between volcanic production and accumulation in beach ridges may not exceed 100 years.

(5) Juxtaposition at the earth surface of repeated eruptive volcanism and the marine realm may lead to the development of landscape features, which are usually considered to form without any intervention of volcanic processes.

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CHAPTER 4: VOLCANISM-INDUCED EPISODIC CHOKING OF RIVER SYSTEMS IN HUMID TROPICAL COSTA RICA

A. Nieuwenhuyse, M. Dechesne, F. van Ruitenbeek, and S.B. Kroonenberg

ABSTRACT

Late Holocene sand deposits fill up former river channels and cover adjacent overbanks in the central fluvial plains of the humid tropical Caribbean lowland of Costa Rica. Two facies types are distinguished. The first is an up to several km long, 3 to 10 m deep, and 10 to 80 m wide coarse sandy channel fill facies, which shows an overall fining upward sequence both in grain size and structure. This facies has a pebbly base, small coarse sandy large troughs in the middle part and sandy trough cross-beds in the upper part. The second facies is a wing-shaped, < 1.5 m thick, sandy overbank facies which shows small cross-beds and ripple-marks and extends laterally up to 300 m from the channel. The deposits testify to episodic supply of huge amounts of loose material on the upper slopes of the volcanoes that border the area, as a consequence of eruptions. Complete choking of the channel and concomitant shifts of the river course appears to occur only during extreme rainfall events. Of a total surface area of about 300 km^2, two smaller areas covered with this kind of channel fill deposits related to eruptions of Turrialba volcano about 2000 yrs BP are described, as well as a similar deposit related to the 1963-1965 eruptions of Irazú volcano. In the distal part of watersheds, more than 1 m thick crevasse splay sediments appear to be a less voluminous manifestation of fluvial sedimentation triggered by volcanic activity. One of such deposits is described, possibly related to the 1864-1866eruptions of Turrialba volcano. In the fluvial plain short periods of highly active sedimentation and landscape formation alternate with longer, rather inactive periods.

INTRODUCTION

Basins bordering uplifting volcanic mountain ranges are among the most dynamic fluvial environments in the world (e.g. Räsänen, 1991). Highly episodic sediment supply by volcanic activity or earthquake triggered landslides, and the continuous subsidence of the basin provide a setting for active sedimentation and frequent changes in river systems. Several papers on the influence of volcanism on sedimentation processes in fluvial basins have been published on both ancient (Mathisen and Vondra, 1983; Smith, 1987; Van der Wiel et al., 1992a and b), and modern deposits (Davies et al., 1978; Vessell and Davies, 1981). Most of these studies describe high-energy deposits within proximal areas, such as hyperconcentrated flood-flows and sheet flood events.

This paper intends to increase knowledge about these phenomena by focusing on sedimentation resulting from volcanism-induced sediment loads choking river channels in more distal areas of alluvial plains, and discusses their effects on landscape evolution in the humid tropical environment of the Limón basin, Costa Rica. Climate of this area is characterized by abundant rainfall (mean annual rainfall between 3.5 and 5.5m) and temperatures of about 26°C throughout the year.

GEOLOGICAL SETTING

The Limón basin forms the southeastern continuation of the Nicaragua depression. It is a back-arc basin related to the subduction of the Cocos and Nazca plates beneath the Caribbean plate and dates back to the Early Tertiary. In some places, the combined thickness of the Tertiary and Quaternary material in the Limón Basin is about 8 to 10 km (Weyl, 1980; Seyfried et al., 1991), and Quaternary deposits reach a thickness of 500 m in the eastern (deepest) part (pers. com. RECOPE, 1991), indicating a maximal average subsidence of 0.25 mm yr^{-1}. Seismic activity is low in the basin itself, in comparison with the surrounding mountainous and offshore areas (Montero et al., 1992).

Four active strato-volcanoes of Quaternary age (Weyl, 1980) are located in the Central cordillera bordering the studied part of the Limón basin. The eruption history of these volcanoes is only partly known, but frequent eruptions occurred throughout the Holocene (Barquero, 1977; Melson et al., 1985; Prosser and Carr, 1986; Reagan, 1987).

Recent infill of the central Limón basin is mainly composed of fluvio-volcanic sediments derived from the Central cordillera. Exposed lava, pyroclastic, and mudflows are confined to the footslopes (Fig.1) and have not descended below 100 m above sea level and 20 to 30 km beyond the crater. Some early Pleistocene volcanic cones up to 300 m, and dissected remains of Pleistocene fluvio-laharic terraces at 10 to 20 m above the actual sedimentation level of rivers occur. Their position indicates that most probably not all parts of the Limón basin are subsiding at the abovementioned rate, but that parts may even be stable or uprising. Holocene fluvial deposits cover an area of about 2900 km^2, of which about 80% (aerial distribution) is derived from the Central cordillera. Sediments deposited in the southeastern part of the Limón basin are derived from the Talamanca cordillera, which is composed of Tertiary volcanic, plutonic and sedimentary rocks (Weyl, 1980).

HOLOCENE FLUVIAL DEPOSITS

Holocene fluvial deposits are bordered in the southwest by the Central and Talamanca cordilleras, and in the east by a 1 to 10 km wide, Holocene coastal plain along the Caribbean sea (Fig.1). Sand derived from the Central cordillera has an andesitic composition and is highly susceptible to chemical weathering (see chapters 6,7). Geomorphologically, the fluvial deposits by rivers draining the Central cordillera can be divided into two main units:

1: Alluvial fans. Upon leaving the Central cordillera, the main rivers form alluvial fans, the largest of which is the Suçio-Toro Amarillo fan (300 km^2). Most likely as a consequence of uplift of the central Cordillera, the upper parts of most fans appear inactive and are being dissected. Radiocarbon dating of wood fragments in an incised part of the Toro Amarillo fan indicate that this fan system has been active for more than 10 500 years (Kesel and Lowe, 1987). Sediments consist mainly of gravel, boulders, and sand, with minor sieve deposits of silt and mud drapes. Gradients vary from 27 m km^{-1} in the inactive parts to 5 m km^{-1} in the distal parts of the fans.

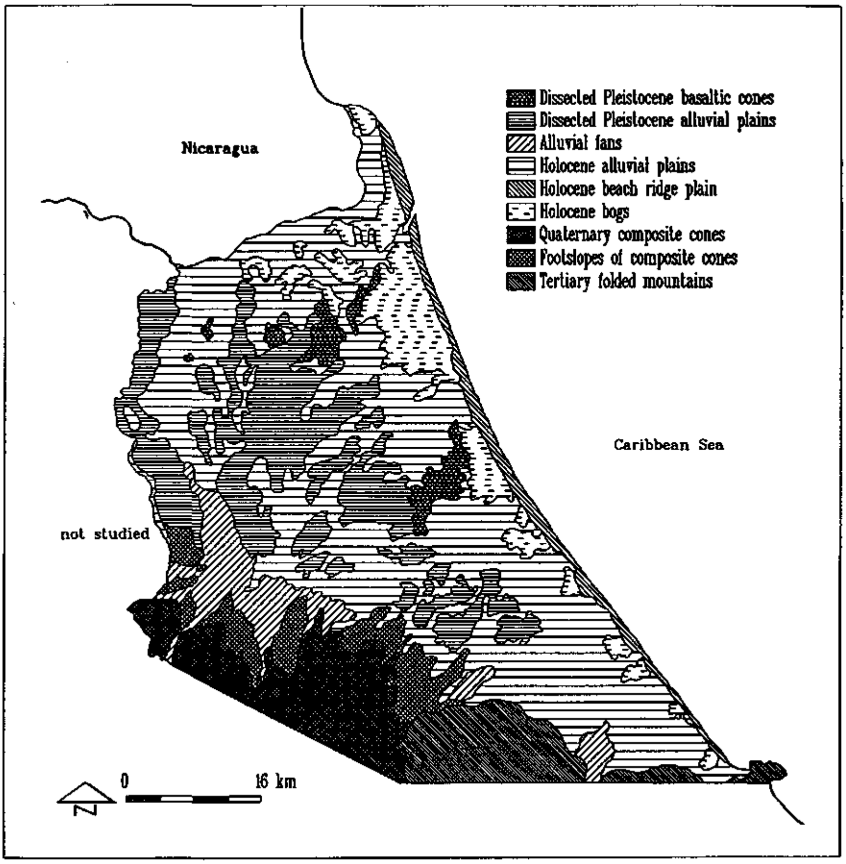


Figure 1: Main geomorphologic units of the Limón basin.

2: Alluvial plains. Downstream the fans, the major rivers follow a low-sinuosity meandering course, with scarce point-bar deposits and oxbow-lakes. Probably due to the low gradients caused by subduction, (Smith, 1983; Bakker, 1990), channel pattern changes to anastomosing in the northeastern part of the basin. Sediment load of rivers in the alluvial plains is mainly coarse (pebbly) sand, changing to medium sand and loam at the transition with the coastal plain. Levee and point bar sediments are mainly sandy, while fine textured deposits dominate in the backswamps. Thin sandy crevasse splay sediments are often found within the overbank deposits. Gradients vary from 5 m km^{-1} at the transition with the alluvial fans, to 0.5 m km^{-1} in the most distal parts. However, throughout the depositional pattern of the alluvial plains, abandoned river channels filled with sand, and extensive sheet like, sandy overbank deposits are common. Such sandy deposits are unusual under the prevalent sedimentation environment, and other sedimentary conditions than the normal meandering flow patterns must be considered to explain their occurrence. Three of these deposits are described below.

SHEET SAND AND CHANNEL FILL DEPOSITS

1: The 2 000 years old sand deposits in the central alluvial plain

In a geomorphological and soil survey of the Limón Basin (Wielemaker and Vogel, 1993), about 300 km^2 of a single sandy soil type, related to choked channels were mapped (Fig.2). Two areas where these sediments occur were studied in detail in newly dug drainage canals in banana plantations. These exposures enabled us to study the sedimentological characteristics and lateral extension of the deposits, but limited the investigated depth to 3 to 6 m. Locally deeper holes were drilled, and where possible, deposits were dated by ^{14}C analysis of buried organic remains.

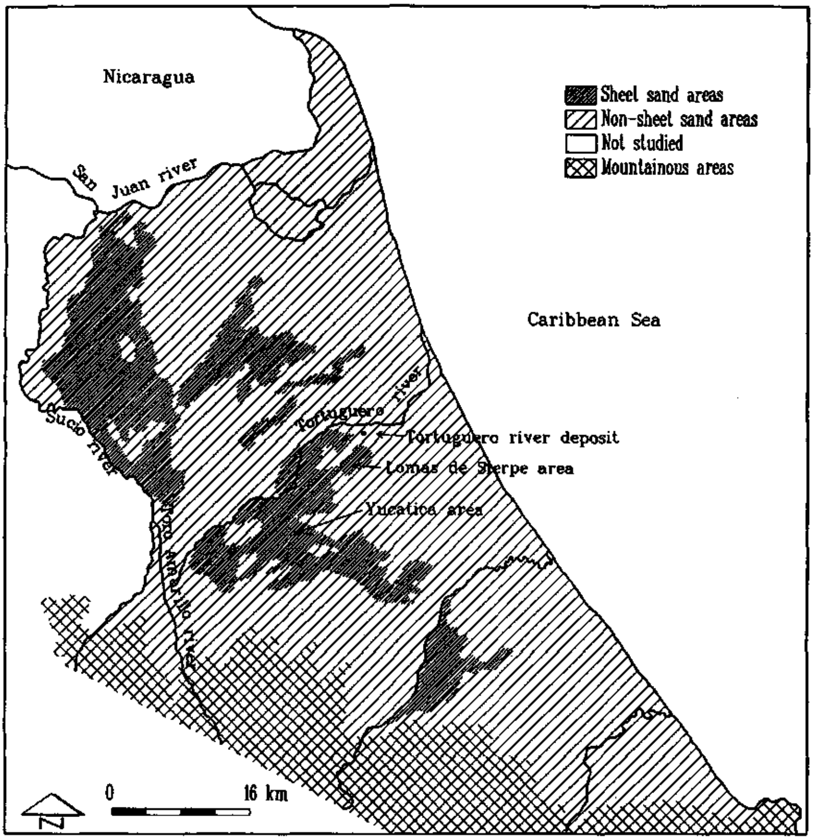


Figure 2: Areas in the Limón basin where sheet sand and channel fill deposits with similar soil development occur, as well as location of the studied deposits.

The Yucatica area

In the central part of the Limón basin, about 70 ha of the Yucatica area (Fig.2) was studied in detail, and adjacent areas were explored. In the exposures three sheet sand deposits have been distinguished, separated by finer grained units (Fig.3):

A lower sheet sand unit, containing a channel fill and an overbank facies. The channel fill is about 70 m wide and at least 2 m thick. Toward the south, at least 0.3 to 1 m thick overbank deposits extend for > 350 m. North of the channel fill overbank deposits are absent or not exposed (Fig.3). The lower part of the channel fill shows mainly trough- and planar cross-beds, while in the upper part a thin soil is present. No sedimentary structures can be recognized in the overbank facies, which we ascribe to homogenization by soil formation.

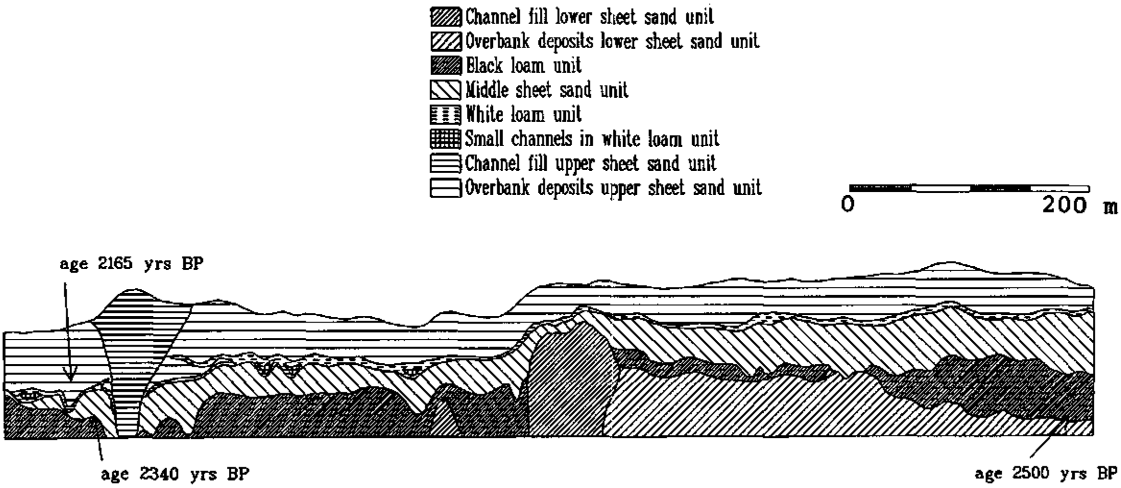


Figure 3: N/S-oriented cross section over the Yucatica area (stream direction of channels is towards NNE).

A black loam unit, which has reduction colours characteristic of a marsh vegetation, overlies the overbank deposits of the lower sheet sand unit, but not the channel fill. Dating of a seed at the base of the black loam unit indicate that its maximum age is 2500 ± 80 ^{14}C yrs (GrN19455) (Fig.3). Maximal thickness is 90 cm and sediments are mostly silt and clay sized, although also some sand lenses up to 40 cm thick and 40 m wide occur. The black loam unit shows mainly very fine horizontal lamination, while the small sand lenses within it show small cross-beds.

The black loam unit has an upper erosive boundary associated with several uprooted trees, overlain by the middle sheet sand unit, so sedimentation of the middle sheet sand unit appears also to have been accompanied by strong erosive currents. Only in the extreme northern part of the Yucatica area, the 40 to 140 cm thick middle sheet sand unit conformably overlies the black loam unit and sometimes contains organic remains. In the southern part of the area a sedimentological discontinuity divides the unit in an upper and lower part, suggesting that deposition took place in two phases. Branches at the transition of the black loam unit and the middle sheet sand unit (Fig.3) are 2340 ± 60 ^{14}C years old (GrN19457), so the middle sheet sand unit must be younger. Frequently load casts can be observed at the transition of overbank deposits of this unit and unconsolidated parts of the black loam unit. The unit is dominated by horizontal lamination and trough cross-beds in the lower part. In its upper part main sedimentary structures are trough cross-beds, with minor planar cross-beds and horizontal stratification, and part is unstratified. The apparent absence of sedimentary structures in the latter part of this deposit may be attributed to deposition as a hyperconcentrated flow, since no signs of soil formation are present.

Throughout the area the middle sheet sand unit is overlain by a 10 to 30 cm thick, horizontally laminated white loam unit. The absence of reduction colours in this unit indicates that at time of deposition the floodplains had been filled up to such extent that no marshy conditions prevailed. We interpret < 10 m wide and < 10 cm thick sand lenses within this unit to be crevasse splay sediments.

Finally, an upper sheet sand conformably covers the whole study area, and is composed of channel and overbank facies. Wood at the base of this unit was dated at 2165 ± 30 ^{14}C years BP (GrN19456), indicating that the unit must be younger. Sediments of the upper sheet sand are derived from a SW-NE oriented feeder channel which decreases in width from 80 m in the southwest to 20 m in the northeast. Overbank deposits can be traced for > 200 m, and decrease in thickness with increasing distance from the channel. The feeder channel of the upper sheet sand is dominated by planar cross-beds and horizontal lamination. Channels are filled with gravelly coarse sand while grain size of overbank deposits varies from pebbly coarse sand at the base of sedimentary structures, to fine sand. An up to 1 m thick soil has formed in the upper sheet sand unit. Burrowing by soil fauna and chemical weathering are the reason that no sedimentary structures are found in the upper part of this unit.

Outside the Yucatica area, several observation points revealed deposits with identical morphological, textural and sedimentological characteristics. Two of such similar deposits were dated at $< 2290 \pm 60$ ^{14}C yrs BP (GrN18197) and $< 2200 \pm 80$ ^{14}C yrs BP (GrN18941), so they seem to be of about the same age as the upper sheet sand unit.

Channel fills can be recognized in the actual landscape as ridges of 20 to 70 m wide and slightly higher than the surrounding surface. In the Yucatica area no younger sediments are found, except for some reworked material in natural drainage gullies.

The Lomas de Sierpe area

20 km downstream the Yucatica area, in the about 1000 ha Lomas de Sierpe area (Fig.2), sheet sand deposits similar to those in the Yucatica area occur (Figure 4). Overbank deposits laterally extend <100 m from the former channel, but are sometimes absent. A channel fill crosses the study area from SW to NE for at least 4 km (Fig.4, 5).

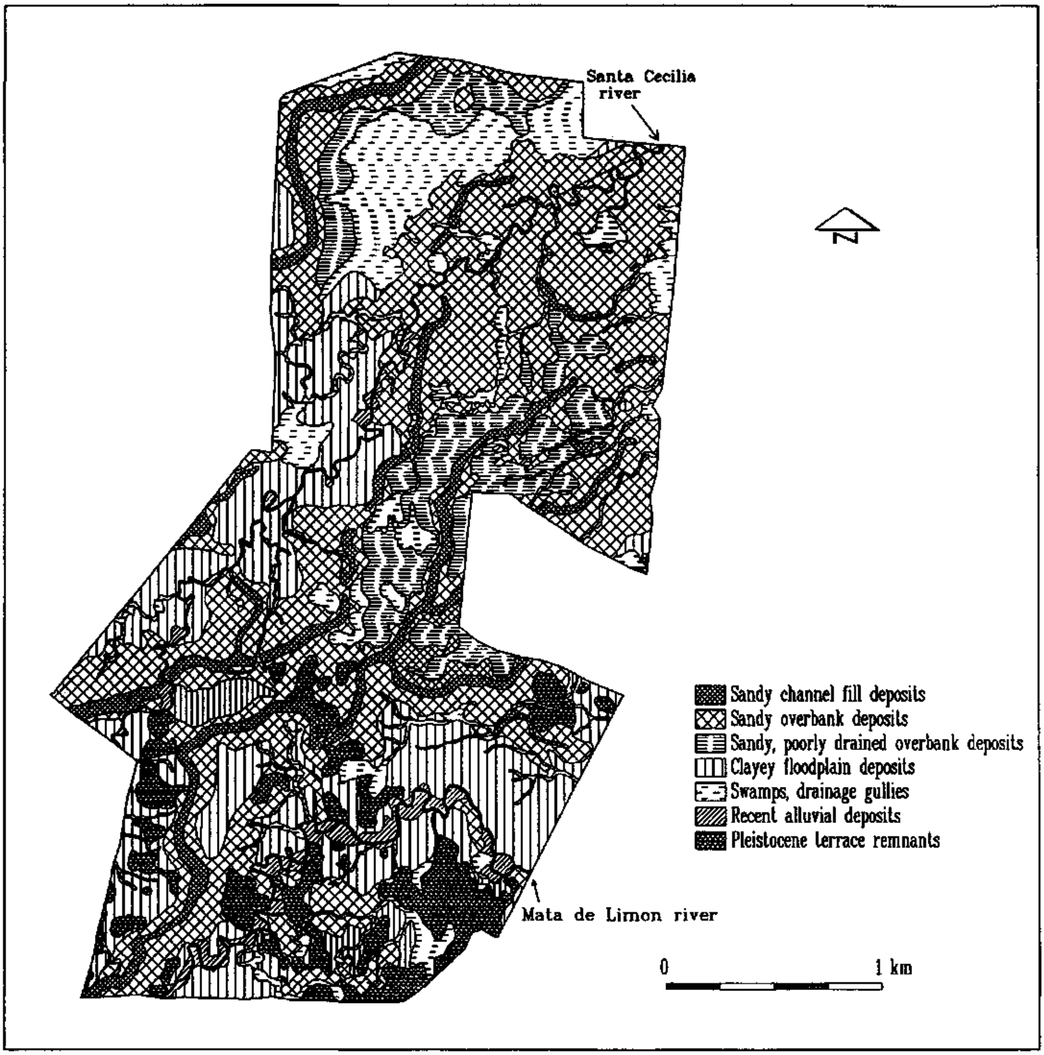


Figure 4: Distribution of main deposits in the Lomas de Sierpe area.

In the extreme northeast channel fill deposits appear to be absent. Leaves buried in sediments directly underlying this channel fill were dated at $2480 \pm 60\ ^{14}C$ yrs BP (GrN19967), indicating that it must be younger and may have about the same age as the upper sheet sand sediments in the Yucatica area. At least 3 other channel fills and their overbank deposits occur in the area. Soils formed in the channel fill or overbank deposits have a similar stage of development, indicating that they must be of about the same age (Nieuwenhuyse et al., 1993).

At the bottom of some of the drainage canals sheet sand deposits buried by the sheet sand deposits belonging to the 2000 yr BP event are exposed, without visible sedimentary structures. The degree of paleosoil development in these deposits suggest that they are at least 1000 to 2000 yrs older than the overlying sheet sand deposits, so similar events must have occurred earlier.

Channels have been filled up asymmetrically with medium to coarse sand deposits. The former outer curves of the river bed have been filled with coarser material, indicating higher stream velocities. Channel depth decreases in downstream direction from about 10 m to 3 m. Small pebbles (< 10 cm) are sometimes concentrated at the base, while finer (< 3 cm) gravel is often present at the base of other sedimentary structures throughout the profile. Channel fills show an overall fining upward sequence. At kilometre scale, a downstream decrease in grain size in the middle of the channel profiles is observed, but within distances of several hundred meters grain size varies in a non-systematical way.

Sedimentary structures in the deposits show little variation. Where exposed, channel fills have a massive lag deposit overlain by 0.5 m thick beds showing high angle cross beds, and by large trough cross-beds (wavelength 120 cm). They grade into trough beds which decrease in dimension (wavelength about 25 cm) in the upper 2.5 m (Fig.5). The upper metre is homogenized by soil formation. Erosive contacts and small clay wedges at the base of the channel fill and between channel fill and overbank deposits indicate that the existing channel may have been deepened and widened during the flood.

Overbank deposits are < 1.5 m thick, and have been extensively influenced by soil formation which makes it difficult to recognize sedimentary structures. Sometimes small (wavelength up to 25 cm), medium to fine sandy trough cross-beds can be observed at the base of the deposits. Grain size gradually decreases from medium sand close to the channel to loamy sand at larger distances. In various exposures a wavy contact zone can be observed between sheet sand and underlying deposits, which we ascribe to deposition in a forest environment without significant erosion or destruction of vegetation. Load-casts are often present at the base of the deposit.

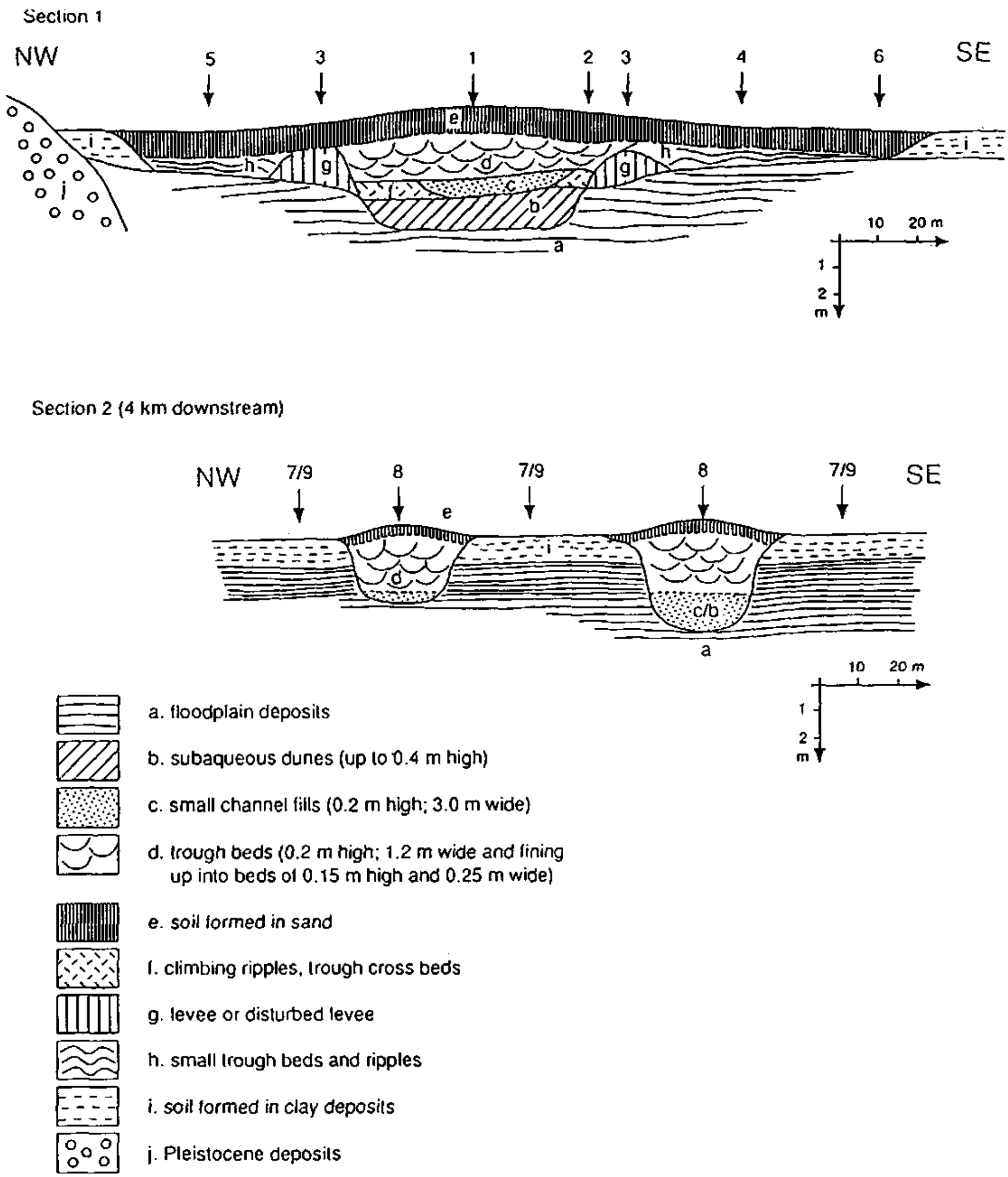
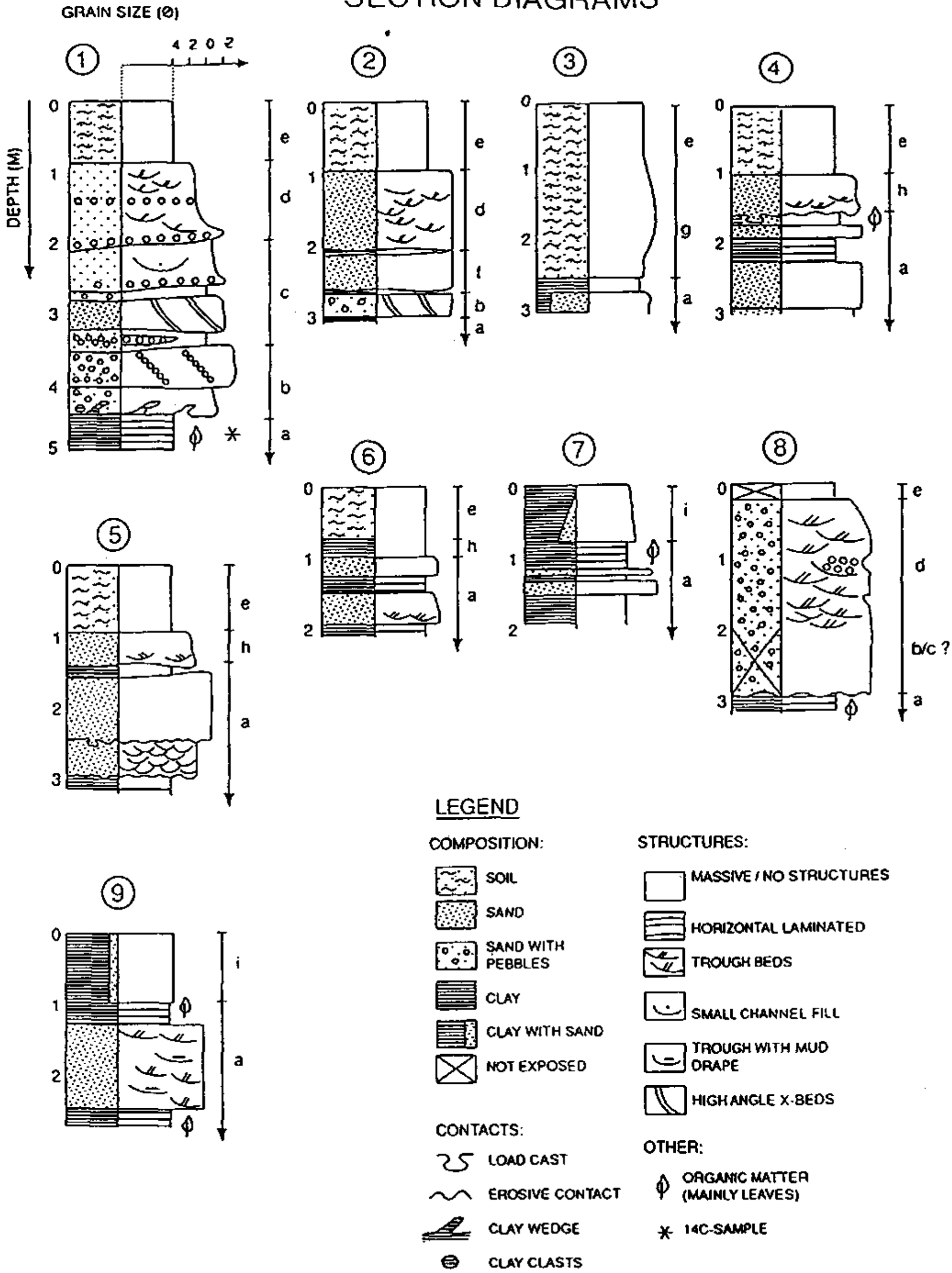


Figure 5: Cross section of channel fill in Lomas de Sierpe area (stream direction of the channel is towards NE).

SECTION DIAGRAMS



Channel fills can be recognized in the actual landscape as ridges of 20 to 70 m wide and up to 1 m higher than the surrounding surface. Toward the northeast sometimes part of the ridges and their overbank deposits are covered by younger sediments, and reappear at the surface further downstream. This indicates whether (1) an incomplete infill of former channels in downstream direction, or (2) different compaction rates at short distances. Occurrence of several more or less parallel orientated channel fills close to each other are thought to have impeded formation of superficial drainage gullies and cause the formation of poorly drained areas between them.

<u>2: The less than 230-year old Tortuguero river deposit</u>

The second case study is a sand deposit covering about 25 ha at 25 km downstream the Sucio-Toro Amarillo alluvial fan, about 50 km from the summit of the volcanoes, in the distal part of the Tortuguero watershed (Fig.2). Leaves at a depth of 1.1 m at the transition with an underlying unconsolidated clayey deposit were dated at 230 ± 30 ^{14}C yrs BP (GrN19966), so the sand deposit must be younger. In the western part of the area a W/E oriented, about 15 m wide, and more than 2.2 m thick sandy ridge is found which may have been the clogged feeder channel of the deposit. On both sides of the ridge there are lower (up to 1m), poorly drained areas covered with medium to fine sand mixed with organic matter, which are thought to belong to the same depositional event. Toward the east, the main ridge branches into several <1.5 m thick ridges, separated by swampy depressions of 5 to 15 m wide and 50 to 100 m long. No sand is found in these depressions; instead unconsolidated greyish loam and clay dominate to a depth of >2 m. Toward the north and east, the deposit is bordered by the present-day Tortuguero backswamps, characterized by reed swamps and patches of forest through which various unconfined creeks flow. No sedimentary structures were observed in pits dug in the deposit although in the upper meter two fining upward sequences occur, ranging in grain size from (very) coarse to medium sand. Since hardly signs of soil formation are present, we think it unlikely that any sedimentary structures present could have been erased by soil formation within 230 yrs. This may indicate deposition in two surges during the same flood, or two successive floods within a short time period.

Pronounced height differences occur between sand ridges and surrounding areas dominated by unconsolidated fine textured deposits. These differences cannot be due to settling, since one would expect that unconsolidated sediments permit rapid settling under the weight of the overlying sand, so sand deposits would be situated lower, rather than higher than the surrounding fine textured deposits. We speculate that differences exist since deposition. Height differences may have been caused by the presence of reed vegetation which prevented deposition of sand, and small creeks which were filled up. We conclude that deposition probably took place as a crevasse-splay in a backswamp environment, filling up mainly existing creeks up to 1-2 km from the Tortuguero river. Spill-over from the feeder channel deposited sand also in swamp areas adjacent to the channel in the proximal part of the deposit.

3: The 1970 choking of the Chirripó channel

About 8 km and further downstream the Sucio-Toro Amarillo alluvial fan, at 45 km from the Irazú crater, sand deposits occur where 1967 topographic maps indicate the Chirripó river (Fig.6). This river ran for about 40 km through the northern fluvial plains before joining the San Juan river. On 1960 aerial photographs the main stream splits into many smaller, apparently anastomosing channels, which seem to end in vast reed swamps. Today, however, the Chirripó river joins the Sucio river and follows a meandering course until it joins the San Juan river, 25 km upstream from its old mouth.

According to local people the old channel was abandoned during a flood in 1970. An area of about 10 to 20 km^2 along the former river bed has been covered with sand deposits. At about 15 km from the beginning of the deposit and further downstream sediments become less continuous and sometimes only the partly filled channel can be recognized, while overbank deposits may be absent. Further downstream overbank sediments may reappear, and the former channel has been filled up completely again. Also along some minor rivers which in the past probably formed part of the Chirripó system, identical sandy overbank deposits are found, indicating that their channels were (temporarily?) connected with the Chirripó system before and during the 1970 flood.

Sedimentary structures in the channel fill are poorly exposed and their study was mostly limited to the upper 2 meters. At the blocking point a deeper exposure shows a massive coarse sand and gravel bed from 6 to 4.5 m depth, overlain by horizontal laminated coarse sand. In the upper part of all channel exposures only coarse and medium sandy horizontally laminated beds up to 50 cm and small channels with cross stratification occur. Sometimes also massive beds are found.

The up to 1.5 m thick overbank deposits overlie older sediments, in which often a buried soil with vegetation remains can be recognized. The observed fine and medium sandy current ripples and small cross-beds may indicate lower stream velocities than in the channel.

23 years after the 1970 Chirripó event, erosion has levelled much of the originally superficial structures which we suppose were present after deposition, and the whole area is vegetated again. In the upstream part of the deposit, no height differences between channel fill and floodplain deposits can be discerned. Later floods are thought to have deposited the thin layers of fine textured sediments observed in lower parts, although some may also have been deposited as mud drapes during waning flood stage. The downstream area has dried up considerably since 1970, and many sandy deposits almost without signs of soil formation occur. As in the second case study, sand deposits are about 1 m higher than the surrounding areas dominated by unconsolidated fine textured and fine sandy deposits.

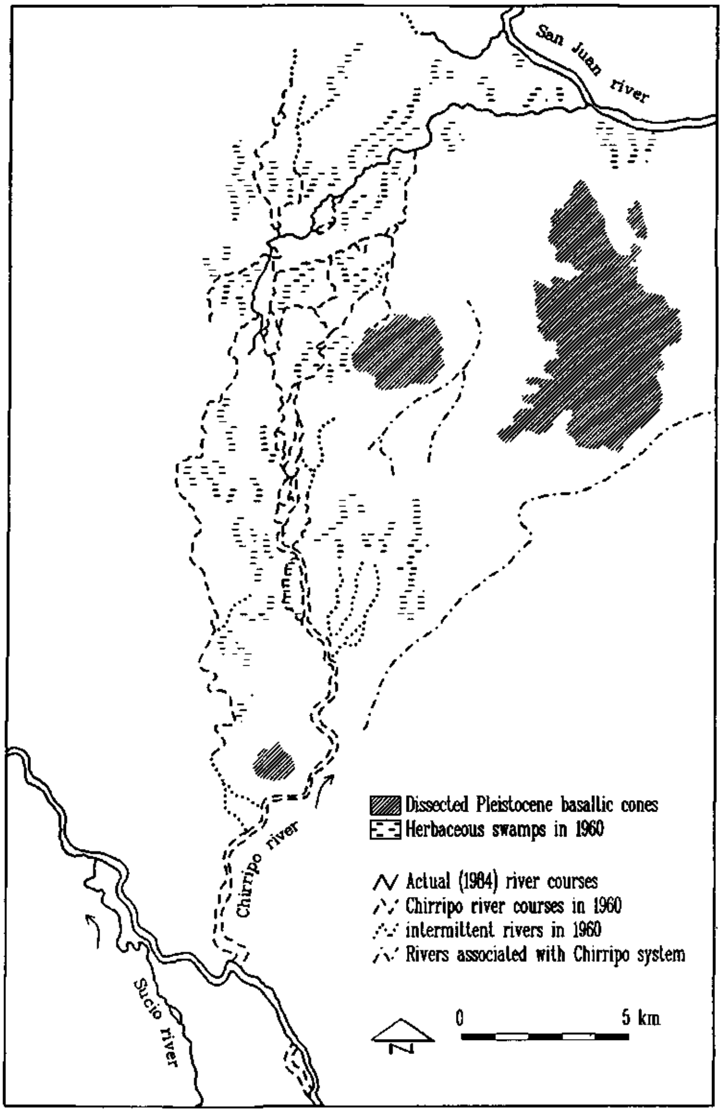


Figure 6: Chirripó river system before (situation 1960) and after 1970 flood (situation 1984).

DEPOSITIONAL MODEL

Based partly on literature and partly on field observations, we first propose a depositional model for the 1970 event. The last eruptive period of the Irazú volcano from 1963 to 1965 deposited thick ash layers in the upper part of river catchments, and destroyed most of the vegetation of the summit area (Barquero, 1977). During and shortly after these eruptions, sediment load of rivers draining the affected area was much higher than normal and was dominated by volcanic ash washed into the rivers by rainfall (ICE, 1964). High rainfall events produced several floods which affected the alluvial fan areas (Waldron, 1967; Kesel and Lowe, 1987). Also lower areas must have been influenced by these floods. We speculate that part or all of the sandy sediments which show hardly any soil formation in the former anastomosing downstream area were deposited between 1963 and 1970 by such floods. This must have influenced subsequent floods, impeding rapid discharge of water and sediment. In December 1970 exceptional rainfall caused large-scale inundations in the whole Limón basin. The Chirripó river bed was choked with sediment downstream from the transition from a braided to a meandering channel, causing a sudden shift of 4 km toward the west, where it joined the Sucio river (Fig.6). It appears that during the flood the river extended its braided system further downstream, but that due to the change in channel geometry and gradient at this point inundations occurred, prograding downstream.

DISCUSSION AND CONCLUSIONS

The climate of the area has been rather stable during the Holocene (Bartlett and Barghoorn, 1973). Extension, morphology, grain size and sedimentological structure of the about 2000 yrs old deposits and the 1970 deposit are alike. Therefore, it seems logical to assume that they were deposited by similar events.

Reagan (1987) describes a relatively big eruption of the Turrialba volcano about 2000 yrs ago, which was preceded by a long period of relative quietness. Since dating of sediments underlying the uppermost Yucatica and Lomas de Sierpe deposits all gave ages slightly over 2000 yrs, it seems reasonable to relate them to this eruption. However, also Barva volcano erupted about 2300 and 1800 yrs BP (E. Fernández, pers.com. 1994), while eruption history of Irazú volcano before 1700 BP is unknown. It is possible that part of the about 2000 yrs BP deposits derived their sediments from other eruptions, especially those in the western part of the central Limón basin (Fig.2).

During several years or decades erosion of the erupted material during rainstorms must have caused frequent flooding of downstream areas, and watersheds influenced by rivers which rose in the summit area of the Turrialba volcano must have been highly dynamic. Most floods probably were similar to the Chirripó 1970 event with only fluvial processes. However, the deposits in the Yucatica area indicate that also hyperconcentrated flows may have occurred in the more proximal areas. As indicated by the white loam unit in the Yucatica area, deposition of the two younger sand sheets was interrupted with a short period of relative quietness.

When comparing the Yucatica with the Lomas de Sierpe area, the much wider extension of deposits in the Yucatica area is remarkable. While in the Yucatica area the middle and upper sheet sand units which slightly differ in age are deposited on top of each other, in the Lomas de Sierpe area deposits which we assume (based on soil development) differ only slightly in age are found next to each other. We ascribe this to a much wider extension of overbank deposits in the more proximal areas (Yucatica area) compared to distal parts (Lomas the Sierpe area).

The studied deposits are different from normal deposits of meandering or anastomosing rivers, but do show some similarity with the Platte and Bijou Creek types for distal braided river deposits described by Miall (1977; 1985). This is in agreement with our assumption that the braided part of the river systems extends further downstream during extreme floods.

Formation of the fluvial landscape is determined by the flatness of the area and the rain forest cover, and is influenced by active volcanism and extreme rainfall events. The studied deposits would not have formed without exceptional sediment supply by volcanic eruptions, nor without high rainfall events which are capable of transporting sediment deep into the alluvial plain. The forest vegetation which we assume was present on levees at time of deposition, apparently did not hamper strongly extension of the deposits, since thick sand layers are found up to 300 m from the former channel. Further away sediments often cannot be related to the choking event. However, this does not mean that no deposition took place. If sediment layers were thin, they may have been mixed with under- and overlying deposits by faunal activity.

Flooding after volcanic eruptions and deposition of large sand bodies may disturb the stable channels systems of meandering and anastomosing rivers. Especially the combination of an anastomosing distal part of a watershed and the occurrence of highly variable sediment loads seems to favour abrupt changes in a river system, or may even be essential to produce such changes.

Channel sediments deposited during floods are often poorly preserved in the sedimentary record, since the river maintains its course and start eroding such sediments as soon as deposition ends (e.g. Pitlick, 1993). In the flat central part of the Limón basin flood level apparently is high enough to cause channel shift, and channel deposits are preserved. Sandy overbank or crevasse-splay deposits in distal parts of watersheds are usually thin and rapidly mixed with older sediments by soil fauna, hampering their recognition in the depositional record. In the study area at least some are thick and can be recognized as individual deposits. Furthermore, point bar deposits hardly occur in the area, probably because the meandering rivers are periodically choked with sediments approximately downstream from their transition with braided river systems as a consequence of extremely high sediment supply by volcanic eruptions.

Volcanic eruptions not always result in drastic changes in a river system, however. Age of the <230 yr BP deposit and its proximity to the Tortuguero river suggest that the deposit may be related to the last (moderate) eruptions of Turrialba volcano from 1864 to 1866

(Reagan, 1987). Since no other sediments related to this eruption were found, it appears that in this case the effect of eruptions was limited to deposition of voluminous crevasse splays in the distal part of the affected watersheds.

About 15% of the Holocene surface sediments in the study area is composed of the described deposits. It is obvious that these deposits by exceptional floods after eruptions have a strong impact on landscape formation in the central part of the Limón basin. In intervening periods landscape formation appears limited to incision of newly formed drainage gullies, swamp formation in areas whose drainage has been blocked by channel fill deposits, and limited lateral migration of remaining river channels. In other areas snowmelts and landslides may have similar effects which suggest that our conclusions have a more than regional validity.

Although the studied time span is limited, data presented in this paper suggest that in the fluvial plains short phases of highly active landscape building due to episodic sediment supply by volcanic activity are separated by longer relatively inactive periods, as is suggested for the alluvial fan by Kesel and Lowe (1987), and for the coastal plain (chapter 3). For example, in the Yucatica area large amounts of sediments were deposited in a time span of maximally 500 years. In the following 2000 years hardly any sediment was deposited and landscape formation was limited to dissection of the sand deposits by small streams. Occurrence of older sheet sand deposits found both in the Yucatica and Lomas de Sierpe areas indicates that similar events occurred earlier, so the described processes appear to occur episodically.

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CHAPTER 5. ANDISOL FORMATION IN A HOLOCENE BEACH RIDGE PLAIN UNDER THE HUMID TROPICAL CLIMATE OF THE ATLANTIC COAST OF COSTA RICA.

Geoderma 57:423-442 (1993)

A. Nieuwenhuyse, A.G. Jongmans and N. van Breemen

Andisol formation in a Holocene beach ridge plain under the humid tropical climate of the Atlantic coast of Costa Rica

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ABSTRACT

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Soil formation has been studied in relation with time in a 5000-year old chronosequence on volcaniclastic beach ridges of the perhumid tropical Atlantic coast of Costa Rica. All soils are under tropical rainforest. Drainage conditions change by subsidence from excessively drained in the two youngest soils to imperfectly drained in the two oldest soils. Parent material is rather homogeneous andesitic sand with a volcanic glass component of less than 10%. It has been found that under these conditions Andisols form within 2000 years. Imperfect drainage caused mottling and accumulation of iron-coatings, as well as the formation of a thin O-horizon in the oldest profiles. Sand content of the soils decreases regularly with soil age, while the amount of fine material increases concurrently. The increase in fine material and the accumulation of organic matter cause an increase of CEC and andic properties, and a decrease in bulk density and pH with soil age. Depth of biological influence increases with soil age, but soil faunal activity is hampered in the oldest three profiles, probably by imperfect drainage. Due to the extreme leaching conditions, the sum of exchangeable cations is less than 2 cmol+ kg^{-1} in the B-horizons of the older soils, notwithstanding the presence of a considerable amount of weatherable primary minerals.

INTRODUCTION

Classification of volcanic-ash derived soils and soils with similar properties has recently been modified. In the Soil Taxonomy, they now form a new Order, the Andisols (Soil Survey Staff, 1990). In order to key out as Andisol, the soil material must have "andic soil properties", which are defined as follows:

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(1) Oxalate extractable Al (Al_o) +oxalate extractable Fe (Fe_o) $\geqslant$2.0% *and*

(2) Bulk density of the <2 mm fraction, measured at 1/3 bar water retention, $\leqslant$0.90 Mg m^{-3} *and*

(3) Phosphate retention >85%.

Special criteria have been defined for young soils which do not meet these requirements but do contain a certain amount of volcanic glass and some oxalate extractable Al and Fe. The minimum content of glass in this requirement can be expressed as: percentage glass $> 35 - 15(Al_o + 1/2\ Fe_o)$.

The rate of formation of Andisols has been studied by various authors under a variety of environmental conditions, but mostly in the humid temperate climates of Japan and New Zealand (Theng, 1980; Wada, 1985). Although some work was done on weathering and neoformation of minerals in Andisols in the humid tropics (summarized in, e.g., Wada, 1989), little is known about the relation of soil formation to time in the humid tropics.

Soils developed in volcaniclastic deposits often have recieved later ash additions, which hampers a straightforward interpretation of the effect of time on soil formation (e.g. Allbrook and Radcliffe, 1987; Bleeker and Parfitt, 1974).

The amount of clay-size materials in volcaniclastic deposits generally increases with soil age (Lowe, 1986). However, the amount of fine material is also influenced by other factors. Clay contents tend to be lower at higher altitudes, due to slower weathering at lower temperature (e.g. Chartres and Pain, 1984). Mineralogy, particle size and porosity of the parent material may play an important role, too.

Under a cool humid climate in Japan 100–500 years were needed for the formation of a clear AC horizon sequence, while an ABC horizon sequence was found only in soils older than 1000 years (Yamada, 1968 in: Wada, 1985). In andesitic–dacitic Quaternary volcaniclastic deposits above 600 m altitude in the dry western part of the USA (Neall and Paintin, 1986), Bw horizons were present in soils older than about 450 years, illustrating that horizon differentiation can be rapid in these parent materials in spite of unfavorable (dry) climatic conditions. In a humid tropical environment differentiation of soil horizons would be expected to occur more rapidly than in the western USA.

The aim of the present study is to investigate Andisol formation in sand of volcanic origin in the humid tropical Atlantic lowland of Costa Rica. A series of beach ridges of different age provided an excellent opportunity to study the effect of soil age on Andisol formation. A second paper (Nieuwenhuyse et al., 1992) focusses on the mineralogical aspects of weathering and neoformation of minerals in this chronosequence.

SITE CONDITIONS

The study area is located in the Limón basin, a sedimentation basin of tectonic origin (Weyl, 1980). In the northeastern part of this basin a sandy beach ridge plain is found. Beach ridges are exposed in a relatively narrow (< 3 km) coastal strip, while further inland they are overgrown by peat or covered by alluvial deposits. Based on a previous soil survey, eight sampling sites were selected (designated AT1 to AT8) in the Tortuguero National Park, along a 2.5 km transect of progressively older beach ridges perpendicular to the coast (Fig. 1).

Beach ridges close to the actual coastline are at < 3 m above sea level, while those further inland are slightly lower, due to the ongoing subsidence of the Limón basin and the Holocene sea-level rise.

Parent material of all soils is well sorted andesitic sand (SiO_2 contents vary from 52 to 60%) of uniform texture (median value M.50 between 150 and 250 μm). The sediments are derived from the Central Cordillera, where active volcanism provides an episodic supply of sandy sediments (Nieuwenhuyse and Kroonenberg, 1992).

The parent material is mineralogically homogeneous, it contains many an-

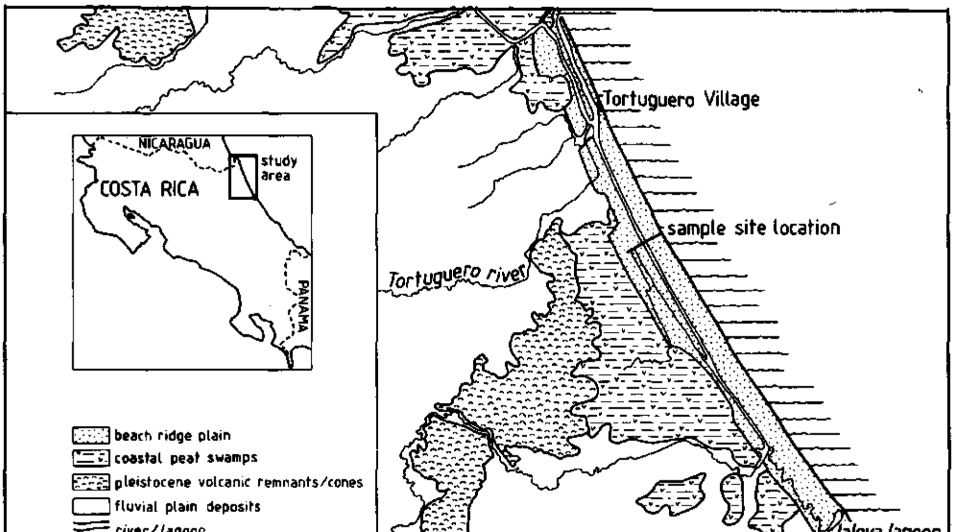

Fig. 1. Location of study area and sampling sites.

desitic rock fragments, composed of plagioclase, pyroxene and magnetite phenocrysts in a matrix in which volcanic glass regularly can be recognized. Few microcrystalline rock fragments occur, composed of quartzite, sandstone and hornstone. Further, many plagioclase and common to many pyroxene (both augite and hypersthene) and very few magnetite, hornblende, biotite and olivine mineral grains are found, sometimes containing volcanic glass inclusions. Finally, common spherical clay bodies are found. In a number of the andesitic rock fragments phenocrysts have been transformed into such clay bodies. Based on point counting in thin sections, the total volcanic glass content is estimated to be less than 10%, on a volume basis (Nieuwenhuyse et al., 1992). Shell fragments were not found, although they may be expected in beach sediments.

The ratios of andesitic rock fragments to pyroxenes vary in the different beach ridge sediments, and their chemical composition varies accordingly (Nieuwenhuyse and Kroonenberg, 1992). Higher concentrations of andesitic rock fragments are accompanied by lower contents of pyroxene and rather high (57–58%) SiO_2 content, while relatively high pyroxene and low andesitic rock fragment contents are reflected in lower (52–54%) SiO_2 contents.

No volcanic ash particles larger than 10 μm were observed in a 4450 year old peat deposit located between the Central Cordillera and the study location. However, very small additions of volcanic dust to the sites cannot be excluded (Nieuwenhuyse, unpubl. data).

At the actual beach, the parent material is homogenized to a depth of about 0.9 m by sea turtles which nest each year in large numbers in the Tortuguero park. Also crabs contribute to this homogenization.

All sites are under humid tropical forest and presumably always have been so. There is no evidence of agricultural use in historical or pre-columbian times, but some logging took place before 1970. At the AT1 and AT2 sites, vegetation composition is influenced by sea salt inputs.

Annual rainfall (1978–1990) ranged from 4100 to 6600 mm with an average of 5350 mm. Mean potential evapotranspiration, based on Penman (using climatic data from nearby stations) amounts to between 12 and 48% of the mean monthly rainfall, and to 22% of the mean annual precipitation (Ta-

TABLE 1

Rainfall and Penman potential evapotranspiration for Tortuguero in millimeters

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Total
Pot. evapotransp.	85	86	113	107	111	96	95	103	106	102	83	81	1168
Precipitation	455	302	235	286	453	403	591	499	326	460	668	666	5345
Surplus	370	216	122	179	342	307	496	396	220	358	585	585	4177

Source: *Instituto Meteorológico Nacional*, Costa Rica.

ble 1). However, each year dry periods varying between one and three weeks occur, causing the younger soils and the topsoil of the older soils to dry out.

Hydraulic conductivity of all soils is very high providing for a rapid infiltration. During many visits to the sites surface runoff has never been observed.

Based on data from nearby stations at the same altitude, mean annual temperature is estimated to be 25–26°C. Difference between hottest (June) and coolest month (January) is 2–2.5°C.

METHODS

Study sites were chosen on a flat part of the top of each ridge. At each site a pit was dug to describe and sample the soils. The soils were described according to FAO (1977) and classified according to Soil Survey Staff (1990). In and around the profile pits evidence of faunal activity was annotated.

To reduce spatial variability, bulk samples of each horizon were collected over a width of about 50 cm along the pit wall. Samples below the ground water table were taken with an auger. After removal of roots and litter and passing the samples over a 2 mm sieve to obtain a well-mixed sample, part of each bulk sample was sealed into plastic bags to keep it at field moisture and was stored at 4°C. Another part of each sample was air-dried and stored at ambient temperature. Field moisture was determined for use as a correction factor to express analyses of field-moist and air-dried samples on the same basis.

Physical analysis

Texture was determined in the field-moist sample after destruction of organic matter with H_2O_2, and dispersion using a mixture of 20 ml sodium polyphosphate (5%) and 20 ml NH_4Cl (10%). Clay (<2 μm) contents were measured by the hydrometer method (Gee and Bauder, 1986). Many of the samples did not disperse well, either in the above mentioned dispersants, or in HCl (pH 4) or NaOH (pH 10). The clay contents reported in this study only refer to samples which did not show flocculation. Sand contents (>53 μm) were determined by sieving.

Bulk density was determined at two to four different depths by measuring the oven-dry weight of 100 ml core samples taken in triplicate.

Moisture retention at 1.5 MPa was estimated in duplicate by equilibrating for 72 h a slurry made from field moist soil on a pressure plate.

Chemical analysis

Soil pH was measured 30 min after intensive stirring for 2 min of a 1:5 (w/w) field moist soil–water mixture. Organic carbon was determined in air-dry samples, using the Walkley–Black wet digestion method. Incomplete recovery was compensated for with a correction factor of 1.3 (Walkley, 1947). Organic matter contents were estimated by multiplying measured C values by 1.72. pH-NaF was measured in field moist samples, after exactly 2 min stirring in 1N NaF (1 g soil to 50 ml NaF).

Exchangeable bases were measured in triplicate in the leachate of a 1M ammonium acetate (pH 7) solution. CEC was determined in triplicate by measuring the absorbed $NH4^+$ after washing with ethanol and replacing ammonium by potassium, using 1M KCl. Exchangeable acidity was determined by leaching the soil sample with 1M KCl and titrating the leachate with 0.1M NaOH. For both methods field moist samples were used.

Phosphate retention was determined according to Blakemore et al. (1987) in air-dry samples. Acid ammonium oxalate extractable iron (Fe_o) and aluminium (Al_o) were determined by shaking a suspension of 1 g of air-dry soil and 100 ml of oxalate solution for 4 h in the dark (Blakemore et al., 1987) and measuring Fe and Al by AAS.

All results have been recalculated on oven-dry (105°C) basis.

Micromorphological analysis

Undisturbed samples of the major soil horizons were taken for preparation of thin sections (7 by 7 and 2 by 2 cm). In the deeper soil horizons of the older profiles disturbance of the soil during sampling could not be prevented. Samples were treated with a fungicide and sealed into plastic bags immediately to maintain field moisture. Field-moist samples were impregnated with acetone without prior drying according to Miedema et al. (1974). Thin sections were made following the method of FitzPatrick (1970), and described according to Bullock et al. (1985).

Age determinations

The ages of the different soils were estimated using ^{14}C datings of organic rich deposits in the coastal plain and between some of the beach ridges (Nieuwenhuyse and Kroonenberg, 1992).

RESULTS

The ages of the different soils (Table 2) range from less than 100 years in the recent beach ridge to maximally 5000 years in the oldest soil, AT8. With increasing age the soils show an increasing profile development as evinced by (i) a change in colour from dark gray in the unchanged parent material to dark yellowish brown B-horizon material, (ii) increasing thickness of the B-horizon from nil in the three youngest soils to more than 50 cm in the oldest

TABLE 2

Soil classification (Soil Survey Staff, 1990), estimated age (Nieuwenhuyse and Kroonenberg, 1992), average depth of groundwater level and drainage (FAO, 1977) of the investigated soils

Soil	Classification	Age (years B.P.)	Av. depth groundwater (cm)	Drainage
AT1	Typic Tropopsamment	< 100	+ 150	excessively drained
AT2	Typic Tropopsamment	<200	+ 130	excessively drained
AT3	Typic Tropopsamment	< 500	70	somewhat exc. drained
AT4	Typic Hapludand	appr. 2000	70	mod. well drained
AT5	Acrudoxic Hapludand	2000–5000	95	well drained
AT6	Acrudoxic Hapludand	2000–5000	90	well drained
AT7	Aquic Hapludand	2000–5000	55	imperf. drained
AT8	Hydric Haplaquands*	2000–5000	30	imperf. drained

*Tentatively. 1.5 MPa water retention value has not been determined.

soils, and (iii) increasingly fine textures from fine sand in the recent beach ridges to loam in the oldest surface horizons (Table 3). The A-horizons tend to increase in thickness and in organic matter content from AT1 to AT7. The two oldest soils show an increasingly thick organic surface horizon, and display iron oxide mottling at depths of about 40 cm (AT7) and 20 cm (AT8) (Tables 3 and 4).

Field observations show that leaf-cutter ants and crabs are important burrowing animals in the three youngest profiles, while in the older profiles earthworms and rodents are active. In the AT7 and AT8 profiles very few worms were observed.

Micromorphological changes with age include (i) increasing degree of alteration of both andesitic rock fragments and pyroxene and plagioclase mineral grains from weakly weathered in the A horizons of the three youngest soils to strongly weathered in the topsoil of the two oldest soils, (ii) increasing content of fine materal over a greater depth, consisting of a mixture of organic matter and pale yellow isotropic material, displaying an undifferentiated b-fabric; (iii) changing of intermineral grain microstructures in the three youngest soils into granular and spongy microstructures in the older profiles. (Sub)angular blocky and locally massive structure dominate A horizons of AT7 and AT8; (iv) decreasing amount of mineral excremental infillings in the topsoils of the oldest profiles; (v) occurrence of (non) laminated isotropic orange-brown to yellowish brown (hypo)coatings in the B and BC horizons of AT7 and AT8. In the oldest soil these coatings cover mineral excremental infillings (Figs. 2 and 3).

Physical changes with increasing age involve (i) decreasing sand and increasing clay contents, (ii) increasing water retention at 1.5 MPa and (iii) a decrease in bulk density from about 1.2 Mg m^{-3} in the parent material to

TABLE 3

Major macromorphological characteristics[1]

Profile	Hori-	Depth	Boun-dary	Color (moist)	Mottles[2]	Field texture	Structure		Consis-tence
							type	grade	
T1	A	0	c,s	10YR 2/2	-	fs	s.grain	-	l
	CA	4	c,s	10YR 2/3	-	fs	s.grain	-	l
	C	20–80		10YR 3/1	-	fs	s.grain	-	l
AT2	A	0	a,s	10YR 2/1	-	ls	crumb	mod.	l
	AC	3	c,s	10YR 2/2	-	fs	s.grain	-	l
	C1	12	c,s	10YR 2/3	-	fs	s.grain	-	l
	C2	45–80		10YR 3/1	-	fs	s.grain	-	l
AT3	A1	0	c,w	10YR 2/1	-	sl	crumb	mod.	vfr
	A2	3	c,w	10YR 2/2	-	sl	sub.bl	weak	vfr
	AC	9	c,s	10YR 2/3	-	ls	s.grain	-	l
	CA	16	g,s	10YR 3/3	-	fs	s.grain	-	l
	C	27–55		10YR 3/2	-	fs	s.grain	-	l
AT4	A	0	c,s	10YR 2/3	-	l	crumb	mod.	fr
	AB	12	c,s	10YR 2/3	-	l	crumb	str.	vfr
	Bw	19	c,s	10YR 4/3	-	sl	sub.bl.	weak	fr
	BCw	35	g,s	10YR 4/3	-	ls	massive	-	fi
	CB	50	g,s	10YR 4/1	-	fs	massive	-	fi
	C	85–120		10YR 4/1	-	fs	s.grain	-	l
AT5	A1	0	c,s	10YR 2/2	-	l	crumb	str.	vfr
	A2	5	c,s	10YR 2/3	-	l	sub.bl.	str.	vfr
	AB	11	c,s	10YR 2/3	-	sl	sub.bl.	mod.	vfr
	BAw	18	g,s	10YR 3/3	-	sl	sub.bl.	weak	fr
	Bw	30	g,s	10YR 4/3	-	ls	sub.bl.	weak	fi
	BCw	50	g,s	10YR 4/2	-	fs	massive	-	vfi
	C	80–150		10YR 4/1	-	fs	s.grain	-	l
AT6	A	0	c,s	10YR 2/3	-	l	crumb	str.	vfr
	Bw1	17	c,s	10YR 4/3	-	sl	sub.bl.	weak	fr
	Bw2	26	c,s	10YR 4/3	-	sl	sub.bl.	weak	fr
	Bw3	38	c,s	10YR 3/4	-	ls	massive	-	fi
	BCw	48	g,s	10YR 4/2	-	ls	massive	-	fi
	BCwr	80	g,s	10YR 4/1	c2d(o)	fs	massive	-	fi
	C	105–120		N 3/1	-	fs	s.grain	-	l
AT7	O	2	a,s	7.5YR 2/3	-	l	crumb	mod.	(nd)
	A1	0	c,s	10YR 3/4	-	l	sub.bl.	weak	(nd)
	A2	12	c,s	10YR 3/3	-	l	sub.bl.	weak	(nd)
	Bw	25	c,s	10YR 4/3	-	sl	sub.bl.	weak	(nd)
	Bwr	36	c,w	10YR 4/2	c1d(o)	sl	massive	-	fi
	BCwr	53	g,s	10YR 4/1	2d(o)	sl	massive	-	vfi
	CBr1	75	g,s	10YR 4/1	c2d(o)	ls	massive	-	fi
	CBr2	90	g,s	10YR 4/1	c2d(r)	fs	massive	-	fi
	CBr3	110	g,s	N 3/1	c2d(r)	fs	s.grain	-	l
	C	150–170		N 3/1	-	fs	s.grain	-	l
AT8	O	5	a,s	10YR 2/3	-	(nd)	crumb	mod.	(nd)
	A	0	c,s	10YR 3/4	-	l	sub.bl.	weak	(nd)
	Bwr	20–35+		10YR 4/2	c2d(o)	l	massive	-	(nd)

[1]Abbreviations according to Soil Survey Staff, 1951, p. 139.
[2]Color mottles: (o) = orange, (r) = red. (nd) = not determined.

TABLE 4

Selected chemical properties of the soils

Soil	Horizon	Sample depth (cm)	Clay ($g\ kg^{-1}$)	pH-H_2O	Organic matter ($g\ kg^{-1}$)	CEC	Sum cations ($cmol+kg^{-1}$)	Exch. acidity
AT1	A	0–4	30	5.6	51	13.0	11.2	nd
	CA	4–10	20	5.7	23	13.5	6.8	nd
	C	20–25	–	6.4	3	12.0	5.3	nd
	C	50–55	0	6.7	2	11.6	4.3	nd
AT2	A	0–2	60	5.7	131	26.1	17.5	nd
	AC	4–8	–	5.8	53	14.2	8.1	nd
	C1	25–30	10	6.5	21	11.3	3.9	nd
	C2	55–60	–	6.6	2	10.2	4.0	nd
AT3	A1	0–3	60	5.7	103	22.6	7.8	nd
	A2	4–8	50	5.9	61	16.1	3.2	nd
	AC	10–15	–	5.8	46	11.8	2.5	nd
	CA	17–23	20	6.1	20	9.2	2.0	nd
	C	28–34	10	6.3	8	9.3	2.2	nd
	C	50–55	0	6.4	4	8.9	2.4	nd
AT4	A	0–3	180	6.0	128	33.9	8.9	0.6
	A	6–10	160	6.0	107	29.8	5.1	nd
	AB	13–17	150	5.9	88	25.7	4.2	nd
	Bw	25–30	110	6.2	42	23.7	1.5	0.3
	BCw	45–50	–	6.1	9	nd	nd	nd
	CB	50–70	–	6.4	6	18.4	2.7	nd
	C	100–110	–	6.5	2	18.7	4.5	nd
AT5	A1	0–3	240	5.4	148	38.7	5.2	0.6
	A2	5–8	170	5.5	121	35.2	2.1	0.5
	AB	12–16	–	5.6	105	30.7	1.6	0.3
	BAw	22–26	–	6.3	51	21.5	1.2	nd
	Bw	40–45	–	6.7	25	16.1	1.0	0.1
	BCw	70–80	–	7.4	7	13.5	2.2	0.1
	C	135–145	–	nd	3	18.1	4.1	nd
AT6	A	0–3	170	5.5	120	35.5	3.2	0.3
	A	6–12	240	5.3	117	34.2	1.6	0.3
	Bw1	18–24	–	5.9	50	26.1	2.0	nd
	Bw2	30–35	–	6.5	30	25.6	1.8	nd
	Bw3	41–46	–	6.5	17	20.0	1.5	nd
	BCw	60–65	–	6.7	10	20.8	1.5	nd
	BCwr	85–95	–	6.6	9	25.0	2.6	0.2
	C	110–120	–	6.4	1	16.4	2.8	nd
AT7	O	2–0	–	4.6	321	79.1	7.9	4.0
	A1	2–8	270	4.9	149	53.2	1.7	1.5
	A2	14–20	–	5.5	147	28.5	1.3	0.2
	Bw	28–33	–	6.1	46	24.0	1.3	nd
	Bwr	38–48	–	6.3	34	18.7	1.5	nd
	BCwr	60–70	–	6.0	25	19.2	2.5	nd
	CBr	90–100	–	6.7	9	24.2	2.9	0.2
	C	150–160	–	6.1	6	16.2	nd	nd
AT8	O	5–0	nd	4.5	413	79.6	7.9	5.9
	A	1–11	nd	5.1	110	56.1	3.3	1.0
	Bwr	15–25	nd	6.0	67	41.3	2.7	0.5
	BCr1	30–45	nd	6.3	35	36.3	5.3	0.3
	BCr2	60–75	nd	6.7	24	29.7	6.7	0.3
	CBr	90–100	nd	6.6	8	nd	8.3	0.1

nd = not determined; – = not determined because of flocculation.

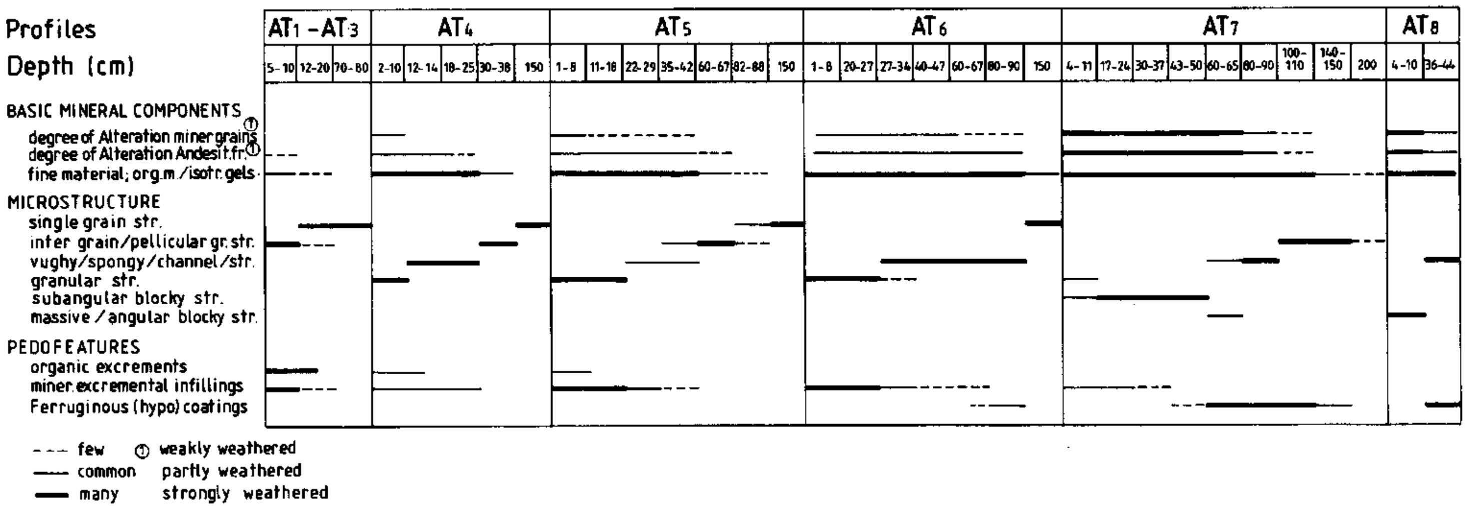

Fig. 2. Selected micromorphological characteristics.

values between 0.9 and 0.3 Mg m^{-3} in the solum of increasingly older soils (Fig. 4, Tables 4 and 5). Chemical development with increasing age is characterized by (i) a decrease in pH-H_2O from 6.5 to 7 in the unchanged parent material to around 5 in the oldest surface soils, (ii) increasing CEC and decreasing exchangeable base cations, (iii) a gradual increase in pH-NaF from just over 8 in recent beach ridge material to between 10 (C-horizon) and 11.5 (surface horizon) in the oldest soils (Tables 4 and 5). Furthermore, P-retention increases from about 25% in the parent material to values over 95% in the A- and B-horizons of AT4 to AT8, and Al_o increases from less than 5 g kg^{-1} in the two youngest soils to over 30 g kg^{-1} in B horizon of the two oldest profiles.

Classification (Soil Survey Staff, 1990) yields Tropopsamments for the three youngest soils and Hapludands for the older soils (Table 2).

DISCUSSION

All soils have developed from essentially the same kind of parent material. No significant differences in the soil forming factors climate, vegetation and human influence occur and rejuvenation by volcanic ash addition is absent or negligible. All soils presumably started their development under conditions of excessive drainage. Due to lowering of elevation, drainage decreased with soil age to imperfectly drained (Table 2). As a result, the differences in soil development can be ascribed to two soil forming factors: age and gradually changing of drainage conditions. The following soil forming processes will be discussed in some detail: organic matter accumulation, biological influence and weathering of primary minerals and neoformation of secondary minerals.

Organic matter accumulation

The organic matter contents of the dark surface soils formed in beach ridge sand (Tables 3 and 4), are high and typical for volcanic ash soils in humid tropical environments (e.g. Wada, 1985; Mizota and Van Reeuwijk, 1989). The relatively high organic matter contents of Andisols is due to the formation of stable complexes of organic material with Al and Fe (Mizota and Van Reeuwijk, 1989; Wada, 1989). Pyrophosphate extractable aluminium and iron increase with soil age from 0% in the AT1 profile to 1.7 and 1.8%, respectively, in the A horizon of the AT7 soil (Nieuwenhuyse et al., 1992). This indicates that the formation of such complexes indeed takes place. The wet climate and the relatively shallow water tables of AT3–AT8 may hamper soil aeration and could also contribute to decreased decomposition of organic matter.

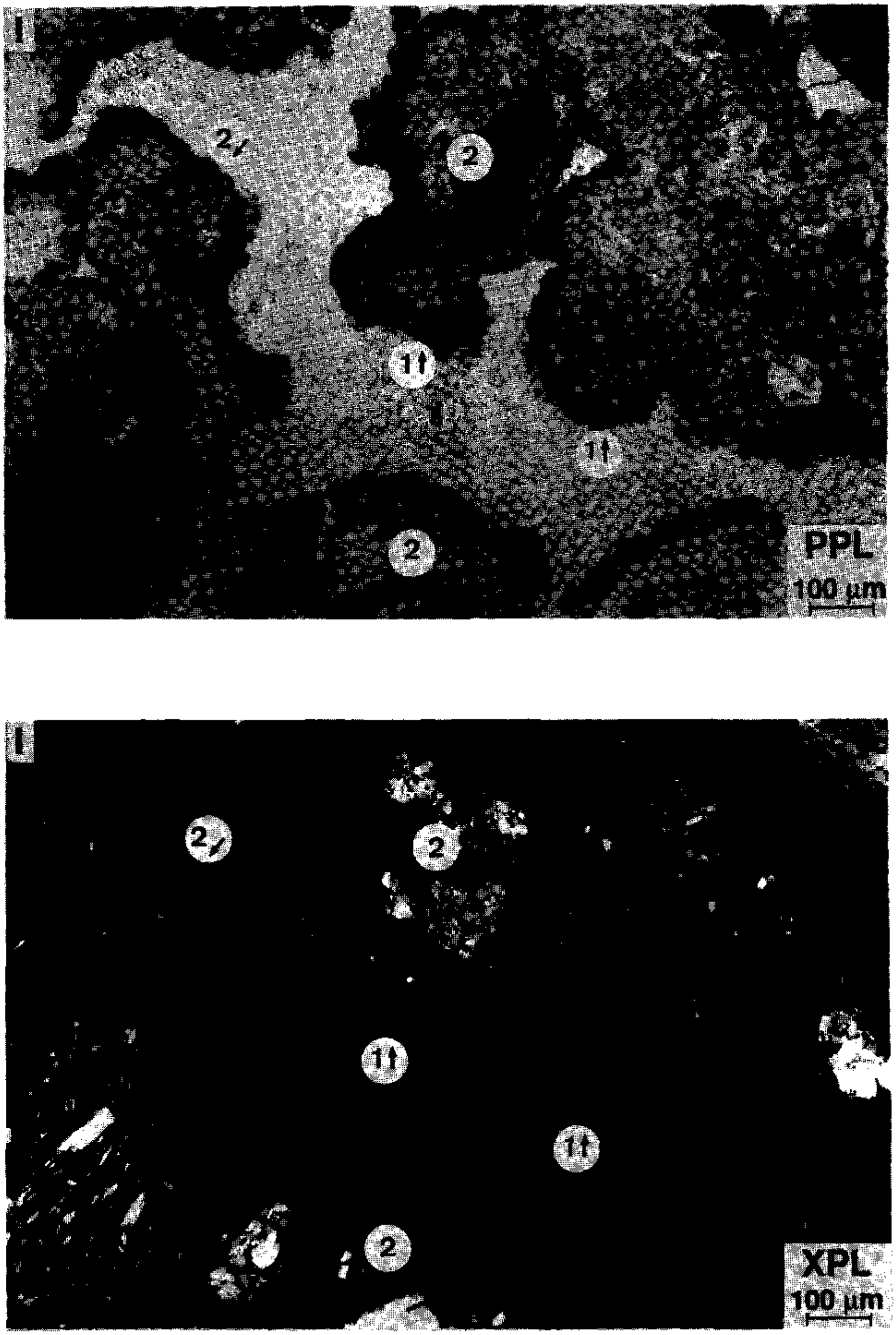

Fig. 3. For caption see p. 437.

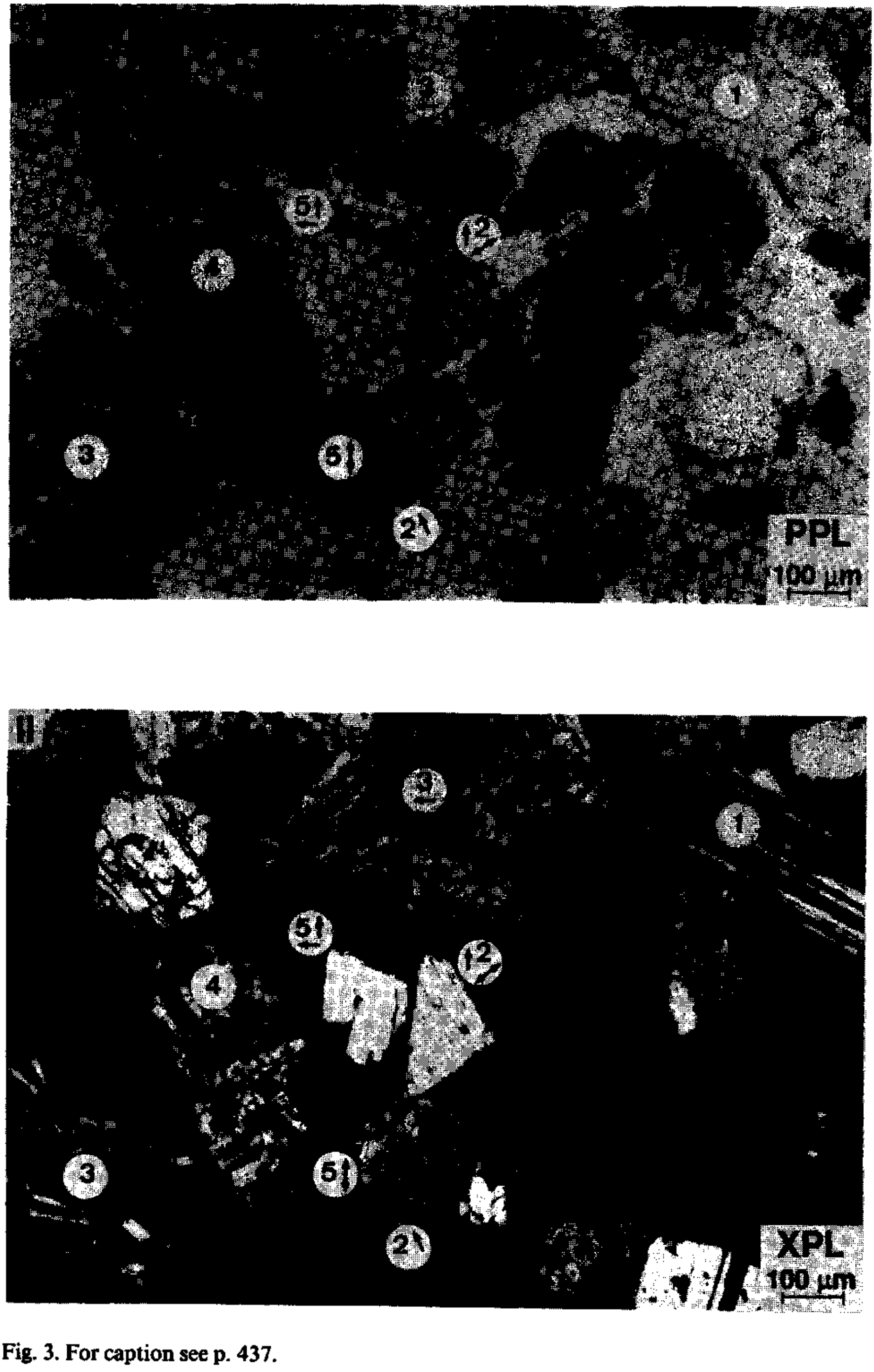

Fig. 3. For caption see p. 437.

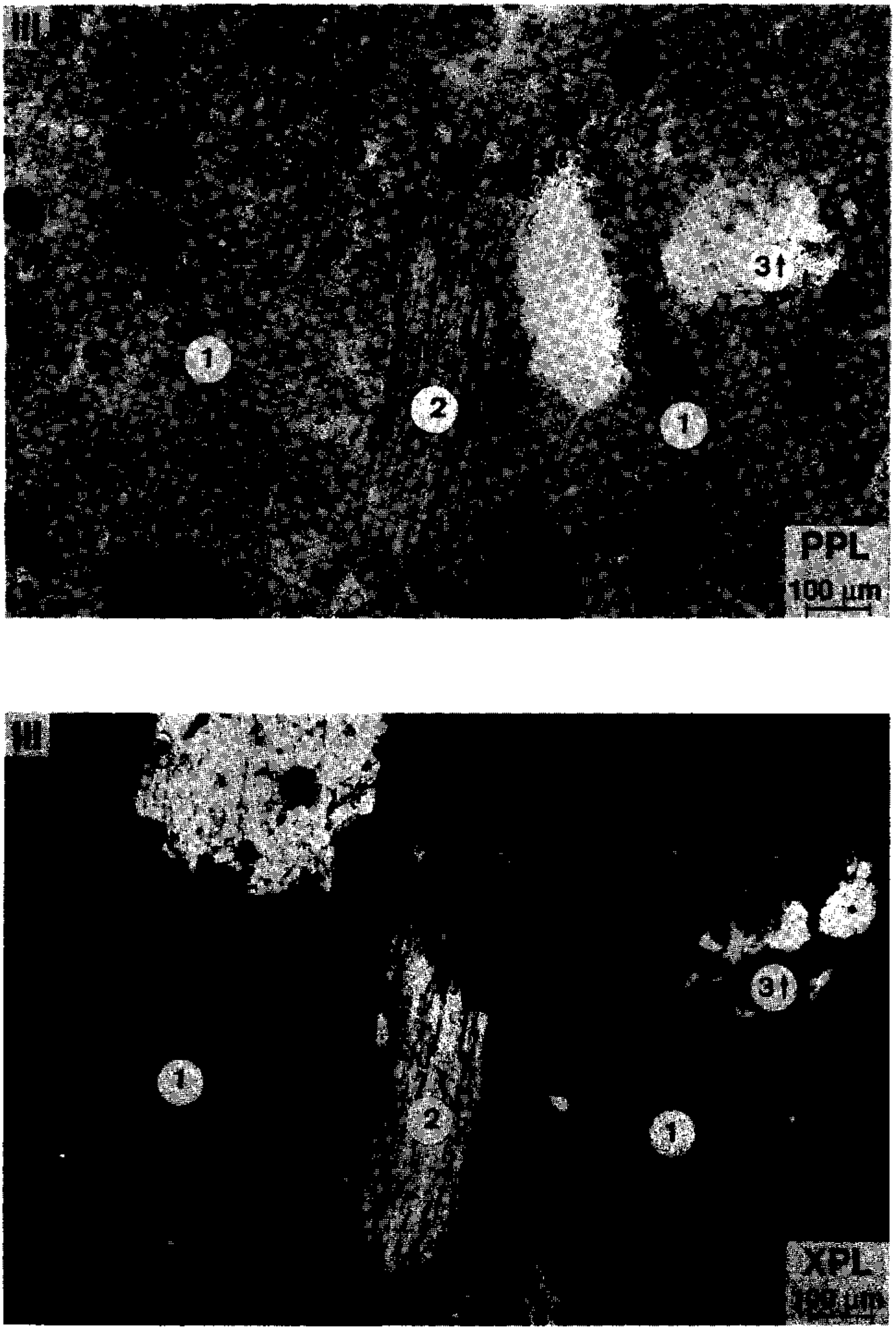

Fig. 3. For caption see facing page.

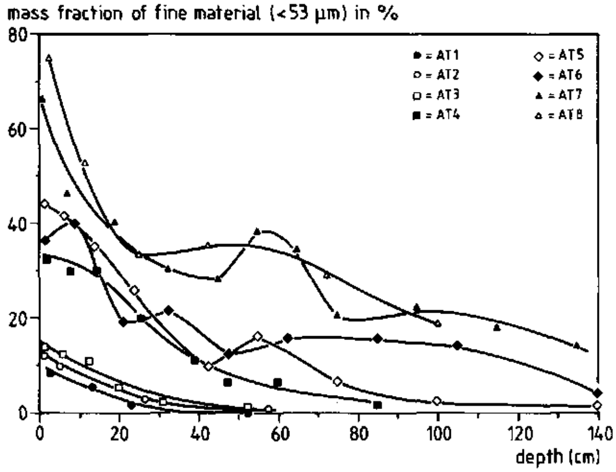

Fig. 4. Content of fine material (< 53 μm).

The organic surface horizons of the two oldest profiles reflects the imperfect drainage of those soils. The lighter color of the A-horizons of these soils suggests a different form of organic matter, perhaps also on account of a different decomposition pathway caused by poor aeration. The low bulk density of the AT7 profile as compared to the younger soils, points to a different form of organic matter accumulation. Possibly more hydrous complexes of organic matter and Al are formed under these wetter conditions.

The organic matter accumulation in the upper meter of the profiles has been estimated, using data about horizon sequence, organic matter content and bulk density (Tables 3–5). Values for the AT1 to AT7 soils are 7, 16, 13, 25, 26, 23 and 25 kg organic matter per m^3 soil, respectively. Due to the simultaneous decrease in bulk density, organic matter pools do not increase from AT4 onward, although organic matter mass fractions do increase.

Biological influence

The observed faunal activity and the granular and spongy structures observed by macro- and micromorphology (Table 3, Fig. 2) indicate a strong biological influence over a greater depth with increasing soil age. Apart from

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Fig. 3. Selected micromorphological features. (I) Microlaminated ferruginous coatings (*1*) covering coalescent mineral excrements (*2*) in the B-horizon of AT8. (II) Basic mineral components in the A-horizon of AT1, mainly consisting of unweathered grains: plagioclase (*1*), pyroxene (*2*), andesitic rock fragment (*3*), limpid anisotropic clay bodies (*4*) and some organic fine material (organic excrements) (*5*). (III) Basic ineral components in Ah horizon of AT8, mainly consisting of isotropic fine material (*1*), pellicular altered pyroxene (*2*), and minute residues of plagioclase (*3*). *XPL* = crossed polarized light; *PPL* = plain polarized light.

TABLE 5

Selected andic properties of the soils

	Horizon	Sample depth (cm)	H-NaF	Bulk density (Mg m^{-3})	P ret. (%)	1.5 MPa wat. ret.	Al_o (g kg^{-1})	Fe_o
AT1	A	0–4	8.4	nd	27	130	1	5
	CA	4–10	8.3	1.0	27	70	2	5
	C	20–25	8.3	1.3	22	30	1	6
	C	50–55	8.0	nd	24	30	1	8
AT2	A	0–2	8.2	nd	40	330	3	3
	AC	4–8	9.2	0.8	41	160	3	3
	C1	25–30	8.9	1.2	27	30	2	3
	C2	55–60	8.3	nd	22	30	1	5
AT3	A1	0–3	9.3	nd	62	320	6	4
	A2	4–8	9.9	0.8	70	240	7	4
	AC	10–15	10.5	nd	68	220	7	5
	CA	17–23	10.2	nd	54	100	5	5
	C	28–34	9.0	1.2	40	70	4	6
	C	50–55	8.6	nd	33	30	2	4
AT4	A	0–3	10.5	nd	91	560	14	4
	A	6–10	10.6	0.7	93	560	16	5
	AB	13–17	10.5	nd	94	520	17	5
	Bw	25–30	10.9	0.8	97	400	21	6
	BCw	45–50	10.5	1.1	81	200	15	2
	CB	50–70	9.8	nd	68	130	13	2
	C	100–110	9.2	nd	44	70	5	3
AT5	A1	0–3	10.6	nd	95	640	24	6
	A2	5–8	10.7	0.6	97	580	23	6
	AB	12–16	10.7	nd	97	500	23	6
	BAw	22–26	10.7	0.7	97	470	24	5
	Bw	40–45	10.6	0.9	88	230	18	2
	BCw	70–80	9.2	0.9	57	130	9	1
	C	135–145	9.5	nd	42	90	4	4
AT6	A	0–3	10.5	nd	96	510	21	6
	A	6–12	10.9	0.7	96	640	24	8
	Bw1	18–24	10.8	0.6	97	490	28	7
	Bw2	30–35	10.5	nd	97	500	24	6
	Bw3	41–46	10.5	0.7	92	400	20	4
	BCw	60–65	10.3	0.8	87	250	20	4
	BCwr	85–95	10.0	nd	80	350	16	3
	C	110–120	9.6	nd	46	90	8	4
AT7	O	2–0	8.7	nd	71	1540	6	10
	A1	2–8	10.3	0.3	93	1340	21	20
	A2	14–20	11.0	nd	98	1070	30	12
	Bw	28–33	11.2	0.4	98	740	31	9
	Bwr	38–48	11.2	0.6	97	600	27	7
	BCwr	60–70	10.3	0.6	93	520	26	7
	CBr	90–100	10.2	nd	88	440	22	6
	C	150–160	10.0	nd	75	220	14	5
AT8	O	5–0		nd	62	nd	5	8
	A	1–11	11.7	nd	96	nd	21	21
	Bwr	15–25		nd	97	nd	33	12
	BCr1	30–45	11.6	nd	97	nd	27	7
	BCr2	60–75	11.3	nd	94	nd	24	8
	CBr	90–100	11.1	nd	87	nd	18	5

nd = not determined.

structure development in a shallow (but increasing) surface soil, the three youngest profiles show hardly any horizon differentiation. The A-horizon gradually deepens from AT1 to AT4; in the AT4 to AT6 profiles depth of the A horizons is more or less constant. In the AT7 and AT8 profiles, differences in color between the A and B horizons become less clear. While homogenization of the parent material by nesting sea turtles facilitates study of soil genesis, later biologic activity may obscure some pedogenetic processes. In AT6, the horizon sequence in the topsoil was found to be affected by burrowing activity of a rodent (*Orthogeomys cherriei*), which has brought part of the B-horizon to the surface, as indicated by values for texture (Table 4, Fig. 4). The decreasing structural grade with age from AT5 onwards (Table 3), as well as the development of organic surface horizons point to reduced faunal bioturbation. Both effects can be caused by poor aeration as a result of imperfect drainage. The few earthworms observed and the absence of a granular microstructure and a low amounts of excremental infillings in the AT7 and AT8 profiles also indicate restricted biological activity. Faunal activity might also be hampered by the acidity of their surface horizons, where pH values lower than 5 and high amounts of exchangeable aluminium are found (Table 4). The spongy microstructure in the B-horizons of these profiles which may result from root activity, faunal biological activity or may be due to the buoyant force of water, contribute to the extremely low bulk densities and high moisture retention values at 1500 kPa (Table 5).

Weathering and neoformation

Weathering of volcanic parent material under well-drained conditions and high rainfall generally leads to the formation of short-range order materials such as allophane and Al-humus complexes (Mizota and Van Reeuwijk, 1989; Wada, 1989). Precipitation of Si–Al gels, liberated by weathering of the mineral grains, and accumulation and incorporation of soil organic matter are thought to be the main source of the increasing amount of fine material in the A and B horizons with soil age (Figs. 2–4). Complexes of Al and Fe with organic matter were found to be the main components of the clay fraction of the A horizons, while allophane was the most important clay component in deeper horizons (Nieuwenhuyse et al., 1992). The rate of weathering of andesitic sand, as can be deduced from the decrease of sand content (Fig. 4) is about 30–40% in the topsoil of AT4 and AT5. This is in the same order of magnitude as that of dacito-andesitic pumice under 2000–2500 mm rainfall in Martinique: Quantin et al. (1991) found an increase in material smaller than 0.05 mm from 12% to about 33% in a soil 1670 years old.

The decrease in bulk density with soil age and the concomitant, gradual increase of CEC, moisture retention at 1.5 MPa values and andic properties

(Tables 4 and 5), must be attributed largely to the formation of amorphous secondary material and the accumulation of organic matter.

The CEC of most samples can be accounted for by contents of clay and organic matter, assuming a CEC of 200 cmol+ kg^{-1} for organic matter and 100 cmol+ kg^{-1} for amorphous clay (allophane). However, in many of the subsurface horizons CEC values are higher than can be explained by their clay and organic matter contents. Possibly, the sand-sized clay bodies derived from the parent material contribute to these high CEC values. On the other hand, CEC values in Andisols obtained by conventional methods are often difficult to interpret (Wada, 1989) and a yet unknown phenomena or analytical errors might be the reason for the high CEC values.

The low sum of exchangeable cations in the B horizons of the AT4–AT7 profiles is remarkable. In all horizons still considerable amounts of weatherable primary minerals are present, and except for the topsoils of AT7 and AT8, soil pH values are relatively high. Therefore, a good supply of bases by weathering is likely, and the low amount of exchangeable cations must be attributed to the extremely high continuous leaching. The somewhat higher amount of exchangeable bases in AT8 may be due to imperfect drainage, which may somewhat hamper leaching.

P-retention increases with Al_o content to values of about 20 g kg^{-1} Al_o (Table 5). At higher contents of Al_o, P-retention remains more or less constant. These results suggest that above the limits defined by Soil Survey Staff (1990) for Al_o and Fe_o to classify a soil as Andisol, values for P-retention are rather insignicant, since they do not distinguish any further between the soils. At least for the studied soils, P-retention determinations have little value for classification purposes.

Fe_o contents are rather low and constant throughout the chronosequence, except in the surface horizons of the two oldest soils. Fe_o contents in the young soils and the less weathered subsoils exceed the content of Al_o. Apparantly, oxalate extracts Fe not only from poorly ordered Fe oxides, but also from the parent material components. Whether this iron is extracted from secondary Fe-oxides present in the parent material, or from primary minerals is not known. Oxalate is know to attack magnetite, which may be the source of some of the Fe_o (Borggaard, 1988). The distinctly higher Fe_o values in the surface horizons of AT7 and AT8 are associated with gley mottling and organic complexes. As a consequence, Fe_o is unsuitable for classification purposes in these soils.

The orange brown to yellowish brown (hypo) coatings and the accumulation of iron oxides in and around voids in the BC- (and B-) horizons of the AT6–AT8 may contribute to the firmness of some of the BC horizons of the AT4–AT8 soils (Table 3).

CONCLUSIONS

Mineralogical and textural data on the parent material indicate that the soils of this chronosequence spanning 100–5000 years of age have developed in essentially the same parent material.

Under the perhumid tropical circumstances of the study area, well developed Andisols may form in sandy andesitic parent material with a low volcanic glass content in less than 2000 years. The dominant soil forming factor in the studied chronosequence is soil age, but except for the youngest two profiles drainage plays an (important) role, too.

Increased weathering to greater depth with increasing soil age, results in decreasing sand contents and increasing contents of fine material over a greater depth with soil age. Biological homogenisation increases with soil age, but biological acivity is restricted in the oldest two soils as a result of gleying.

This study suggests that P-retention and Fe_o are of limited use for classifying these Andisols.

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CHAPTER 6. MINERALOGY OF A HOLOCENE CHRONESEQUENCE ON ANDESITIC BEACH SEDIMENTS IN COSTA RICA.

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Mineralogy of a Holocene Chronosequence on Andesitic Beach Sediments in Costa Rica

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ABSTRACT

We investigated weathering and neoformation of minerals in a Holocene (<5000 yr) soil chronosequence on sandy, andesitic, ocean beach ridges in humid tropical Costa Rica using micromorphological, mineralogical, and chemical analyses. Tropopsamments are present on the younger beach ridges and Hapludands are on the older ones. The parent materials of all soils are sands with similar mineralogical composition: andesitic rock fragments, plagioclase, and pyroxene dominate, with minor amounts of opaque minerals. None of the parent materials contained >13% (v/v) volcanic glass. Weathering and neoformation of minerals with increasing soil age is characterized by (i) increasing pellicular and linear alteration of sand grains and (ii) decrease of the sand fraction and concomitant increase of finer material. Andesitic rock fragments weather more rapidly than plagioclase and pyroxene mineral grains. The alteration rates of the latter two are similar. Clay content in the ≈2000-yr-old soil is several times higher than in soils developed in rhyolitic parent materials of similar age in New Zealand. Formation of allophane with Al/Si ratios ranging from 1.9 to 3.8 takes place mainly in the B horizons. Aluminum-humus complexes, allophane, and Al oxides and hydroxides are mainly formed in the A horizons. Small amounts of gibbsite were noticed in soils older than 2000 yr. Small amounts of 2:1 and 1:1 clay minerals present in the clay fraction of all soils are thought to be inherited from the parent material, which contains sand-sized bodies of clay and andesitic rock fragments with clay pseudomorphs, both consisting of 2:1 and 1:1 clay minerals.

The effect of time on the amount and nature of the clay minerals formed in volcanic deposits has been studied by several authors. Investigations in New Zealand (summarized in Lowe, 1986), which at present has temperate humid conditions, showed that predominantly rhyolitic tephra deposits younger than ≈3000 yr have <50 g clay kg^{-1}, deposits that are 3000 to 10 000 yr old contain 50 to 100 g clay kg^{-1}, and those that are 10 000 to 50 000 yr old contain 150 to 300 g clay kg^{-1}. The amount of clay in these soils is thus correlated with age. Weathering of volcanic material under well-drained conditions in a humid climate usually leads to formation of Al- and Fe-humus complexes and short-range order material such as allophane, imogolite, ferrihydrite, noncrystalline Al and Fe oxides and hydroxides, and opaline silica (Wada, 1985; Mizota and Van Reeuwijk, 1989). The composition of allophane may reflect the weathering environment: Al-rich allophane (Al/Si = 2) is normally formed where the soil solution is relatively Si poor, while Si rich allophane (Al/Si = 1) is formed at higher Si concentrations (Parfitt, 1990). At relatively high Si concentrations, which occur in drier climates or under restricted leaching conditions, halloysite also may form directly from volcanic parent material (Parfitt and Wilson, 1985; Wada and Kakuto, 1985).

After further weathering, this material may be transformed to more stable minerals. Allophane and imogolite appear to be transformed to halloysite and kaolinite if conditions favor resilication, or to gibbsite if conditions favor desilication (Fieldes, 1955; Wada, 1989). Aluminum and Fe from the humus complexes will gradually become available for formation of oxyhydrates, and ferrihydrite will transform into goethite (Mizota and Van Reeuwijk, 1989). Layer silicates other than halloysite are often present in young soils derived from volcanic ash. Direct transformation of short-range order material into such layer silicates, however, has not been demonstrated (Wada, 1989).

Nieuwenhuyse et al. (1993) described soil formation in a Holocene (<5000 yr) soil chronosequence on sandy andesitic beach ridges in the Atlantic lowland of Costa Rica, focusing on the role of soil-forming factors on morphological, physical, and chemical soil properties. The objective of this study was to investigate mineralogical aspects of Andisol formation in the humid tropical climate of the study area, and to relate weathering and neoformation to soil age and drainage conditions.

MATERIALS AND METHODS

Sample Site Locations and Environmental Conditions

A beach ridge plain on the Atlantic coast of Costa Rica is composed of volcanic sediments from the Costa Rican Central Cordillera, where volcanism provides large amounts of material. Eight beach ridges, designated AT1 to AT8, in a 2.5-km transect stretching inland perpendicular to the coastline in Tortuguero National Park were selected for study (Fig. 1). All sites are on the flat tops of the beach ridges at <3 m above sea level. Fluvial influence is absent. Age estimations of the different soils were based on ^{14}C dating of peats east of the beach ridge plain and between the AT6 and AT7 sites, as well as on comparison of the studied soils with well-dated soils developed on fluvial sand deposits within 50 km of the study sites (Nieuwenhuyse, 1992, unpublished data). Drainage varies from excessively drained in the two youngest soils to poorly drained in the oldest soil. We think that drainage of the profiles slowly became poorer with increasing age due to the ongoing subsidence of the beach ridge plain. Soil classification, drainage, estimated ages, and texture of the soils are listed in Table 1. At all sites, vegetation is humid tropical rain forest and has been so since vegetation colonized the originally barren beach ridges. We estimate that the vegetation succession from the pioneer stage to evergreen forest takes about 50 to 100 yr after termination of active beach ridge formation. No vegetation changes are known to have occurred in the area during the past 4200 yr (Bartlett and Barghoorn, 1973). We did not detect

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Abbreviations: AAS, atomic absorption spectrometry; Al_o, acid oxalate extractable Al; Al_p, pyrophosphate-extractable Al; DCB, dithionite-citrate-bicarbonate; Fe_d, dithionite-extractable Fe; Fe_o, acid oxalate extractable Fe; Fe_p, pyrophosphate-extractable Fe; Si_o, acid oxalate extractable silica; TEM, transmission electron microscopy; TGA, thermogravimetric analysis; XRD, x-ray diffraction.

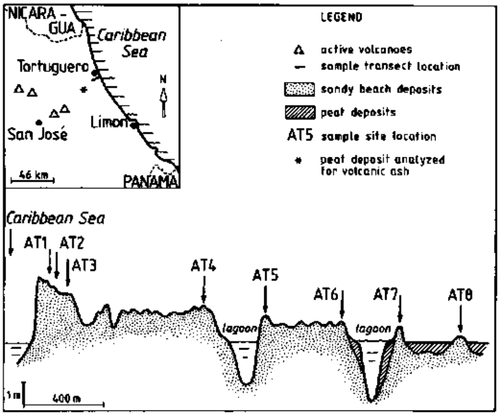

Fig. 1. Location of the study area and cross section of the sample transect.

volcanic ash (>10 µm) in thin sections of a peat deposit formed during the last 4450 yr at 20 km southwest of the study sites at 20 m above sea level (Fig. 1 inset). Since, moreover, prevailing wind directions are inland, addition of appreciable volcanic ash to the study area seems unlikely. However, addition of small amounts of volcanic dust cannot be excluded.

At the modern beach, the upper 80 to 90 cm of the sediments are mixed by nesting sea turtles (*Chelonia mydas*) and crabs, which probably accounts for the lack of stratification. The youngest soils have been homogenized by leaf-cutter ants and crabs, while earthworms and rodents burrow in the older soils. Mean annual rainfall is 5345 mm, with totals varying between 4100 and 6600 mm. Penmann potential evapotranspiration was estimated to be 1168 mm, using data from climatologic stations at the same altitude and within 40 km of the study site. Mean monthly precipitation exceeds mean potential evapotranspiration in every month of the year (Table 2). Dry periods up to 3 wk in duration occur each year, drying the younger soils and the A horizons of the older soils. Soil moisture measurements using a tensiometer at a depth of 20 cm in a loamy soil at a site 40 km west of the study area showed that in these dry periods maximum suction values are about 0.1 MPa. It is likely that in the sandier soils on the beach ridges these values are higher. The mean annual air temperature is 25 to 26 °C, with the highest temperatures in June (mean = 27.0 °C) and the lowest temperatures in January (mean = 24.6 °C).

Sampling and Analysis

At each site, a pit was dug to describe and sample the soil profile. Samples beneath the groundwater table were taken with an auger. After removing roots and litter, samples were sieved to obtain a homogeneous sample of <2-mm material. Part of each sample was stored at 4 °C at field moisture in sealed plastic bags within 1 wk after sampling; another part was air dried, mixed, and stored.

Acid oxalate extractable Al, Fe, and Si, and Al_p, and Fe_p, and Fe_d were determined in air-dry fine earth samples, as described by Van Reeuwijk (1987). Aluminum, Fe, and Si in the extracts were measured by AAS. All results have been recalculated on oven-dry (105 °C) basis. Soil pH was determined in 1:5 soil/water suspensions of field-moist samples, about 30 min after 5 min of intensive stirring (Van Lierop, 1990).

Undisturbed samples of 7 by 7 cm or 2 by 2 cm were taken for preparing thin sections and kept at field moisture. Disturbed samples were taken from the deepest horizons of the older soils. The samples were impregnated after replacement of water with acetone (Miedema et al., 1974). Thin sections were made following the method of FitzPatrick (1970), and described according to Bullock et al. (1985). The mineral composition of the coarse silt and sand fractions (>20 µm) was quantitatively estimated from thin sections by point counting at least 300 mineral grains (Van der Plas and Tobi, 1965).

Clay fractions of all major horizons of all profiles were separated by sedimentation after destruction of organic matter with H_2O_2 from field-moist samples using a mixture of 20 mL of 50 g kg^{-1} sodium polyphosphate and 20 mL of 100 g kg^{-1} NH_4Cl as dispersants. In sedimentation cylinders where flocculation of part of the samples was observed, the suspended material was separated for further study. Two samples of the flocculated material were studied with XRD, but results did not vary from those of the suspended material, and were not further taken into account. A second field-moist sample was sieved over a 53-µm mesh screen after adding dispersants, dried, and, under gentle rubbing, passed through a nest of sieves to determine weights of the following fractions: 2000 to 425, 425 to 250, 250 to 212, 212 to 150, and 150 to 53 µm.

Mineralogy of the clay fractions was determined by XRD, TGA, and TEM. Relative abundance of minerals was based on peak intensity. Various clay samples were treated with acid

Table 1. Soil classification, drainage, estimated age, and weighted average of clay and sand contents of the soil horizons.

Profile	Classification†	Drainage‡	Age	Texture§ A horizon clay	A horizon sand	B horizon clay	B horizon sand	C horizon clay	C horizon sand
			yr	g kg^{-1}					
AT1	Typic Tropopsamment	excessively drained	<100	30	930	na¶	na	0	1000
AT2	Typic Tropopsamment	excessively drained	<200	60	890	na	na	0	1000
AT3	Typic Tropopsamment	somewhat excessively drained	<500	55	870	na	na	0	1000
AT4	Typic Hapludand	moderately well drained	~2000	170	690	110	800	0	1000
AT5	Acrudoxic Hapludand	well drained	2000–5000	205	570	100#	810	0	1000
AT6	Acrudoxic Hapludand	well drained	2000–5000	205	620	80#	800	0	1000
AT7	Aquic Hapludand	imperfectly drained	2000–5000	270	500	110#	700	0	980
AT8	Hydric Haplaquand	imperfectly drained	2000–5000	210#	480	100#	650	0	990

† Soil Survey Staff, 1992.
‡ Food and Agriculture Organization of the United Nations, 1977.
§ Nieuwenhuyse et al., 1993.
¶ na = not applicable.
# These values should be considered minimum values since flocculation was observed in the sedimentation cylinders.

ammonium oxalate (Van Reeuwijk, 1987), to dissolve the poorly ordered part of the clay fraction so as to obtain better diffractograms of the crystalline components. For all clay separates, a preliminary scan was made of a moist sample in order to detect halloysite with d_{001} of 1 nm. Differentiation between kaolinite and halloysite species with d_{001} of 0.7 nm was made using the formamide test described by Churchman et al. (1984).

Allophane contents were calculated using the method proposed by Parfitt and Wilson (1985) and later simplified by Mizota and Van Reeuwijk (1989), as follows:

$$\text{allophane\%} = 100\mathrm{Si_o}/[23.4 - 5.1(\mathrm{Al_o} - \mathrm{Al_p})/\mathrm{Si_o}] \quad [1]$$

Ferrihydrite contents were estimated using Fe_o by (Childs, 1985):

$$\text{ferrihydrite\%} = 1.7\mathrm{Fe_o} \quad [2]$$

Microquantities of sand-sized clay bodies in uncovered thin sections were sampled with a microscope-mounted drill (Verschuren, 1978; Beauford et al., 1983). X-ray diffraction patterns were obtained by a step-scan method (Meunier and Velde, 1982).

RESULTS AND DISCUSSION

Parent Material

The presumably unaltered subsoil material from all soils consisted almost exclusively of sand between 53 and 425 μm in diameter, with particles between 425 and 2000 μm accounting for <20 g kg^{-1} and coarse silt (25–53 μm) making up <10 g kg^{-1} of the subsoil material. Coarser or finer material could not be detected or was not present.

The parent materials (C horizons) contained both single and compound mineral grains (Table 3). The single mineral grains were mainly plagioclase and pyroxene, commonly with inclusions of volcanic glass and Fe-bearing minerals. Traces of green and brown hornblende, biotite, olivine, and opaque ferrimagnetic minerals were present. The main compound mineral grains were holocrystalline to holohyaline, fine-grained andesitic rock fragments, with porphyritic and seriate textures. Dominant phenocrysts were plagioclase, pyroxene, and numerous small opaque grains. The groundmass regularly contained violet- to brown-colored volcanic glass. Holocrystalline fragments of quartzarenite and arkose had subhedral to anhedral granular textures. Mineral grains as well as rock fragments had smooth grain surfaces and were unfractured. Finally, variable amounts of greenish yellow, smoothly rounded, spherical to ellipsoidal anisotropic grains (50–400 μm) consisting of crystalline clay were found. In a number of the andesitic rock fragments, phenocrysts appeared to have been transformed into pseudomorphs consisting of such clay bodies. Step-scan x-ray diffractograms of small quantities obtained by microdrilling revealed the presence of 2:1 and 1:1 layer silicates (data not shown).

The ratios of andesitic rock fragments to pyroxene varied in the sediments of the different beach ridges. Relatively high concentrations of andesitic rock fragments were accompanied by lower pyroxene contents, and vice versa. Plagioclase concentrations were fairly constant throughout the chronosequence. In spite of these small differences, the parent material of the beach ridge soils can be considered fairly homogeneous.

Weathering of Primary Minerals

Table 3 gives the mineralogical composition of the >20-μm fractions. The amount of plagioclase, and to a lesser extent of pyroxene and andesitic rock fragments, varied little with depth in the profiles. In general, variation in mineralogy between profiles was larger than within the profiles. The proportion of andesitic rock fragments in the uppermost A horizons of the oldest soils (AT7 and AT8) is reduced relative to the lower horizons, suggesting that these fragments break up into the individual constituents, including plagioclase and pyroxene phenocrysts. Such dislodged phenocrysts would cause an increase in the volume fraction of such minerals, unless they underwent further weathering. The fact that contents of pyroxene and plagioclase increased only slightly with age suggests that these minerals weather almost as fast as they are released from the andesitic rock fragments. Most individual grains of all three dominant constituents showed increasing evidence of weathering with increasing soil age. Most grains in the A and B horizons showed pellicular alteration. Plagioclase and pyroxene grains showed irregular linear alteration in the A horizons of the oldest soils (Fig. 2). Point counts in which two or three weathering classes were distinguished (Table 4) revealed that the volume fraction of fresh mineral grains generally decreases with age along the chronosequence, particularly at shallow depth in the profiles, while the contribution of weathered grains increases in the same order.

According to Allen and Hajek (1989), volcanic glass is expected to weather more rapidly than augite, which in turn weathers more quickly than plagioclase. In this study, however, with the exception of profile AT7, ratios of plagioclase and pyroxene within the soil profiles tended to remain similar (Table 3), suggesting that they weather at roughly equal rates.

Detailed textural analyses of the AT4 to AT8 soils showed that sand is finer toward the soil surface. This is illustrated by data of the AT5 profile in Fig. 3. Taking

Table 2. Rainfall and Penman mean monthly potential evapotranspiration in Tortuguero.†

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Total
	mm												
Potential evapotranspiration	85	86	113	107	111	96	95	103	106	102	83	81	1168
Precipitation	455	302	235	286	453	403	591	499	326	460	668	666	5345
Surplus	370	216	122	179	342	307	496	396	220	358	585	585	4177

† Instituto Meterológico Nacional, Costa Rica, 1991 (unpublished data).

Table 3. Contents of main mineral components in the 20- to 550-μm fraction.

Horizon	Sample depth	Andesitic rock fragments	Pyroxene (Pyr)	Plagioclase (Plag)	Clay body	Volcanic glass†	Pyr/Plag ratio
	cm	% (v/v)					
			Profile AT1				
A/AC	2–8	50 ± 4	18 ± 3	23 ± 3	4 ± 2	8	0.8
C	15–20	53 ± 4	19 ± 3	22 ± 3	4 ± 2	9	0.8
C	75§	42 ± 4	35 ± 3	17 ± 3	3 ± 1	13	2.1
			Profile AT2				
A/AC	2–8	43 ± 5	19 ± 4	26 ± 4	6 ± 2	9	0.8
C1	15–21	48 ± 4	16 ± 3	29 ± 3	5 ± 2	10	0.5
C2	75§	48 ± 5	18 ± 4	26 ± 5	5 ± 2	11	0.7
			Profile AT3				
A	2–8	50 ± 5	19 ± 4	21 ± 4	3 ± 2	7	0.9
AC	13–19	53 ± 4	16 ± 3	24 ± 3	3 ± 1	6	0.7
C	30–36	45 ± 3	17 ± 3	29 ± 3	6 ± 2	10	0.6
C	75§	45 ± 5	18 ± 4	29 ± 5	4 ± 2	13	0.6
			Profile AT4				
A	2–8	50 ± 5	10 ± 3	28 ± 5	7 ± 3	7	0.4
AB	12–14	55 ± 6	5 ± 3	23 ± 5	12 ± 4	6	0.2
Bw	18–24	62 ± 5	7 ± 3	18 ± 4	5 ± 2	4	0.4
Bw	30–32	58 ± 7	6 ± 3	23 ± 6	9 ± 4	10	0.3
BCw	45–50	61 ± 5	7 ± 3	21 ± 4	8 ± 3	13	0.3
C	150§	56 ± 4	3 ± 1	24 ± 4	13 ± 3	9	0.1
			Profile AT5				
A	2–8	56 ± 5	7 ± 3	22 ± 4	8 ± 3	3	0.3
AB	11–17	57 ± 5	5 ± 2	23 ± 4	9 ± 3	6	0.2
BAw	23–29	52 ± 5	7 ± 3	24 ± 4	11 ± 3	7	0.3
Bw	35–41	61 ± 4	4 ± 2	21 ± 4	10 ± 3	4	0.2
BCw	60–66	65 ± 4	4 ± 2	21 ± 3	6 ± 2	6	0.2
CBw	82–88	57 ± 5	5 ± 2	25 ± 4	9 ± 3	11	0.2
			Profile AT6				
A	2–8	48 ± 6	11 ± 4	24 ± 5	9 ± 4	3	0.5
Bw1	20–26	44 ± 6	11 ± 4	20 ± 5	12 ± 4	3	0.6
Bw2	27–33	54 ± 5	12 ± 3	19 ± 4	10 ± 3	4	0.6
Bw3	40–46	56 ± 4	11 ± 3	19 ± 4	8 ± 2	4	0.6
BCw	60–66	60 ± 4	11 ± 3	16 ± 3	10 ± 3	8	0.7
BCwg	87–93	62 ± 4	10 ± 3	18 ± 4	7 ± 3	4	0.6
C	150§	60 ± 4	12 ± 2	18 ± 3	7 ± 2	3	0.7
			Profile AT7				
A1	2–8	30 ± 5	34 ± 5	26 ± 5	3 ± 2	2	1.4
A2	15–21	40 ± 5	32 ± 5	19 ± 5	2 ± 1	3	1.7
Bw	28–34	43 ± 5	27 ± 5	17 ± 4	7 ± 3	4	1.6
Bwg	41–47	39 ± 5	28 ± 5	24 ± 5	2 ± 1	1	1.1
BCwg	60–66	45 ± 5	20 ± 4	23 ± 4	6 ± 3	2	0.9
BCwg	81–87	42 ± 5	23 ± 4	24 ± 5	9 ± 3	3	0.9
			Profile AT8				
A1	2–8	27 ± 5	34 ± 5	29 ± 5	6 ± 3	6	1.2
BCg1	31–37	56 ± 5	21 ± 4	17 ± 4	4 ± 2	8	1.3

† Volcanic glass refers to the percentage of grains that contain clearly recognizable glass. Contents within the grains may vary considerably.
‡ Standard deviation according to Van der Plas and Tobi (1965).
§ Disturbed samples.

into account that some mass losses due to leaching may have occurred, more than one-half of the sandy parent material has been comminuted to sizes <53 μm in the A horizons of the AT7 and AT8 soils (Table 1). In these soils the amount of fine material continued to increase despite increasingly poor drainage. This suggests that comminution of sand by weathering continues at about the same rate under alternating oxidizing–reducing conditions as under aerobic conditions. In the AT8 soil, however, sand grains are less weathered below 30 cm than above (Table 4). If poor aeration and limited leaching caused by the permanently wet conditions at ≈30 cm and deeper have hampered weathering of the sand grains, then hydromorphic conditions may have prevailed at the site since shortly after deposition of this beach ridge.

Clay content in the 2000-yr-old soil AT4 is 170 g kg^{-1} in the A horizon and 110 g kg^{-1} in the B horizon (Table 1). Equally old volcanic soils in New Zealand contain <50 g clay kg^{-1} (Lowe, 1986), indicating a much higher rate of clay formation in Costa Rican soils. We think that higher leaching and higher temperatures are the most important factors causing these differences. In addition, differences in weatherability due to mineralogical composition (andesitic vs. mainly rhyolitic) between the parent materials and comminution of clay bodies present in the parent material may have contributed to the higher clay content observed in Costa Rica.

Formation of Noncrystalline Secondary Minerals

Textural and mineralogical data indicate that the parent material of the beach ridge soils contains only primary

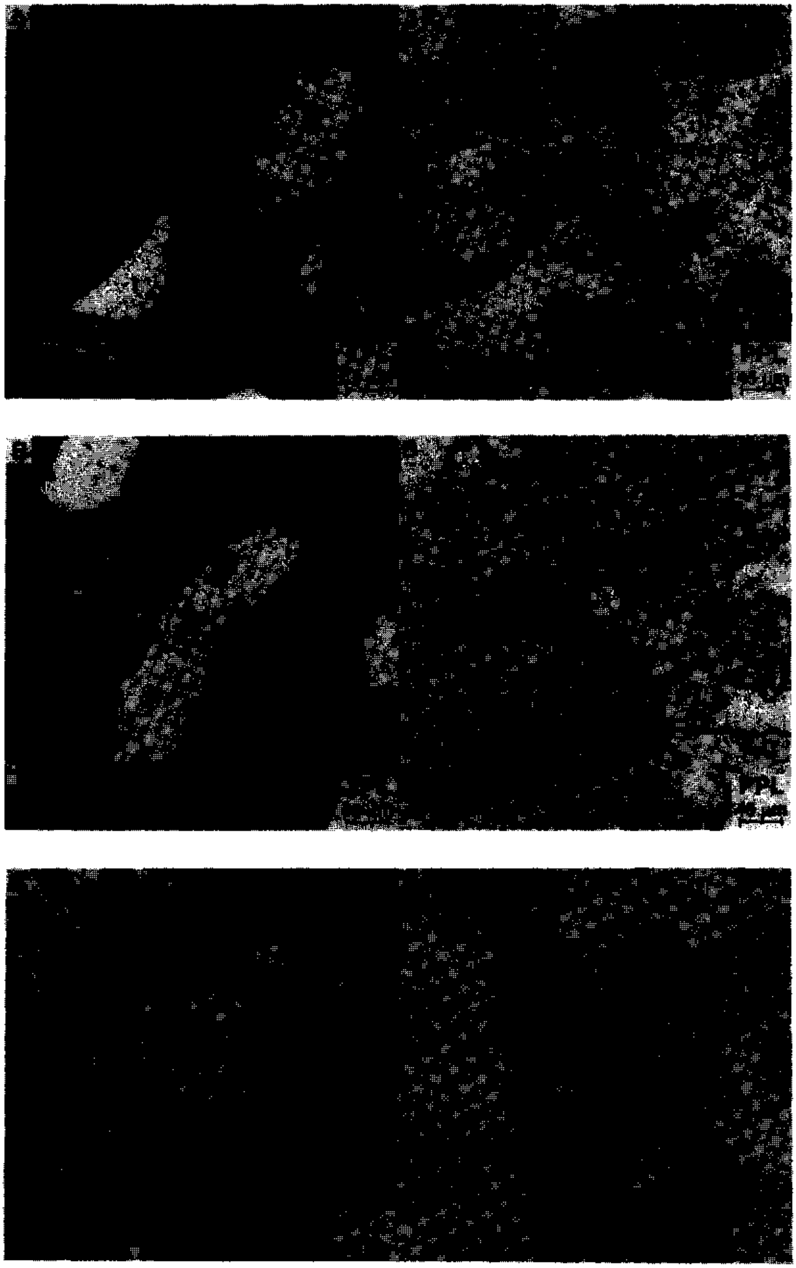

Fig. 2. Alteration features of the most important parent material components in the A horizon of profile AT7 in cross-polarized (XPL) and plane-polarized (PPL) light: (A) irregular linear alteration of plagioclase (1); (B) pellicular altered pyroxene (1); (C) thick pellicular and complex alteration of an andesitic rock fragment (1).

Table 4. Degree of weathering of the grains (20–550-μm fraction).

Horizon	Sample depth	Andesitic rock fragments F†	Andesitic rock fragments PW‡ + SW§	Pyroxene F	Pyroxene PW	Pyroxene SW	Plagioclase F	Plagioclase PW	Plagioclase SW
	cm	% (v/v)							
				Profile AT3					
A	2–8	23	77	42	56	2	29	65	6
AC	13–19	11	89	51	48	2	26	72	2
C	70–80	58	42	71	29	0	77	24	0
				Profile AT5					
A	2–8	4	96	13	54	33	18	59	24
AB	11–17	6	94	24	67	10	16	72	13
BAw	23–29	7	93	39	57	4	13	63	25
Bw	35–41	7	93	40	55	5	21	62	17
BCw	60–66	13	87	35	61	4	22	65	13
CB	82–88	34	66	38	52	10	34	45	21
				Profile AT7					
A1	2–8	1	99	1	62	37	4	41	56
A2	15–21	0	100	5	62	34	2	43	56
Bw	28–34	2	98	4	54	43	6	57	37
Bwg	41–47	4	96	5	59	36	4	55	42
BCwg	60–66	8	92	5	64	31	7	54	39
CBg	81–87	9	91	21	63	15	11	65	24
				Profile AT8					
A1	2–8	3	97	11	54	35	4	39	57
BCg1	31–37	18	82	33	55	12	10	66	25

† F = fresh, without signs of in situ weathering.
‡ PW = partly weathered, with pellicular alteration.
§ SW = strongly weathered, with thick pellicular and/or linear alteration.

minerals, in addition to the crystalline clay bodies. Assuming eolian contributions from inland sources have been negligible, any other secondary products present must therefore be ascribed to neoformation.

The fine material (<5 μm) observed in thin sections consisted of a mixture of organic matter and pale yellow isotropic inorganic material with an undifferentiated b-fabric, indicating dominant occurrence of noncrystalline material. Thermogravimetric analysis of selected clay separates invariably showed a gradual, continuous mass loss of 270 to 300 g kg^{-1} between room temperature

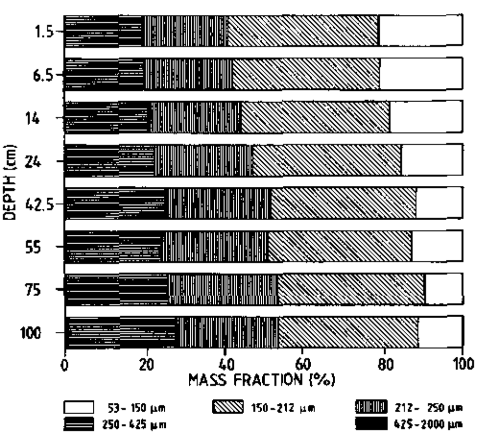

Fig. 3. Mass of various sand-size fractions of the AT5 profile. Depths are midpoints of soil horizons.

and 500 °C, characteristic for allophane (Wada, 1989). X-ray diffraction patterns of clay separates showed low peaks that became more intense after acid ammonium oxalate treatment (Fig. 4a and 4b), indicating that x-ray amorphous material had been removed. Transmission electron microscope images of clay separates of both young and old soils showed dominant occurrence of amorphous material. All these observations indicate that short-range order material and Al– and Fe–humus complexes make up a major part of the clay fraction.

Contents of Al_o, Si_o, Al_p, Fe_p, and Fe_d gradually increased with soil age (Table 5), indicating increasing amounts of Al– and Fe–humus complexes and short-range order material. Within the profiles, Al_o and Si_o were highest in the B horizons, while Al_p and Fe_p were highest in the A horizons, and Fe_d was generally higher in the A and B horizons than in the C horizons. The Fe_o values showed no clear trend with depth. The Fe_o, Fe_p, and Fe_d contents were considerably higher in the AT7 and AT8 soils than in all other profiles. The high variability in Fe_o/Fe_d ratios and the occurrence of ratios >1 in all profiles may be explained by dissolution of some Fe from magnetite by acid ammonium oxalate, while dithionite extraction does not affect magnetite (Borggaard, 1988).

With the exception of the B horizon of profile AT8,

Fig. 4. X-ray diffraction patterns of air-dried clay separate of the A horizon of the AT7 profile: (A) untreated subsample, Mg saturated; (B) acid ammonium oxalate treated subsample, Mg saturated; (C) glycerol-treated subsample, Mg saturated; (D) K-saturated subsample, after heating to 250 °C.

ratios of (Al_o - Al_p/Si_o) were close to 2, which is about the most common ratio of allophane (Parfitt and Kimble, 1989). The high ratios in the AT8 soil indicate low Si concentrations in soil solution relative to Al.

Calculated allophane contents gradually increased with soil age from zero in the youngest soil to >100 g kg^{-1} in the B horizon of the oldest soil. The Al_p/Al_o ratios clearly decreased with depth (Table 5), which suggests

Table 5. Chemical properties of profiles, calculated allophane (All.) contents, and ratios of selectively dissolved components.†

Horizon	Depth	pH	Al_o	Al_p	Fe_o	Fe_p	Fe_d	Si_o	All.	Al_p/Al_o	Fe_o/Fe_d	$(Al_o - Al_p)/Si_o$
	cm		g kg^{-1}									
						Profile AT1						
A	0–4	5.6	1	1	5	0	7	0	0	1.0	0.7	nd‡
CA	4–10	5.7	2	1	5	0	6	0	0	0.5	0.8	nd
C	20–25	6.4	1	0	6	0	5	0	0	0.0	1.2	nd
C	50–55	6.7	1	0	8	0	4	0	0	0.0	2.0	nd
						Profile AT2						
A	0–2	5.7	3	2	3	1	6	0	0	0.7	0.5	nd
AC	4–8	5.8	3	2	3	1	6	0	0	0.7	0.5	nd
CA	25–30	6.5	2	0	3	0	4	1	8	0.0	0.8	2.0
C2	55–60	6.6	1	0	5	0	4	0	0	0.0	1.3	nd
						Profile AT3						
A1	0–3	5.7	6	4	4	2	7	1	8	0.7	0.6	2.0
A2	4–8	5.9	7	4	4	2	7	1	12	0.6	0.6	3.0
AC	10–15	5.8	7	4	5	1	6	2	13	0.6	0.8	1.5
CA	17–23	6.1	5	2	5	1	5	2	13	0.4	1.0	1.5
C	28–34	6.3	4	1	6	0	4	1	12	0.3	1.5	3.0
C	50–55	6.4	2	0	4	0	3	1	8	0.0	1.3	2.0
						Profile AT4						
A	0–3	6.0	14	6	4	2	9	4	30	0.4	0.4	2.0
A	6–10	6.0	16	8	5	3	10	4	30	0.5	0.5	2.0
AB	13–17	5.9	17	7	5	3	10	5	38	0.4	0.5	2.0
Bw	25–30	6.2	21	6	6	2	13	7	56	0.3	0.5	2.1
BCw	45–50	6.1	15	2	2	0	6	6	49	0.1	0.3	2.2
CB	50–70	6.4	13	1	2	0	6	6	45	0.1	0.3	2.0
C	100–110	6.5	5	1	3	0	3	3	18	0.2	1.0	1.3
						Profile AT5						
A1	0–3	5.4	24	11	6	4	12	6	49	0.5	0.5	2.2
A2	5–8	5.5	23	10	6	4	11	6	49	0.4	0.6	2.2
AB	12–16	5.6	23	9	6	4	12	7	53	0.4	0.5	2.0
BAw	22–26	6.3	24	5	5	2	11	8	71	0.2	0.5	2.4
Bw	40–45	6.7	18	2	2	0	7	7	60	0.1	0.4	2.3
BCw	70–80	7.4	9	1	1	0	4	5	33	0.1	0.3	1.6
C	135–145	nd	4	1	4	0	3	2	13	0.3	1.3	1.5
C	190–200	nd	2	0	6	0	7	1	8	0.0	0.9	2.0
						Profile AT6						
A	0–3	5.5	21	8	6	4	14	7	50	0.4	0.4	1.9
A	6–12	5.3	24	11	8	7	15	6	49	0.5	0.5	2.2
Bw1	18–24	5.9	28	6	7	2	13	10	82	0.2	0.5	2.2
Bw2	30–35	6.5	24	3	6	0	12	10	79	0.1	0.5	2.1
Bw3	41–46	6.5	20	2	4	0	10	8	67	0.1	0.4	2.3
BCw	60–65	6.7	20	2	4	0	8	9	68	0.1	0.5	2.0
BCwg	85–95	6.6	16	2	3	0	9	7	53	0.1	0.3	2.0
C	110–120	6.4	8	1	4	0	5	4	28	0.1	0.8	1.8
C	190–200	nd	2	0	7	0	4	1	8	0.0	1.8	2.0
						Profile AT7						
O	2–0	4.6	6	7	10	10	18	0	0	1.1	0.6	nd
A1	2–8	4.9	21	17	20	18	26	2	15	0.8	0.8	2.0
A2	14–20	5.5	30	12	12	9	21	9	68	0.4	0.6	2.0
Bw	28–33	6.1	31	6	9	3	15	11	93	0.2	0.6	2.3
Bwg	38–48	6.3	27	3	7	0	16	12	91	0.1	0.4	2.0
BCwg	60–70	6.0	26	3	7	1	21	11	86	0.1	0.3	2.1
CBg	90–100	6.7	22	2	6	0	8	10	76	0.1	0.8	2.0
C	150–160	6.1	14	1	5	0	7	7	50	0.1	0.7	1.9
C	210–220	nd	4	1	8	0	3	2	13	0.3	2.7	1.5
						Profile AT8						
O	5–0	4.5	5	4	8	6	13	0	0	0.8	0.6	nd
A	1–12	5.1	21	10	21	15	27	6	43	0.5	0.8	1.8
Bwg	15–25	6.0	32	8	12	4	18	7	118	0.3	0.7	3.4
BCg1	30–45	6.3	24	4	7	1	24	6	94	0.2	0.3	3.3
BCg2	60–75	6.7	21	2	8	0	20	5	125	0.1	0.4	3.8
CBr	90–100	6.6	16	1	5	0	9	6	57	0.1	0.6	2.5

† Al_o = oxalate-extractable Al; Al_p = pyrophosphate-extractable Al; Fe_o = oxalate-extractable Fe; Fe_p = pyrophosphate-extractable Fe; Fe_d = dithionite-extractable Fe; Si_o = oxalate-extractable Si.
‡ nd = not determined.

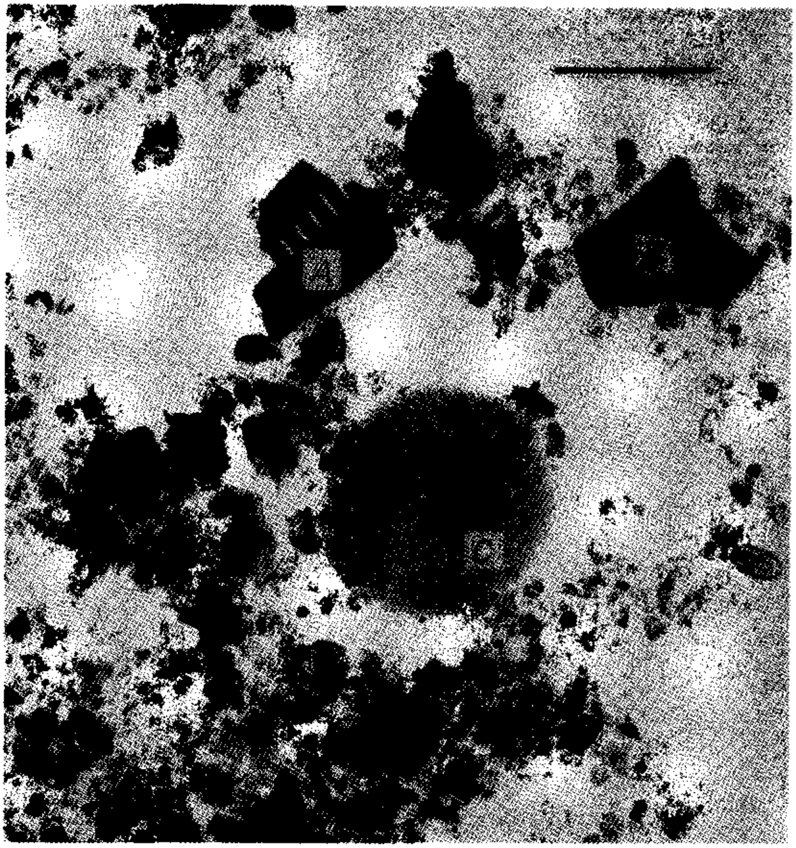

Fig. 5. Transmission electron micrograph of clay separate of A horizon of AT2 soil, showing: diatom fragment (A), crystalline clay particle (B), and diffuse amorphous floc interpreted as opaline Si (C). The other material shown is interpreted as short-range order material. Reference line represents 2 μm. Micrograph taken at 80 kV.

that formation of complexes of Al with organic matter mainly occurs in the A horizons. These complexes apparently coexist with allophane in the A horizons, but allophane dominates in the B and C horizons. Soluble silica and humic acids may compete for dissolved Al in the A horizon, and the relative availability of these acids determines whether formation of allophane or of Al-humus complexes dominates (Mizota and Van Reeuwijk, 1989). Calculations (using Eq. [1] and [2]) illustrate the dominance of Al- and Fe-humus complexes and short-range order material in the clay fraction and their distribution with depth. In the B horizon of the AT7 profile, average allophane content was calculated to be ≈ 90 g kg^{-1}, while ferrihydrite content was estimated to be ≈ 15 g kg^{-1}. At a clay content of at least 110 g kg^{-1} (Table 1), allophane appears to be dominant in the clay fraction of this horizon. In the A horizons of AT7, however, allophane contents were 15 and 68 g kg^{-1}, while ferrihydrite accounted for some 20 to 34 g kg^{-1}. Few layer silicates or opaline Si particles were observed in TEM images of this horizon. Given a clay content of about 270 g kg^{-1}, the major part of the clay fraction must be composed of Al-humus complexes or (amorphous) Al oxides and hydroxides.

The AT1 to AT6 soils are slightly acid (pH 5.5–6) in the A horizons and are near neutral in lower horizons. The A horizons of the two oldest profiles are acid, with pH values close to 5 (Table 5). A soil pH of at least 4.8 is required for allophane to precipitate (Shoji et al., 1982; Parfitt and Kimble, 1989). In the A horizons of the AT7 and AT8 profiles, pH values are apparently too low for the formation of allophane, and any allophane present might be transformed into other minerals.

In the B horizons of the AT5 and AT6 profiles and in the A horizons of the AT7 and AT8 profiles, low 0.485 peaks indicative for gibbsite were present (data not shown). Detectable amounts of gibbsite occurred only in soils older than 2000 yr. The absence of gibbsite in the parent material and its presence in the clay fraction of the older profiles suggests that short-range order materials are being transformed into gibbsite. However, the formation of gibbsite directly from a soil solution poor in Si cannot be excluded, or both processes may occur (Hsu, 1989).

Optical observations of thin sections showed that opal phytoliths (5–20 μm) are common in the surface horizons of the younger profiles and abundant in the surface horizons of the older profiles. Transmission electron microscope images revealed the presence of diatom fragments and particles resembling opaline silica (Fig. 5). After weathering, both Al and Si are released from primary minerals. In the A horizons probably most of the Al forms complexes with organic matter, causing relatively high soluble Si levels. Part of this Si may be absorbed by plants and converted to phytoliths, or may appear as diffuse opal flocks (Drees et al., 1989).

Presence of Layer Silicates

All clay separates produced small 0.7- and 1.4-nm peaks, which became more pronounced after acid ammonium oxalate treatment (Fig. 4a and 4b). Peak intensity varied little throughout the chronosequence. After glycerol treatment of a Mg-saturated subsample, the 1.4-nm peak partly shifted to higher d-spacing (Fig. 4c), indicating the presence of smectite. Following K saturation and heating to 250 °C of a different subsample, the 1.4-nm peak moved to a d-spacing between 1.4 and 1 nm (Fig. 4d), indicating random interstratifications of chlorite and (hydroxy-interlayered) vermiculite.

Field-moist samples showed no halloysite with a d_{001} spacing of 1.0 nm. After formamide treatment of the dried sample, the 0.7-nm peak did not expand to 1.0 nm, so halloysite appears to be absent. The 0.7-nm peak must probably be attributed to kaolinite, since no third and fourth order peak of chlorite at 0.475 and 0.354 nm were detected, and the 1.4-nm peak almost disappeared at 250 °C (Fig. 4d). Occasionally crystalline clay particles were observed in TEM images (Fig. 5). Under restricted drainage conditions, halloysite is often formed from volcanic parent materials (Parfitt and Wilson, 1985). The fact that no halloysite was found in the imperfectly drained AT7 and AT8 soils indicates that Si concentrations in the soil solution are too low for halloysite formation, in spite of the imperfect drainage conditions.

Given the presence of identical clay minerals in the parent material, we conclude that no in-situ formation of silicate clays takes place and that layer silicates found in the clay separates originated from the clay bodies and clay pseudomorphs in the andesitic rock fragments. The crystalline clay probably released by the breaking up of these components during soil formation. However, some of the clay may have been mechanically separated from the clay aggregates during clay separation in the laboratory.

Similar explanations for the presence of 2:1 layer silicates in young profiles developed on volcanic parent material were given by Dudas and Harward (1975) and Pevear et al. (1982), who found 2:1 layer silicates inherited from the parent material. In Costa Rica, Dondoli (1965) observed clay particles in recently erupted ash from the Poas volcano that were probably formed hydrothermally within the volcano before eruption. Subsequent erosion, transportation, and deposition can be responsible for incorporation of these layer silicates in the beach ridge sediments.

CONCLUSIONS

1. The humid tropical climate of the study area leads to a rapid weathering of andesitic sand, evidenced by decreasing sand content and increasing content of fine material with increasing soil age. Clay contents as high as 170 g kg^{-1} in the A horizon of a 2000-yr-old soil were found.
2. Under humid tropical conditions, weathering of andesitic sand composed of mainly plagioclase, pyroxene, and rock fragments containing volcanic glass does not show clear preferential weathering of one of the components within 2000 yr. After further weathering, rock fragments seem to weather at a faster rate than other components. Waterlogging in deeper horizons appears to slow down the weathering rate.
3. Independent of drainage conditions, the main neoformed weathering products are Al–humus complexes, (amorphous) Al oxides and hydroxides, and allophane in the A horizons and allophane in the B horizons.
4. Detectable amounts of gibbsite were present in soils older than 2000 yr. It was not clear whether it was formed by transformation of short-range order material, or by precipitation from soil solution.
5. The 2:1 and 1:1 layer silicates found in clay separates of all soils are thought to be derived from disintegration of sand-sized clay bodies consisting of 2:1 and 1:1 clay minerals present in the parent material.

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Inheritance of 2:1 Phyllosilicates in Costa Rican Andisols

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ABSTRACT

The occurrence of 2:1 phyllosilicates in Andisols is variously ascribed to in situ pedogenic origin, aeolian addition, or the presence of hydrothermally altered rock fragments. We studied the origin of 2:1 phyllosilicates that occur in Holocene Hapludands on andesitic, sandy beach ridges in Costa Rica by micromorphological, mineralogical, and submicroscopical techniques. The 2:1 phyllosilicates also occur as pseudomorphs after primary minerals in fresh rock of the inland volcanoes, from which the parent material of the beach ridges was mainly derived. Hydrothermal processes are most likely responsible for the formation of such pseudomorphs. Rock weathering produces sand-sized rock fragments with clay pseudomorphs and also liberates individual pseudomorphs. Subsequent erosion and alluvial transport affect their shape, but not their internal fabric. In the beach ridges, clay pseudomorphs appear as individual, sand-sized clay bodies, and inside sand-sized andesitic rock fragments. Submicroscopical analyses of these individual clay bodies and andesitic rock fragments with clay pseudomorphs indicate a predominance of 2:1 phyllosilicates. This implies that they are inherited from the parent material and are not due to postdepositional soil formation in the beach ridges. Weathering and biological activity affect the clay bodies and rock fragments with clay pseudomorphs, leading to the formation of clay-sized particles consisting of 2:1 phyllosilicates. Toward the soil surface, these particles are incorporated into the allophanic groundmass resulting from actual soil formation. The geographically extensive occurrence of 2:1 phyllosilicates in Andisols suggests that the genetic processes described here may have more than regional validity.

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ALLOPHANE and 1:1 phyllosilicates are the most common secondary minerals in soils formed in pyroclastic deposits in humid tropical areas without a distinct dry season (Mizota and van Reewijk, 1989; Parfitt and Kimble, 1989; Quantin et al., 1990). The occurrence of 2:1 phyllosilicates has often been reported in such soils

Abbreviations: EDS, electron diffraction spectometer; EG, ethylene glycol; SSXRD, step-scan x-ray diffraction; TEM, transmission electron microscope; XRD, x-ray diffraction.

CHAPTER 7: QUANTITATIVE ASPECTS OF WEATHERING AND NEOFORMATION IN VOLCANIC SOILS IN PERHUMID TROPICAL COSTA RICA.

A. Nieuwenhuyse and N. van Breemen

ABSTRACT

Assuming Ti to be immobile, gains and losses of major elements in were calculated from total element contents. Initially soil formation (0 to 0.5 ka) in Tropopsamments involves dilation of the sandy deposits by incorporation of organic matter and formation of structure and biopores, without detectable gains or losses of elements. In 2 to 5 ka old sandy Hapludands, primary minerals are still abundant, but up to 20% of the mineral soil consists of X-ray amorphous materials. Dilation continues and losses of Mg, Ca, Na and K, and to a lesser degree Si, have become measurable. In a <18 ka old, stony Melanudand primary minerals and, especially, volcanic glass, become depleted in the fine-earth fraction. A mixture of short-range order material, metal-humus complexes, gibbsite and kaolin minerals dominate in the A horizon, while gibbsite, halloysite, and short- range order material are the most important secondary minerals at greater depths. Dilation of the soil mass prevails in the A horizon, while collapse has occurred in the B horizon. Considerable amounts of mobile elements are lost: 50 to 85% of Si, Mg, Ca, Na, and K have been leached from the soil profile. The strongly collapsed Haploperox, >50 to 450 ka old, thought to have formed from Andisols, are devoid of primary minerals (except opaques) in the upper meters, and are dominated by gibbsite, kaolin minerals and iron (hydr)oxides. Under the prevailing environmental conditions, weathering and neoformation of primary volcanic minerals lead to almost complete losses of basic cations, probably in a time period between 20 and 50 ka. Si and P mineral reserves are depleted considerably, but part is still present after such long time periods.

INTRODUCTION

Knowledge about rates of mineral weathering and neoformation in soils is important, to estimating (1) acceptable annual soil loss by erosion (Wischmeier and Smith, 1978), (2) nutrient supply by weathering (Stoorvogel, 1993), (3) the rate of buffering of soils by acid rain (Van Grinsven, 1988) and (4) the sink strength of soils for atmospheric CO_2 (Chadwick et al., 1994). Mass balances can be derived for soils developed in homogenous parent material containing an immobile element (Haseman and Marshal, 1945; Brimhall and Dietrich, 1987), or for isovolumetrically weathered rock (Millot and Bonifas, 1955; Veldkamp et al., 1990). Next, time-averaged weathering rates can be calculated if the age of the parent material is known. Present-day volume-averaged weathering rates of soil and underlying rocks can be evaluated from element input-output budgets of watersheds (Bruijnzeel, 1990).

Although weathering pathways of volcanic materials are well known (Wada, 1989; Shoji et al., 1993), rates of weathering in such materials have received little attention, especially in humid tropical climates. The scarcity of estimates of weathering rates of volcanic deposits is remarkable, because they can often be dated precisely. However, rejuvenation by small ash additions often make volcanic deposits unsuitable for profile balance studies. Moreover, watertight watersheds are rare in volcanic areas, precluding estimates of present-day weathering rates from input-output budget of solutes.

In this paper we utilize the availability of dated volcanic parent materials of soils with little or no ash rejuvenation to estimate such weathering rates in the humid tropical Atlantic lowland of Costa Rica, and relate the results to soil mineralogy and morphology.

MASS BALANCE CALCULATIONS

A set of equations to estimate open-system mass transport (Brimhall and Dietrich, 1987; Brimhall et al., 1991) permits calculation of chemical gains and losses of elements for a soil material sample compared to the parent material. First, volumetric changes are estimated from the strain $\epsilon_{i,w}$, i.e., the ratio of volume change during weathering and soil formation, to the initial volume $((V_w-V_p)/V_p)$. Positive strains are dilations of the soil mass and negative strains represent collapse. Assuming that an immobile component (element) is present in the soil, strain can be calculated as follows:

$$\epsilon_{i,w} = (\rho_p C_{i,p}/\rho_w C_{i,w}) - 1 \qquad \text{(equation 1)}$$

in which:

ρ = dry bulk density in g cm^{-3} for soil sample (ρ_w) and parent material (ρ_p)

C_i = mass concentration in % of element i in the soil sample ($C_{i,w}$) and parent material ($C_{i,p}$)

The loss or gain of element j from the soil, $\tau_{j,w}$, is calculated as follows:

$$\tau_{j,w} = (\rho_w C_{j,w}/\rho_p C_{j,p})(\epsilon_{i,w} + 1) - 1 \qquad \text{(equation 2)}$$

If $\tau_{j,w}$ is calculated for each horizon, the net mass flux of each analyzed element (m_j,flux) can be calculated for the entire studied soil with depth $Z = D_{j,w}$:

$$m_{j,flux}\ (g\ cm^{-2}) = \rho_p (C_{j,p}/100) \int_{Z=0}^{Z=D_{j,w}} \tau_{j,w(z)} dZ \qquad \text{(equation 3)}$$

The equations are applicable only if: (1) the soil has been formed in a homogeneous parent material of the same age, and (2) the soil material contains an immobile element (Chadwick et al., 1990). To calculate rates of soil formation and mineral weathering, also the soil age should be known.

These conditions are rarely met. Many parent materials, in particular fluvial or pyroclastic deposits are stratified, and chemically and structurally heterogeneous. Furthermore, additions of material of a similar composition may occur after soil formation has started, e.g. volcanic ash deposition or sedimentation of thin alluvial deposits. Often these cannot be recognized in the soil profile, even shortly after their deposition.

Generally, elements like Cr, Ti, V, and Zr are thought to be immobile in soils since minerals in which they occur are often stable in soil environments (Haseman and Marshall, 1945; Milnes and FitzPatrick, 1989). However, this may not always be so. For example, an Oxisol in South West Wales, Australia, had lost considerable Ti by leaching (Mohr et al., 1972), presumably in the form of organic Ti-complexes under acid and reducing conditions. Ti was also leached from a Paleustalf formed in granodiorite (Kaup and Carter, 1987), and Ti losses were reported from weathering rinds in alkali basalt pebbles in France (Jongmans et al., 1993).

Dating soils is difficult, and provide in many cases only approximations of the true age of a soil profile. Sometimes soils can be dated by ^{14}C datings of soil organic matter (Matthews and Dresser, 1983) or buried A horizons. However, soil age estimations usually are based on the position of the soil in the landscape in relation to known geological events, e.g. uplift rates of terraces, or by dating parent material or under- or overlying deposits.

SOILS OF THE ATLANTIC LOWLAND OF COSTA RICA

The study area forms part of the Nicaragua depression, a back-arc basin whose Costa Rican part is filled up mainly with fluviovolcanic sediments derived from the volcanic Central cordillera (Weyl, 1980). In the footslopes of this cordillera soils of various age are found, formed on alluvial fans, lava and mud flows. In the extensive lowlands fluvial deposits of two age classes occur: <5 ka old Holocene soils of variable texture, and >50 ka old Pleistocene clayey soils on slightly elevated (about 10 to 25 m) terrace remains (Fig.1).

Fluvial deposits in which the older soils have been formed are thought to have been deposited during a former high sea-level, possibly during the Eemian/Sangamonian interglacial (± 125 ka ago). Along the Caribbean Sea a <5 ka old sandy beach ridge plain occurs. The texture of soils on lava is different from the alluvial soils. The flows contain massive boulders relatively resistant against weathering, while also porous, more easily weatherable coarse fragments occur, as well as considerable amounts of sand and silt sized fine material.

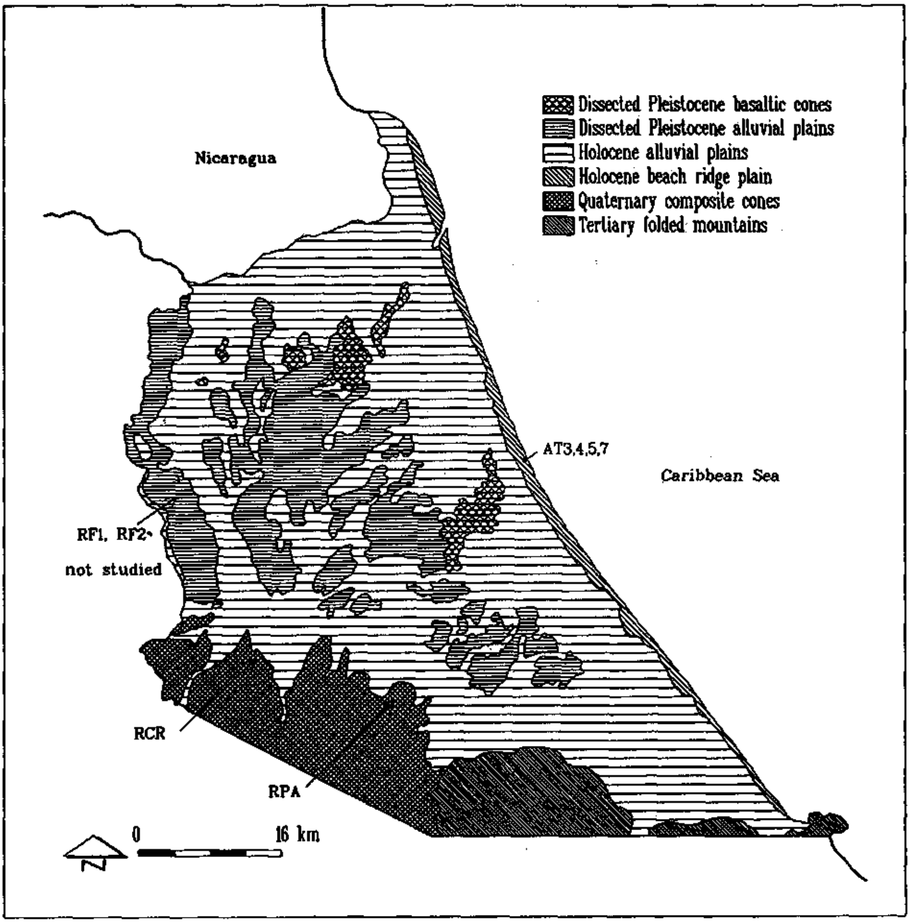


Figure 1: ***Location of the studied soils and mayor landforms in the study area.***

Main annual air temperature is 25 to 26°C, with differences between the coolest (January) and hottest (June) month of about 2°C. The mean annual rainfall of about 3500 to 5500 mm is well distributed throughout the year and exceeds mean evapotranspiration in all months (Instituto Meteorológico Nacional de Costa Rica, 1992, unpublished data). All sites were deforested less than 50 years ago or are still under forest.

Throughout this area, 8 soils were selected for which approximate ages are known (Table 1). These soils meet the requirements mentioned earlier: they have been formed in thick, more or less homogeneous parent materials, and as far as we could deduce from observations in nearby peat deposits the AT3,4,5, and 7, they received no detectable additions of volcanic ash or sediments during the past 5000 years (Nieuwenhuyse et al., 1993, 1994). Micromorphological observations suggest that the RCR soil and, possibly, both RF soils, have been slightly enriched with ash, although no signs of it could be detected in the field. No evidence for ash enrichment was found in the RPA soil. Finally, they contain easily measurable amounts of the slowly mobile elements Cr, Ti, V, and Zr.

Table 1: Selected characteristics of the studied soil profiles.

Profile	Age[1] -ka-	Landform[1], Parent material[2]	Classification[3]
AT3	<0.5 (0.2)	sandy beach ridge, basaltic andesite	Typic Tropopsammment
AT4	2	sandy beach ridge, andesite	Typic Hapludand
AT5	2-4 (2)	sandy beach ridge, andesite	Acrudoxic Hapludand
AT7	2-4 (3.5)	sandy beach ridge, basaltic andesite	Aquic Hapludand
RPA	<18 (18)	blocky lava, basaltic andesite	Pachic Melanudand[4]
RF1,2	>50 (125)	fluvial deposits, mainly andesitic sand	Typic Haploperox
RCR	<450 (450)	lava, (trachy)andesite	Typic Haploperox

[1] Data taken from previous studies (Nieuwenhuyse et al., 1993, 1994) and unpublished material (Nieuwenhuyse, 1993). The most likely age is given between parentheses.

[2] According to MacKenzie and Guilford (1986).

[3] Soil Survey Staff, 1992.

[4] Tentative, since no melanic index was determined.

Except for the RF soils which contain clayey layers in the subsoil, the soils have a similar parent material mineralogy: plagioclase, pyroxene and opaque minerals in a matrix with variable amounts of volcanic glass. Clayey layers in the RF soils probably were originally composed of kaolinite/halloysite, 2:1 layer silicates, and short-range order materials, just as present-day clayey river sediments (Van Seeters, 1993). Any sedimentary layering in the topsoil has been erased by faunal homogenization.

Although present-day climate is similar at all sites, the older-than-Holocene sites are likely to have experienced climate changes. As indicated by palynological data (Bartlett and Barghoorn, 1973), a more seasonal rainfall distribution may have prevailed during glacial periods.

Except for the youngest AT3 soil and the three oldest soils all soils are Andisols (Table 1). We assume that also the older soils initially were Andisols, because (1) the parent material is like that of the younger soils; (2) even during glacial maxima climate differed little from the actual (Bush and Colinvaux, 1990) (3) bulk density is still low (about 0.9) and (4) in the subsoil of the RF soils short-range order material can still be found.

METHODS

Soils were described in profile pits or fresh exposures in quarries and road cuts (Nieuwenhuyse et al., 1993, 1994). Care was taken to select sites in flat positions in the landscape where risk of lateral migration of particles was minimal. Horizons were described according the FAO (1990). Bulk density of each horizon was determined by measuring the oven-dry mass of 100 ml core samples taken in triplicate. Petrographical composition of material >20 μm was determined qualitatively by studying thin sections, using standard optical techniques. X-ray diffraction (XRD) on <20 μm fractions, and thermogravimetric techniques on whole soil samples were used to obtain further information on the soil's mineralogical composition.

For this study, chemical composition of the <2 mm material was determined by X-ray fluorescence (XRF). After ignition at 900°C, glass disks were obtained by melting aliquots of the samples with $Li_2B_4O_7$, and analyzed on a Philips XRF assembly. Major element contents were presented as mass fractions of oxide components and recalculated to a volatile-free basis.

Strain, mass fractions added to or subtracted from each horizon, and loss or gain of elements during pedogenesis were calculated using equations 1, 2, and 3. Of the potentially immobile elements Ti, Zr, Cr, and V (Wedepohl, 1967-1974), Ti had the highest concentrations and was used as index element. However, in order to check immobility of Ti against the other supposedly immobile elements, for some profiles strain was also calculated using Cr, V, and Zr as index elements.

No fresh parent material was exposed in the RF profiles. Chemical composition and bulk densities of their parent materials were assumed to be similar to those of actual sediments from rivers draining the Central cordillera (Van Seeters, 1993), because (1) volcanism in the volcanic source areas of the sediments has been more or less the same during the past two millions years (Weyl, 1980), and (2) fresh sandy sediments in the RF2 saprolite protected against weathering by sesquioxide/silica coatings had a similar mineralogical composition as fresh river sediments. Calculations for these layered profiles were done taking into account chemical differences between parent material of different sedimentary layers: when sedimentological structures indicated fine sediment (e.g. fine horizontal lamination), parent material was assumed to be silty clay sediment, while for layers with structures typical for higher stream velocities (e.g. cross bedding) a medium sand composition was used. Tentatively, a medium sand composition was also used for the homogenized topsoils in which no sedimentary structures were observed.

Quantifying element gains and losses in a stony soil as the RPA soil is difficult. Firstly, a meaningful value for bulk density is hard to obtain (Chadwick et al., 1990). Secondly, fresh stone cores may occur adjacent to strongly weathered soil matrix. Based on observations in the exposures, we estimated that 40 to 60% of this soil is composed of unaltered, massive andesite. To compensate for variability, all samples of this soil were taken in triplicate. The

RCR soil, also developed on lava, was completely weathered over the studied depth.

To test whether the estimated losses during pedogenesis resemble present-day leaching losses from watersheds, some water samples from the Tortuguero river were taken and analyzed for major elements. This river drains soils similar to those studied and does not receive water from the mountain slopes. Furthermore, soil solutions of a soil similar to the RF soils and of a young sandy soil similar to the AT5 soil, as well as rain water composition were sampled and analysed about weekly for several months under both relatively dry and wet weather conditions.

RESULTS AND DISCUSSION

The use of Ti as immobile element

Ti is residually enriched in all profiles. The mobility of Ti was compared for four profiles with that of Cr, V, and Zr through calculation of strain values (Fig.2). Ti, Cr and V give comparable strain values, but ϵ_{Zr} values are higher, indicating that Zr is more mobile than the other elements. Zr mobility is difficult to explain, but may be related to presence of Zr as a contaminant in pyroxene (Wedepohl, 1969-1974), which may have been leached upon release by weathering. We conclude that under the conditions of the study area Ti appears to be best index element, partly because of its relatively high concentration.

<0.5 ka old Tropopsamment (AT3).

Except for a thin A horizon in which organic matter has accumulated, the soil profile is not yet differentiated (Table 2). Positive strain values (Fig.2) are due to incorporation of organic matter (which has a lower bulk density and lower concentration of immobile elements than mineral soil material, thus contributing to dilation) and formation of (bio)pores, which are the major soil forming processes in such young soils. Small amounts of X-ray amorphous fine material have been formed, presumably from the sandy parent material (Nieuwenhuyse et al., 1994). Mass balance calculations indicate that the soil has undergone little if any detectable removal by weathering. By contrast, the soil even appears to be enriched in Si and Mg (Table 3), which is unlikely. Assuming that Ti is immobile, these apparent enrichments probably result from parent material variability. So, even in these strongly homogenized beach ridge sediments (Nieuwenhuyse et al., 1994), gains and losses may be off by 10% due to inhomogenities in the parent material.

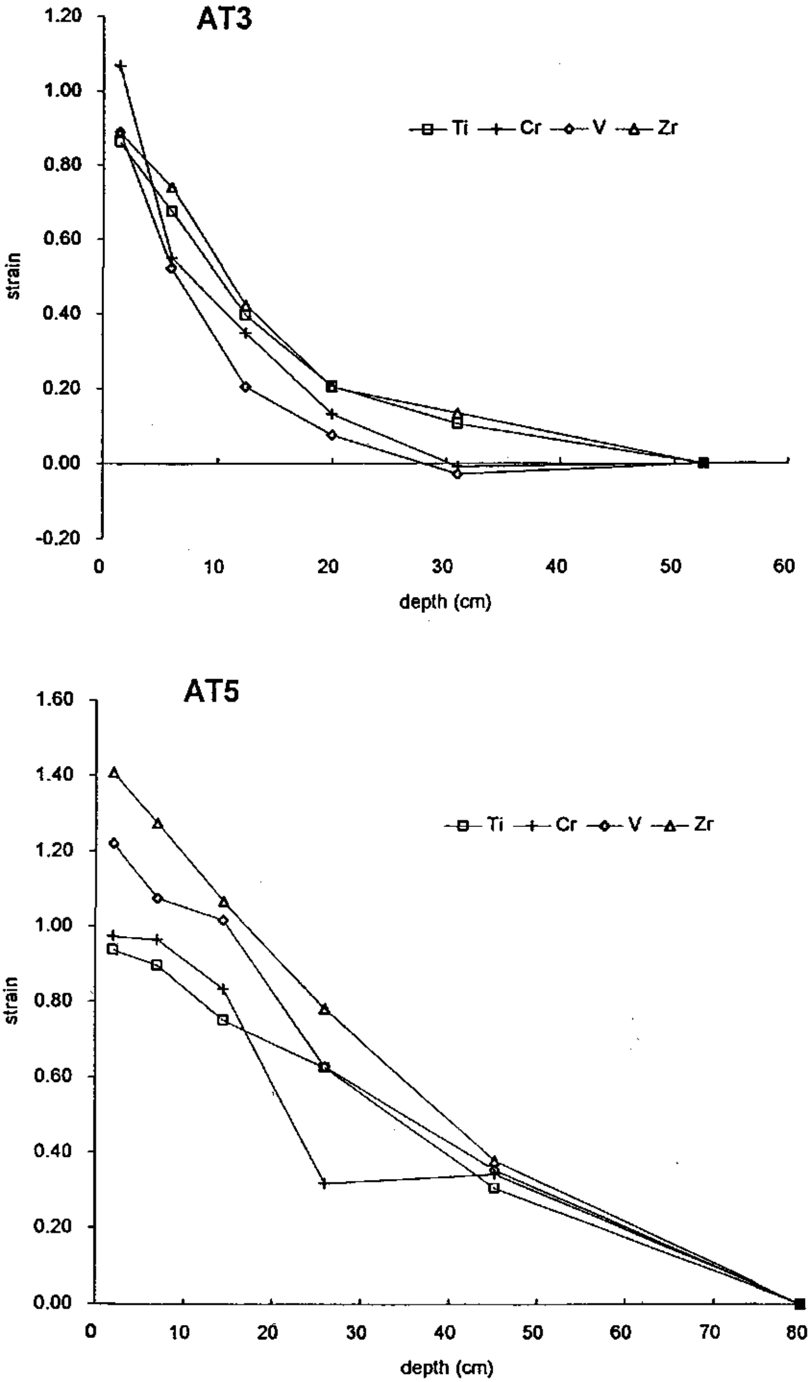
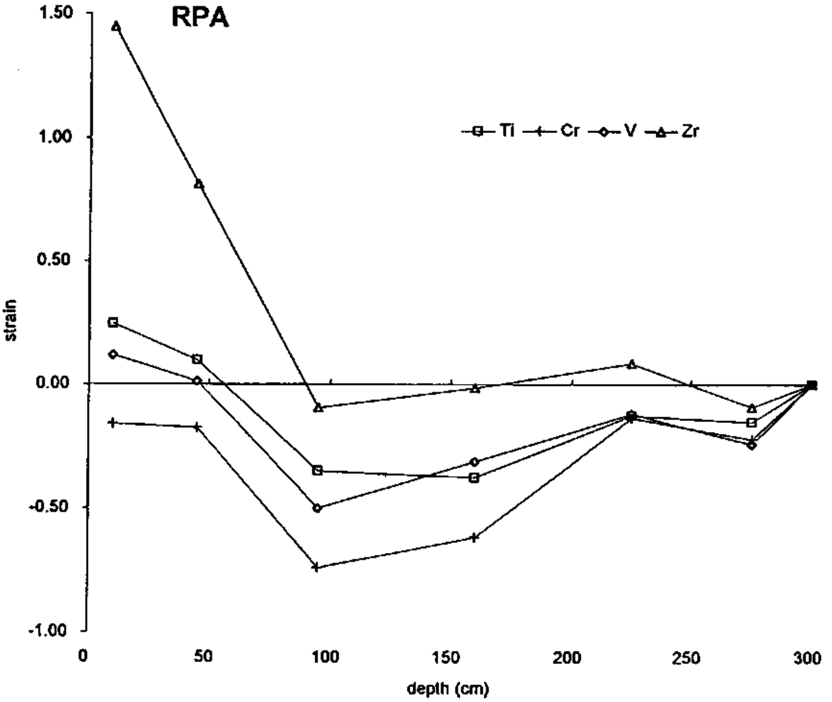
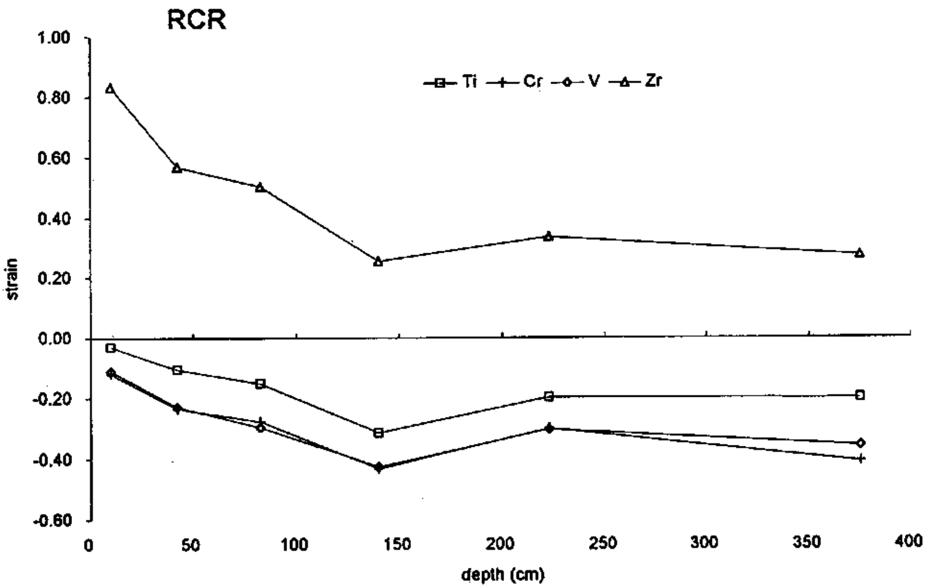


Figure 2: **Strain values for selected profiles.**



RPA
1.50
1.00
0.50
0.00
-0.50
-1.00
strain
Ti
Cr
V
Zr
0
50
100
150
200
250
300
depth (cm)



RCR
1.00
0.80
0.60
0.40
0.20
0.00
-0.20
-0.40
-0.60
strain
Ti
Cr
V
Zr
0
50
100
150
200
250
300
350
400
depth (cm)

2 to 5 ka old Hapludands (AT4, 5, and 7).

These soils have about 1 m thick ABC profiles and contain more clay- and silt-sized material and organic matter in the A and B horizon than the AT3 soil. In all soils, strain values are positive indicating dilation. Greatest dilation occurs near the soil surface (Fig.2), where incorporation of organic matter, structure development, and formation of (bio)pores are maximal. Dominant secondary minerals are short-range order materials and Al- and Fe-humus complexes (Table 2).

Losses of Mg, Ca, Na, and K, and some Si are evident (Table 3), and must be attributed to dissolution of primary minerals. Again, slight variations in parent material composition may have affected the outcome of the calculations. For example, greater calculated losses of Si in AT4 than in AT5 are probably due to vertically heterogeneous parent material. However, it cannot be excluded that for whatever reason, the differences are real, and AT4 has lost more Si than the older AT5.

≤ 18 ka old Melanudand (RPA).

Strain values in this deep ABC profile are slightly positive in the A horizon, but smaller than in the younger soils. Moreover, they become slightly negative in the B (Fig.2). Clearly, the soil has undergone collapse relative to the younger soils in the A horizon and relative to the parent material in the B horizon. Collapse is to be expected when chemical weathering and subsequent leaching has removed part of the minerals, and organic matter incorporation and biological activity are not able to stabilize the remaining mineral structure under the pressure of overlying soil material and vegetation.

With increasing soil age, Andisols of the area become depleted in the most easily weatherable minerals in the fine-earth fraction, especially volcanic glass. This was apparent already in the topsoil of AT7, and becomes more clearly expressed throughout the RPA soil. Volcanic glass is absent from the < 2 mm material, but is still present in fresh cores of coarser fragments. In RPA about 60% of Si, and 50 to 80% of Ca, Mg, Na, and K has been leached (Table 3).

In addition to weathering of the remaining plagioclase and pyroxene, secondary minerals may disintegrate too, and contribute to leaching of Si. Short-range order materials, metal-humus complexes, gibbsite, and kaolin minerals are present throughout the profile (Table 2). Thermogravimetric analyses showed that in the B horizon the gibbsite content of the fine-earth fraction is about 15%, while 1.0 nm halloysite makes up about 18% (data not shown). By mean of micromorphological techniques, Jongmans et al. (1994) showed that X-ray amorphous material in the B horizon was being transformed into gibbsite, thus contributing to leaching of Si.

Table 2: Organic matter content, texture and mineralogical composition of studied soils.

Horizon	Depth	Organic Matter	Texture < 50 µm	> 50 µm	Most important mineral phases[1] > 20 µm	< 20 µm
	-cm-	g kg^{-1}				
Profile AT3 (<0.5 (0.2) ka)						
A	0-16	60	122	818	prim,ls	sro,mh,ls
CA	16-41	10	40	950	prim,ls	na[2]
Profile AT4 (2 ka)						
A	0-21	93	282	625	prim,ls	sro,mh,ls
B	21-50	28	194	778	prim,ls	sro,ls
CB	50-80	4	30	966	prim,ls	na
Profile AT5 (2-4 (2) ka)						
A	0-19	107	384	509	prim,ls	sro,mh,ls
B	19-80	24	185	791	prim,ls	sro,ls
C	80-150	3	20	977	prim,ls	na
Profile AT7 (2-4 (3.5) ka)						
A	0-24	130	435	435	plag,pyr,ls	mh,sro,gi,ls
B	24-80	32	290	678	prim,ls	sro,ls
CB	80-120	8	119	873	prim,ls	na
Profile RPA (<18 (18) ka)						
A	0-70	160	na	na	pyr,pla	sro,mh,gi,kao
B	70-200	24	na	na	pyr,pla	hall,gi,sro
CB	200-300	1	na	na	prim	sro
Profile RF1 (>50 (125) ka)						
A	0-10	78	876	46[3]	op,gi	gi,kao,goe
B	10-200	10	911	79[3]	op,gi	gi,kao,goe
CB	200-610	0	480-990	520-10	op,gi	hall,gi,goe,cri
Profile RF2 (>50 (125) ka)						
A	0-10	83	826	91[3]	op,gi	gi,kao,goe,ls
B	10-195	10	901	89[3]	op,gi	gi,kao,goe,ls
CB	195-570	0	50-810	190-950	prim,op,gi	hall,sro,cri
Profile RCR (<450 (450) ka)						
A	0-8	104	762	134[3]	op,gi	gi,kao
B	8-387	9	832	159[3]	op,gi	gi,kao
CB	387-500	0	na	na	prim,ls	na

[1] cri = cristobalite; gi = gibbsite; goe = goethite; hall = halloysite; kao = kaolin minerals; ls = 2:1 layer silicates; mh = Al- and Fe-humus complexes; op = opaques; pla = plagioclase; prim = primary minerals (volc.glass, pyroxene, plagioclase, opaque minerals); pyr = pyroxene; sro = short range order material. Mineral phases are listed in order of importance.

[2] na = not analyzed.

[3] mostly present as sand-sized gibbsite.

$\geq$50 to 450 ka old Haploperox (RF1,2; RCR).

Horizon differentiation in the oldest soils is less pronounced than in the younger soils. The thick, dark A horizon present in the Andisols from which these soils are thought to have formed, is absent. Instead, the older soils have a < 15 cm thick A horizon and a > 1 m thick, homogeneous B horizon, which gradually grades into a saprolitic CB horizon.

Strain values are negative throughout the profile, indicating collapse of up to 65% of the volume occupied by the parent material (Table 3; Fig.2). Collapse is greater in both soils formed on alluvial sediments (RF) than in the soil formed on lava (RCR), in spite of greater leaching in the older soil. This is probably due to the much greater porosity of river sediments compared to that of lava. Coarse textured layers in the RF1 soil show a greater collapse than fine textured layers (Table 3).

Except for small quantities of opaques, the soils have lost all primary minerals, and consist mainly of gibbsite, kaolin minerals, and crystalline iron oxides (Table 2). If these soils indeed have gone through an Andisols stage, weathering apparently leads to dissolution or transformation of short-range order materials and metal-humus complexes. In both RF soils differences caused by sedimentary layering have been erased in the part of the profile influenced by biological activity. In both profiles, layering is still clearly present in the CB horizons which are virtually devoid of biological activity.

The two alluvial soils RF1 and RF2 are thought to have the same age (they are located about 400 m from each other), but have lost different amounts of various elements. RF1 lost much more cations and Fe, and became enriched in Al (relative to Ti, indicating some mobilization of Ti). However, the two soils lost similar amounts of Si. A striking aspect is that RF1 lost Si, Mg, and K mainly from the coarse sedimentary layers (Table 3), and less so from the fine sedimentary layers. By contrast, more Fe and Ti (as deduced from Al-enrichment) appear to have been lost from the finer than from the coarse layers. In RF2 there is no clear relationship between sedimentary layer type and leaching behaviour. These patterns can be explained by differences in hydrology between the two profiles, with RF1 acting as an aquifer for lateral drainage from a larger area, and RF2 being well-drained and subject mostly to steady vertical percolation of rain water infiltrating at the soil surface only. That would be consistent with RF1 (1) being generally more strongly leached than RF2; (2) showing strongest desilication in its coarse layers (due to more rapid percolation in coarse- than in fine-textured layers), and (3) being at least periodically watersaturated and anoxic (particularly in the fine-textured layers with their greater waterholding capacity), leading to reduction and mobilization of Fe, and perhaps Ti. Periodical reduction and oxidation in this profile is apparent from mottling. Stronger leaching of coarser layers is corroborated by the greater collapse of such layers.

The RCR profile seems to have been weathered more homogeneously: all horizons have been leached more or less to the same extent. The original rock structure is morphologically visible from about 4 m depth downward. Virtually all basic cations and 60 to 90% of the Si present originally have been leached from shallower depths.

Table 3: Calculated losses and gains of mayor elements for studied soil horizons influenced by weathering and neoformation, using Ti as immobile element.

Depth	strain	Si	Al	Fe[1]	Mg	Ca	Na	K	P
- cm -		- % of parent material lost (-) or gained (+) -							
Profile AT3 (<0.5 (0.2) ka)									
0-4	0.86	10	8	4	-4	1	7	9	7
4-10	0.67	4	2	8	2	-2	-1	-6	3
10-16	0.40	2	3	4	-0	-1	-2	-1	7
16-41	0.15	2	1	6	5	0	-2	-2	7
mean[2]	0.33	3	2	6	4	0	-1	-1	7
Profile AT4 (2 ka)									
0-5	0.60	-5	-6	-4	-10	-19	-17	-6	-2
5-12	0.57	-8	-6	-5	-16	-24	-19	-6	-9
12-21	0.57	-8	-4	-5	-22	-29	-18	-4	4
21-38	0.33	-12	-4	-5	-19	-30	-21	-9	1
38-50	0.01	-7	-3	-8	-19	-23	-10	2	-16
50-80	0.03	-5	-2	-5	-14	-20	-10	5	-19
mean	0.23	-8	-3	-6	-16	-24	-14	-1	-10
Profile AT5 (2-4 (2) ka)									
0-4	0.94	-1	0	-1	-22	-32	-17	-13	27
4-10	0.90	-3	0	-1	-24	-36	-18	-15	24
10-19	0.75	-3	0	-1	-26	-37	-17	-12	13
19-33	0.63	-5	3	-1	-22	-36	-17	-7	13
33-80	0.30	-3	3	1	-8	-15	-6	-3	-8
mean	0.49	-3	2	0	-15	-24	-11	-6	2
Profile AT7 (2-4 (3.5) ka)									
0-2	3.00	-2	-29	5	-16	-23	-42	-45	-6
2-12	1.89	-21	-27	4	-26	-37	-51	-50	-42
12-24	1.92	-10	10	11	-22	-32	-36	-30	-31
24-35	1.38	-17	-3	4	-14	-29	-36	-29	-52
35-54	0.74	-8	5	11	-8	-21	-27	-22	-54
54-80	0.68	-11	7	7	-21	-29	-24	-23	-53
80-120	0.21	-8	-1	6	-8	-17	-17	-12	-65
mean	0.86	-11	0	7	-14	-25	-27	-23	-53
Profile RPA (≤18 (18) ka)									
0-20	0.24	-63	-13	-6	-57	-88	-96	-90	10
20-70	0.10	-69	-4	-5	-54	-89	-99	-94	-12
70-120	-0.35	-74	-7	-6	-94	-99	-100	-96	-61
120-200	-0.38	-73	2	-4	-84	-97	-100	-97	-63
200-250	-0.13	-37	14	5	-4	-58	-68	-70	-45
250-300	-0.15	-18	-6	-5	-3	-23	-22	-43	-8
mean	-0.17	-57	-1	-3	-52	-77	-81	-82	-37

Table 3: continued

Depth - cm -	strain	Si	Al	Fe[1]	Mg	Ca	Na	K	P
		% of parent material lost (-) or gained (+)							
Profile RF1 (>50 (125) ka)									
0-100	-0.41	-71	8	0	-99	-100	-100	-97	-68
100-200	-0.46	-76	4	-5	-98	-100	-100	-100	-74
200-290c[3]	-0.65	-82	-12	-31	-98	-100	-100	-100	-52
290-375f	-0.20	-28	24	-23	-84	-100	-100	-94	-61
375-445c	-0.40	-54	9	-1	-96	-100	-100	-96	-37
445-470f	-0.20	-26	25	-36	-85	-100	-100	-93	-59
470-490c	-0.34	-46	25	-27	-95	-100	-100	-96	-60
490-510f	-0.22	-26	17	-47	-79	-100	-100	-85	-49
510-545c	-0.53	-63	-21	-13	-95	-100	-100	-98	-48
545-560f	-0.15	-16	19	-44	-78	-100	-100	-82	-51
560-595c	-0.35	-46	13	-16	-94	-100	-100	-96	-59
595-610f	-0.11	-20	31	-4	-75	-100	-100	-88	-34
mean	-0.40	-57	8	-15	-93	-100	-100	-96	-58
Profile RF2 (>50 (125) ka)									
0-75	-0.46	-77	3	0	-98	-100	-100	-97	-62
75-150	-0.48	-80	3	4	-99	-100	-100	-99	-73
150-240c	-0.62	-84	-5	-1	-97	-100	-100	-99	-40
240-300vc	-0.59	-74	-34	-1	-66	-97	-95	-77	-22
300-330c	-0.34	-46	-9	-2	-51	-86	-80	-33	-9
330-340vc	-0.36	-53	5	-2	-55	-93	-88	-45	14
340-420c	-0.23	-37	3	-3	-29	-73	-75	-4	-7
420-425vc	-0.31	-50	25	6	-86	-99	-100	-78	-53
425-440f	-0.34	-44	-7	-10	-82	-98	-100	-94	-68
440-520c	-0.21	-34	23	-13	-56	-85	-86	-33	-9
520-570c	-0.16	-34	16	2	13	-41	-81	-38	-23
mean	-0.40	-60	1	-2	-65	-88	-91	-64	-33
Profile RCR (<450 (450) ka)									
0-8	-0.03	-84	-7	9	-99	-100	-100	-100	-81
8-68	-0.11	-85	-6	10	-99	-100	-100	-100	-88
68-135	-0.15	-89	-2	9	-99	-100	-100	-100	-92
135-227	-0.31	-93	-3	8	-99	-100	-100	-100	-96
227-307	-0.20	-87	-5	0	-100	-100	-100	-100	-95
307-387	-0.16	-87	6	14	-100	-100	-100	-100	-87
mean	-0.19	-88	-2	8	-100	-100	-100	-100	-92

1 Iron has been reported as Fe_2O_3, although actually both Fe_2O_3 and FeO may be present.

2 Weighted average over studied depth.

3 Sediment type: c = coarse f = fine vc = very coarse.

ELEMENTAL BEHAVIOUR AND MINERALOGY

All soils older than 0.5 ka have lost basic cations and Si. Basic cations are liberated simultaneously with Si upon weathering of primary minerals. Since under the prevailing conditions basic cations are not retained in significant amounts in any of the neoformed products, they are leached from the soil profile. When cations are set free from primary minerals, the Al-Si-O frameworks of the original silicate minerals are partly reconstituted into secondary minerals (e.g. allophane-like minerals), partly leached and lost from the soil profile. Some Mg is retained in many of the soil horizons, probably in 2:1 layer silicates that contain Mg. Such layer silicates occur in most studied soils from very young to old. Formation of 2:1 layer silicates is often related to dry climates or poor drainage conditions (Allen and Hajek, 1989; Borchardt, 1989), so their occurrence in the studied soils is surprising. Most likely, they are inherited from minerals formed by hydrothermal transformation in the volcanoes and subsequently deposited as volcanic ash or alluvial sediments after erosion (Jongmans et al., 1994). Although 2:1 layer silicates usually are thought to be unstable in the well drained tropical environment (Allen and Hajek, 1989), these minerals are remarkably stable in our soils. This may be due to the dense structure of the sand sized bodies in which they occur, and to Al incorporation in interlayers.

In most soils P is leached to some extent. However, in the AT5 and RPA soils, P is slightly enriched in the topsoil. A similar enrichment was observed in the A horizon of Andisols in Vanuatu by Quantin (1992). At the RPA site, topsoils spatially varied strongly in their P_2O_5 content from 0.6 to 1.6% (the lower value was used in calculations). Surface accumulation caused by nutrient pumping and input of airborne P which is retained in the topsoil by short-range order materials and metal-humus complexes (Wada, 1989), may explain this enrichment. Input of airborne P in the study area was estimated to be 0.1 to 0.2 kg $ha^{-1}yr^{-1}$ (Parker, 1985), and the pool of P cycling between the soil and the tropical forest vegetation has been estimated to be about 30 to 100 kg P ha^{-1} (Jaffré, 1985). Since at the RPA site indian pottery was found, we speculate that also P brought into the soil by indian activity has locally contributed to the enrichment.

Although Al is generally highly immobile, in the AT7 and RPA soils it appears to have been removed from the A horizon and accumulated in B horizons (Table 3). Although it cannot be excluded that this partly is due to parent material variability, it may also be due to transport of Al as complexes with organic matter, which is abundant in the A horizons of these soils.

Using equation 3 and most likely age of the soils (Table 1), average annual elemental gains and losses per hectare during the time pedogenesis has been active can be estimated (Table 4). For the RF 1 and 2 profiles as well as for the RCR soil, these estimates must be considered as minimal values, since the weathered soil extends to a greater depth than the studied soil profile. Furthermore, part of the elements originally present in the parent material may have been removed by erosion of surface soil material. Present-day leaching from the upper 5 to 6 meters of the older soils will be lower than indicated in Table 4, because they are now almost devoid of easily weathereable elements. Obviously cation losses from the weathering material are important only during a relatively short time period.

Table 4: Average gains and losses of elements during time pedogenesis has been active[1], using Ti as immobile element.

Profile	Depth	Si	Al	Fe	Mg	Ca	Na	K	P
	cm					kg ha^{-1} yr^{-1}			
AT4	80	-99	-17	-13	-16	-48	-11	-1	-0.4
AT5	80	-91	26	-0	-24	-54	-15	-5	+0.2
AT7	120	-107	-2	19	-22	-47	-13	-9	-2.4
RPA[2]	300	-664	-4	-5	-44	-137	-64	-31	-2.8
RF1	610	-92	+5	-6	-11	-17	-10	-9	-0.5
RF2	570	-90	+1	-1	-8	-17	-10	-6	-0.3
RCR	387	-51	-0	+1	-5	-13	-6	-4	-0.3

[1] Using the likely soil age mentioned in Table 1.

[2] Values for fine-earth fraction of the soil. It should be kept in mind that an estimated 40 to 60% of the soil mass is composed of coarse fragments, which are weathered to a much lesser extend. True leaching losses are thought to be about half of these values.

As indicated by data from the RPA profile where only little of the basic cations that were originally present are left in the fine-earth fraction, the length of this time period in sandy or finer deposits may be about 20 to 50 ka. If soils are preserved for longer time periods, no significant basic cation reserves are present in the part of the profile which can be exploited by roots.

For several months, soil solutions of soils similar to the RF and AT5 were sampled using commercial Si-free samplers. Concentrations of basic cations below the rooting zone of both soils were sometimes in the same order of magnitude as those in incoming rainwater, indicating that atmospheric inputs may play an important role in determining base cation concentrations in the soil solution. Si concentrations were significantly lower in the older soil (13 to 99 mmol m^{-3}), than in the younger soil (72 to 225 mmol m^{-3}). Moreover, Si-concentrations in rainwater were negligible (0 to 2 mmol m^{-3}). These data indicate that actual Si leaching from the older soils is considerably less than from younger soil with abundant primary minerals.

We compared data from Table 4 with preliminary leaching data from rivers. However, data on solute concentrations in stream waters are less useful in this regard, because of possible contribution of solutes from active weathering zones at depth below older soils. Assuming an annual average drainage of 1.5 to 2.5 m of water (= 1.5 to 2.5 10^4 m^3 ha^{-1}) and a negligible input of Si by atmospheric deposition, the annual losses of this element from the studied soils (Table 4) should be reflected in stream water concentrations of rivers draining such soils by concentrations of 73 mmol Si m^{-3} in case the RCR leaching rate is correct, to 785 mmol Si m^{-3} assuming the RPA leaching rate prevails in the whole area. Samples were taken at various points in the upper and middle part of the Tortuguero river, whose upper and middle basin mainly comprise young soils on volcanic sandy sediments similar to the studied ones. Si in river water was analysed in samples taken during a short dry spell when the river presumedly drained base flow water from the weathering zone. The Si concentrations were 490 to 890 mmol Si m^{-3}, similar to the Si-concentrations expected from the RPA profile balance, which supports the soil mass balance calculations.

CONCLUSIONS

1: Weathering and formation of secondary minerals on andesitic parent material in the humid tropical Atlantic lowland of Costa Rica follow pathways as described by other authors in different regions of the world.

2: Ti appears to be a reasonable index element at least under oxic weathering, and permits quantification of element behaviour. However, even the most homogeneous parent material had a variability in chemical composition that caused errors in apparent gains and losses of up to 10% of the initial element pools.

3: Neither the short-range order materials and metal-humus complexes, nor the tropical rain forest vegetation are capable of retaining a significant part of the basic cations released by weathering.

4: In a <18 ka old Melanudand about 50% of Mg, and about 80% of Ca, Na, and K are leached from the upper meters. In soils older than 50 ka, Mg, Ca, Na, and K reserves are completely depleted to depths >2 m.

5: In Andisols in the study area Si losses become clearly measurable between 2 and 18 ka. The mean annual Si losses over the lifetime of the soils studied, calculated from mass balance and soil age, are in the same order of magnitude expected from Si leaching losses estimated from the composition of base flow water draining such soils.

6: In Andisols of the study area, P may be accumulated in A horizons by P-retention and possibly human activity. However, the result on older Oxisols indicate that mineral P reserves are gradually lost from the soil profile, although even after long time periods (50 to 450 ka) still mineral P reserves are present.

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CHAPTER 8: SYNTHESIS

The papers presented in this thesis reflect the results of a research project focused on influence of volcanism on landscape genesis, and formation of soils on volcanic parent material in the humid tropical Atlantic lowland of Costa Rica. Although these topics have been studied extensively elsewhere, the area presents a combination of environmental conditions which permit research approaches that are often not possible in other volcanic regions. These conditions, which determine landscape and soil formation in the study area are discussed briefly below:

1 **A highly dynamic tectonic and volcanic environment.** Although in the Limón basin itself volcanism and strong earthquakes are absent, in the surrounding mountainous areas they are frequent, and influence profoundly fluvial processes in the lowland (chapter 4).

2 **Regular catastrophic rainfall.** Almost every year rainfall events of 100 - 300 mm day^{-1} occur. Under normal circumstances, such huge amounts of water can be drained reasonably rapid and do not cause major inundations, except in areas at < 20 m above sea level. However, when watersheds are severely disturbed by the deposition of ash, by landslides and/or by destruction of vegetation by volcanic eruption or earthquakes, the transport capacity of river systems may be exceeded after heavy rainstorms; often leading to catastrophic inundations and deposition of thick sediments (chapter 4). When no inundations occur, river carry huge amounts of volcanic ejecta to the sea, where they are distributed by marine currents and wave action and influence coastal landscapes (chapter 3).

3 **Lava, fluvial, and marine volcanic deposits of similar chemistry and mineralogy.** The beach ridge deposits described in chapter 3, as well as the fluvial sand deposits described in chapter 4, are often thick and homogeneous and consist of almost pure andesitic sand of a mineralogical and chemical make up similar to that of lavas. In such thick deposits soils can form whose formation is not interrupted by later deposition of different nature or by underlying layers, as often occurs in other fluvial sediments. Also lava flows show little variation in chemical and mineralogical composition, although texturally they are heterogeneous. Chemistry of soils parent material is similar, facilitating budget calculations (chapter 7) and comparison of soil formation (chapter 5 and 6). Textural differences, on the other hand, may lead to differences in weathering pathways.

4 **Near absence of volcanic ash deposition on large parts of the area.** Volcanic areas are often dominated by extensive ash deposits of different age. As argued in chapters 5, 6, and 7, there is little evidence that significant amounts of volcanic ash have been

deposited in the study area, except perhaps in the footslopes of the Central cordillera. This is due to dominant northeastern to eastern trade winds. However, during the 1963 - 1965 eruptions of Irazú volcano, ash reached the footslope areas around the town of Guápiles at least twice, and reached the city of Limón once (Barquero, 1977). This indicates that ash addition cannot be excluded. According to eyewitnesses, ash was of dust size and just concealed the surface of leaves and roofs, so the deposited amount may have been less than 1 mm. On the other hand, as mentioned in chapters 5 and 6, and indicated by unpublished data of the author on a 6200 years old peat in the northeastern part of the area, in large parts of the area the deposited amount must have been minimal since not even traces were observed in thin sections, or in loss-on ignition residues. Contrary to many other regions bordering volcanic mountain chains, landscape formation in the basin is not influenced by deposition and erosion of ash deposits. Furthermore, at least the Holocene soils studied in this thesis have not been influenced significantly by ash additions. Nevertheless, even small additions may affect results of chemical budget calculations as presented in chapter 7.

5 **A high constant temperature, and rainfall exceeding evapotranspiration throughout the year.** The climate is rather even throughout the year over the whole lowland, so landscape forming processes are not influenced by strong seasonal climatic changes. Also, differences in soil formation in the lowland cannot be ascribed to variations in soil moisture or temperature regime, as in many other volcanic areas (Parfitt et al., 1983; Chartres and Pain, 1984; Quantin, 1985). Mean annual air temperature is about 25°C and varies as little as 2°C between the coolest (January) and hottest (June) month. It decreases with altitude at a rate of about 0.52°C every 100 m (Herrera, 1985). The research presented in chapter 3 to 7 of this thesis was carried out from sea-level to about 300 m, where differences in temperature are probably insignificant. Rainfall in this area varies from 3500 to 5500 mm annually and is well distributed throughout the year. It exceeds potential evapotranspiration throughout the year, as shown for Tortuguero in chapter 6. These climatic conditions favour rapid continuous soil formation.

6 **Hydrology ranging from well-drained to poorly drained.** Large variations in drainage occur in the fluvial lowland as well as in the beach ridge plain due to small differences in relief, blocking of natural drainage systems by events as described in chapter 4, and presence of impermeable layers in the subsoil. Therefore, it is possible to study soil formation as a function of drainage.

Different combinations of these factors lead to particular landscape and soil forming processes, some of which were investigated in the present study:

8.1. Landscape Genesis

8.1.1. Influence of volcanism on fluvial processes (channel shifts)

Volcanic landscapes are shaped by a sequence of nearly instantaneous events lasting 10^0 to 10^1 years, such as eruptions of ash and lava, followed by periods of 10^1 to 10^2 years in which strong erosion prevails and vegetation slowly restores. If not interrupted by new eruptions, erosion slows down during the following period (10^2 to 10^3 yrs). Much of the erupted material is transported into the sea, although a significant part is stored on mountain slopes and in alluvial plains.

During and shortly after eruptions, rivers in the affected area behave different as compared to periods of volcanic quietness. The volcano slopes influenced by an eruption undergo extreme erosion, and rivers which flow through such areas receive a much higher amount of sediment than during (quiet) periods without ash deposition. Apart from normal fluvial processes, events such as mudflows and hyperconcentrated flood-flows may occur in more proximal areas such as alluvial fans, as indicated by Kesel and Lowe (1987). In chapter 4 we show that also in distal parts of watersheds eruptions profoundly affects the landscape. As a result of heavy rainfall, inundations occur, and thick sand layers are deposited on levees and overbanks in a time period of hours or days. Since also channels may be clogged with sediment, rivers are forced to seek new channels. We argue that landscape formation in fluvial plains is dominated by such exceptional floods which choke river channels. Afterwards, the landscape stabilizes and some erosion occurs, giving rise to formation of drainage gullies and incision of existing river channels.

Recently, these phenomena could be observed on a small scale after one single phreatic eruption of Irazú volcano in 1994 in an area with fumaroles. The eruption caused hardly any ash fall, but destroyed vegetation and caused many mass movements in an area of about 25 ha. During the following year, the Sucio river which drains this area was saturated with sediments and changed its course frequently in its proximal alluvial fan area. Further downstream no effects on the river course were observed. During this period, most of the transported material must have reached the sea. In August 1995, a lahar-like event (probably triggered by a high rainfall event) filled up the proximal alluvial fan area with about 1 m thick, very course sediment. After this, the river was observed to transport less sediment, and at the moment of writing (December 1995), appears to transport about the same amount of sediment as before the eruption. Just as the crevasse splay event of Tortuguero river described in chapter 4, this is an example of volcanic activity which influences the landscape of only a limited area.

8.1.2. Coastal plain development influenced by volcanism

When the rapid sea level rise during the Early Holocene slowed down about 6000 years ago (Fairbanks, 1989), on many coasts sandy beach ridge plains could form wherever sediment supply exceeded coastal erosion. On the Caribbean coast of Costa Rica formation of a beach ridge plain started about 6000 years ago (chapter 3). We show that material from volcanic eruptions which reaches sea is distributed along the coast and subsequently deposited as beach ridges. We argue that even individual eruptions can be recognized, illustrating the great importance of volcanism for their formation. Since active volcanism is the main sediment source, it can be speculated that without such catastrophic events few or no beach ridges would have formed and that the coastline would have been different from the actual one.

8.1.3. Pleistocene terrace levels (climatic change?)

As mentioned in chapter 2, two distinctively different terrace levels are found in the area: Pleistocene terrace remnants and Holocene floodplains. The occurrence of the high Pleistocene terraces merits some attention.

The dissected remnants of Pleistocene deposits are at about 5 to 25 m above the actual floodplain deposits. Their high position within a (supposedly) subsiding area indicates that they may have been deposited during a former high sea level stand, and were subsequently dissected during glacial periods with lower sea levels. Assuming that (1) these sediments were deposited 130,000 years ago during the Eemian/Sangamonian interglacial, and (2) the subsidence rate has been 0.25 mm yr^{-1} since then (chapter 2), these terrace levels are consistent with a former sea level at least 42.5 to 57.5 m higher than at present. Their degree of soil development is in accordance with Eemian/Sangamonian age. However, taking into account that Eemain/Sangamonian sea levels were only slightly higher than at present (Zagwijn, 1983), there is little room for any subsidence. Furthermore, De Jong (1994) encountered shells in Pleistocene deposits at 4-6 m below actual sea level dated at 34,000 BP (GrN-19968), a period during which sea level reconstructed on the base of raised coral terraces at Huon Peninsula, Papua New Guinea and Barbados was estimated to be situated at around - 35m. Both data indicate that at least during the last 150,000 years the coast was stable or slightly uplifting. The subsidence rate of 0.25 mm yr^{-1} calculated from the total thickness of Quaternary deposits in the extreme northeast of the Limón graben was apparently not uniform over time, or is only valid for certain parts of the graben. Especially toward the Central cordillera and in the western part of the studied area this explanation seems feasible. Furthermore, terrace remains are absent at about 10 km or less from the coastline where RECOPE oil company actually did measure subsidence.

The presence of these old terraces can also be explained by a higher sediment supply in the past, e.g. due to a more active volcanism (e.g. by caldera explosions of volcanoes of the Central cordillera, see e.g. Prosser and Carr, 1987), which led to formation of an absolute higher landscape. However, one would expect that in that case sediments would be coarser than surrounding Holocene sediments, for which there is no evidence (Van Ruitenbeek, 1992).

8.2. Soil Genesis

8.2.1. Rates of soil formation

Compared to colder and dryer volcanic regions of the world, soil formation in the Atlantic region is rapid. On andesitic beach ridge sediments, as well as on sandy andesitic channel fill sediments, Hapludands form within 2000 years (chapters 4, 5, and 6). Similar values have been found in other humid tropical areas on volcanic parent materials, e.g. Quantin et al. (1991). However, most soils described in literature have formed on tephra deposits which may be richer in glass than the alluvial and marine sediments in the study area, which contains less than 20% glass. Furthermore, tephra deposits often contain more silt-sized material, which tends to weather more rapidly than the sand-sized material, (chapter 5). That weathering rate in the Atlantic zone were as high as those elsewhere, in spite of lower ash content and higher texture, indicates that environmental factors in the study area indeed promote a fast rate of soil formation.

8.2.2. Mineralogical evolution

In soils on volcanic parent materials

Under well-drained conditions in humid climates, short-range order materials usually dominate the clay minerals formed in the upper layers of young Holocene soils formed in volcanic materials (Lowe, 1986; Mizota and Van Reeuwijk, 1989; Shoji et al., 1993). Soils of the study area do not present exceptions to this findings, and mineralogical evolution of soils follows pathways also described in other volcanic areas of the world (e.g. Lowe, 1986; Quantin, 1992)

The mineralogy of sandy sediments derived from the Central cordillera is essentially similar to that of volcanic rock in this cordillera. Because, moreover, no significant differences in the soil forming factors climate, vegetation, and human influence occur, the differences in development between soils formed on these sand deposits can be ascribed to the soil forming factors time, parent material texture, and drainage.

Under well-drained conditions, chemical weathering of volcanic glass, plagioclase, and pyroxene is rapid in the hot and humid Caribbean lowland (chapter 6). Formation of short-range order materials from these minerals and incorporation of soil organic matter is thought to be the main source of the increasing amount of fine material with increasing soil age. As a consequence, texture and other soil properties are correlated with age (chapter 5).

Coarse sandy parent material weathers more slowly than fine sandy parent material (chapter 6), due to the smaller surface area at which weathering can act. Furthermore, for the same reason, soils containing coarse fragments often present fresh primary minerals next to a weathered soil mass. It is obvious that as the content of coarse fragments in the parent material increases, the less soil-weathering has proceeded. On the other hand, the fine-earth fraction between the coarse fragments is further weathered than that of a similar soil without coarse fragments due to more intense leaching (e.g. the RPA soil in chapter 7). Parent material texture thus clearly influences weathering.

Sands deposited by catastrophic events as described in chapter 4 occur throughout the alluvial plains. Soils formed on these deposits differ in mineralogical composition according to drainage position: short-range order material dominates the clay fraction in well-drained soils, and halloysite at poorly drained sites (chapter 2). Even within one soil profile these differences in mineralogy may be found: short-range order material dominates in the better drained upper part of a soil profile, while halloysite is practically the only mineral present in lower horizons.

Given the fairly uniform parent material, the identical age of the soils and the similar environmental conditions of the sites, the observed differences in soil properties and mineralogy must be ascribed to different drainage conditions. This confirms the finding in literature that a stagnant moisture regime leads to relatively high Si concentrations in the soil solution, which favour formation of halloysite over short-range order material (Parfitt et al., 1983). Strong leaching leads to relatively low solute concentrations, while higher solute concentrations may develop under restricted leaching. Especially the silica concentration in the soil solution was found to be of paramount importance for the formation of secondary clay minerals (Parfitt and Wilson, 1985). In New Zealand soils, concentrations of dissolved Si lower than 7 to 10 g/m^3 were found to favour formation of allophane, while higher Si concentrations favour halloysite formation (Singleton et al., 1989). Halloysite is also described as the main secondary product in volcanic soils developed in climates with a pronounced moisture deficit and in deeper soil horizons (e.g. Aomine and Mizota, 1973; Bleeker and Parfitt, 1974). These observations make clear that halloysite does not necessarily form only, and slowly, from short-range-order precursors like allophane, as supposed by e.g. Fieldes and Claridge (1975), but that both short-range order materials and halloysite may form directly from volcanic parent materials (e.g. Wada and Kakuto, 1985).

However, it is somewhat paradoxical that formation of short-range order material seems to be favoured by the presence of easily weatherable parent rock, (giving relatively high concentrations of solutes) while in those conditions, formation of crystalline halloysite is favoured over that of allophane by higher concentrations of Si. Future research will be needed to explain what exactly favours halloysite formation.

Upon further weathering, short-range order materials disintegrate, and halloysite and/or gibbsite forms (e.g. profile RPA in chapter 7). Environmental factors probably determine whether kaolin minerals or gibbsite form eventually. Although both halloysite and gibbsite may form directly from primary minerals in volcanic parent materials (Parfitt and Wilson, 1985, Lowe 1986; Hsu, 1989), halloysite is often thought to form by resilication from short-range-order materials (Wada, 1989). Gibbsite may form by desilication of secondary minerals. The presence of halloysite and gibbsite in the same soil horizons, as often found in the well drained older soils of the area, is therefore paradoxical: while halloysite seems to form in silica-rich solutions, gibbsite is related to silica poor conditions. The simultaneous presence of these minerals may therefore be indicative for different conditions to which the soil has been exposed at different times. Halloysite could have formed during dryer climatic conditions (e.g. during glacial maxima in the Pleistocene), while gibbsite formed in more humid climates. In thin sections halloysite was observed to transform into gibbsite, but the abundance of kaolin minerals indicate that this transformation is slow. On the other hand, Jongmans (1994) showed that differences on a microscale in weathering environment, especially in the deeper parts of a profile, may lead to simultaneous formation of different clay minerals. Weather gibbsite and halloysite can form simultaneously from short-range order material in different microenvironments or sequentially due to changes in climate or hydrology, remains an open question.

Other parent materials

On other rock or sediment types, weathering follows different pathways. Holocene fine sediments are composed mainly of a mixture of 1:1 and 2:1 layer silicates, with minor short-range order material, gibbsite, and iron compounds (chapter 2). Variations in mineralogy of soils on such sediments appear to be a consequence in differences in deposited material. With increasing age, it may be expected that continuous leaching alters mineralogical composition as it does in sandy textured sediments. Since little evidence for such changes were found, it appears that they are too young to permit further mineralogical differentiation, or that such mineralogical changes proceed much more slowly than in sandy soils.

In the saprolites of the Pleistocene terraces, remnants of sedimentary structures (e.g. fine horizontal lamination) often indicated that sediments were fine-textured. Since neither geological setting, nor source areas appear to have changed significantly over several hundreds of thousands of years (chapter 2), it seems reasonable to assume that the clay minerals present in these sediments were deposited as a mixture of 1:1 and 2:1 layer silicates, just as fine sediments today. Nowadays, these layers are composed almost exclusively of halloysite, indicating that 2:1 layer silicates transform into halloysite under continuous leaching.

8.2.3. Changes in sediment mineralogy by fluvial transport

Formation of clay minerals appears to occur less in young fine textured alluvial soils than in young sandy soils, due to the relatively greater stability of the deposited clay minerals in the prevailing environment compared to primary volcanic minerals. Rather than formed *in-situ*, clay minerals are thought to have been derived from the cordilleras and to a lesser degree, from erosion in of the alluvial fan and plain deposits. Thus, sedimentary parent materials should contain clay minerals similar to those formed in the source areas: short-range order materials and halloysite from volcano slopes and alluvial fans, kaolin minerals and gibbsite from older (Pleistocene) deposits and saprolites, and interlayered 2:1 layer silicates from hydrothermally altered rocks in the Central cordillera (Jongmans et al., 1994).

In < 10 μm fresh suspended sediments of rivers draining the Central cordillera dominant clay minerals are 1.4 nm minerals and 0.7 nm kaolinite or halloysite. The presence of short-range order material was suspected because of limited peak height and sharpness, but was not investigated (Van Seeters, 1993). No 1.0 nm halloysite was observed. Soil formed on fine textured sediments (chapter 2) are usually dominated by a mixture of 1:1 and 2:1 layer silicates, with little or no short-range order materials and gibbsite. Thus, mineralogical composition of clay-sized material in source areas, in transported river sediments and in fine-textured fluvial deposits is not the same. Differences in amount, or absence of a certain clay mineral in soils formed on fine textured sediments compared to presence in source areas may be ascribed to several causes:

(1) Differences in sediment supply in source areas. Streams which derive their sediments mainly from saprolites in areas with frequent landslides are likely to contain more layer silicates than streams where sediment is provided mainly by surface erosion of Andisols dominated by short-range order materials. However, in the lower reaches of the main rivers that cross the alluvial plains, it may be expected that such differences are levelled out by mixing.

(2) Particle size. It is to be expected that certain clay minerals remain dispersed during transportation in stream water. Others, for example short-range order materials, may flocculate and are concentrated in the coarse clay and silt fractions. In older soils gibbsite was often found to form rather large crystals, which consequently favour their transport in the silt and sand fraction.

The reason why clay mineralogy of soils in source areas of watersheds and poorly drained clayey deposits in distal parts are different, merits further attention. Especially the behaviour of eroded short-range order materials and halloysite in stream water and sedimentary deposits appear worthwhile to study in some detail.

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SUMMARY

The influence of volcanism on landscape genesis, and formation of soils on volcanic parent material was studied in the Atlantic lowland of Costa Rica. This lowland is a subduction basin of tectonic origin, in which thick alluvial and marine sediments are accumulated. At its southwestern side it is bordered by active volcanoes. The climate of the area is hot and humid throughout the year, with a constant mean air temperature of about 25°C and a well-distributed mean annual rainfall of about 3500 to 5500 mm.

Landscape formation was found to be strongly influenced by volcanism. I investigated two particular phenomena: (1) the formation of beach ridges and (2) landscape dynamics in the middle and lower parts of the alluvial plains.

The formation of beach ridges along the Caribbean coast, ranging in age between about 100 and 5000 years, appears to be related to discontinuous sand supply by eruptions of the volcanoes which border the study area. This is indicated by small but consistent and statistically significant differences in chemical composition between sediments of individual beach ridges, which cannot be explained by textural differences. The differences are thought to be caused by variations in magma composition of different eruptions in the volcanic hinterland. SiO_2 contents of the ridge sediments vary between about 52% and 58%, other elements show variations corresponding to magmatic differentiation. Petrographically this is expressed in variations in the amount of andesitic rock fragments and pyroxene grains. Chemical composition of beach ridge sediments is similar in composition to the erupted products, in spite of the removal of part of the mobile elements and mixing with weathered sediments.

Late Holocene sand deposits fill up former river channels and cover adjacent overbanks in the central fluvial plains of the study area. Two facies types are distinguished. The first is an up to several km long, 3 to 10 m deep, and 10 to 80 m wide coarse sandy channel fill facies, which shows an overall fining upward sequence both in grain size and structure. This facies has a pebbly base, small coarse sandy large troughs in the middle part and sandy trough cross-beds in the upper part. The second facies is a wing-shaped, less than 1.5 m thick, sandy overbank facies which shows small cross-beds and ripple-marks and extends laterally up to 300 m from the channel. The deposits testify to episodic supply of huge amounts of loose material on the upper slopes of the volcanoes that border the area, as a consequence of eruptions. Complete choking of the channel and concomitant shifts of the river course appears to occur only during extreme rainfall events. Of a total surface area of about 300 km^2, two smaller areas covered with this kind of channel fill deposits related to eruptions of Turrialba volcano about 2000 yrs BP are described, as well as a similar but much younger deposit related to the 1963-1965 eruptions of Irazú volcano. In the distal part of watersheds, thick (up to 1 m or more) crevasse splay sediments appear to be a less voluminous manifestation

of fluvial sedimentation triggered by volcanic activity. One of such deposits is described, possibly related to the 1864-1866 eruptions of Turrialba volcano. In the fluvial plain short periods of highly active sedimentation and landscape formation are alternated with longer, rather inactive periods.

Soil formation has been studied in relation with time in a Holocene (<5000 yr) soil chronosequence on sandy, andesitic, ocean beach ridges along the Caribbean coast. All soils are under tropical rainforest. Tropopsamments are present on the 3 younger beach ridges and Hapludands are on the older ones. Drainage conditions change by subsidence from excessively drained in the two youngest soils to imperfectly drained in the two oldest soils. The parent materials of all soils are sands with similar mineralogical composition: andesitic rock fragments, plagioclase, and pyroxene dominate, with minor amounts of opaque minerals. None of the parent materials contained >13% (v/v) volcanic glass. It has been found that under these conditions Andisols form within 2000 years. Imperfect drainage caused mottling and accumulation of iron-coatings, as well as the formation of a thin O-horizon in the two oldest profiles. The increase in fine material and the accumulation of organic matter cause an increase of CEC and andic properties, and a decrease in bulk density and pH with soil age. Depth of biological influence increases with soil age, but soil faunal activity is hampered in the oldest three profiles, probably by imperfect drainage. Due to extreme leaching, the sum of exchangeable bases is less than 2 cmol+.kg^{-1} in the B-horizons of the older soils, notwithstanding the presence of a considerable amount of weatherable primary minerals.

Using micromorphological, mineralogical, and chemical analyses, weathering and neoformation of minerals was investigated in the same chronosequence. Weathering and neoformation of minerals with increasing soil age is characterized by: (i) increasing pellicular and linear alteration of sand grains and (ii) decrease of the sand fraction and concomitant increase of finer material. Andesitic rock fragments weather more rapidly than plagioclase and pyroxene mineral grains. The alteration rates of the latter two are similar. Clay content in the about 2000-yr-old soil is several times higher than in soils developed on rhyolitic parent materials of similar age in New Zealand. Formation of allophane with Si/Al ratios ranging from 1.9 to 3.8 takes places mainly in the B horizons. Aluminium-humus complexes, allophane, and Al oxides and hydroxides are mainly formed in the A horizons. Small amounts of gibbsite were noticed in soils older than 2000 yr. Small amounts of 2:1 and 1:1 clay minerals present in the clay fraction of all soils are thought to be inherited from the parent material, which contained sand-sized bodies of clay and andesitic rock fragments with clay pseudomorphs, both consisting of 2:1 and 1:1 clay minerals.

Furthermore, an attempt was made to quantify aspects of soil formation in eight soil profiles developed on volcanic parent material in the area. Assuming Ti to be immobile, gains and losses of major elements in were calculated from total element contents. Initially soil formation (0 to 0.5 ka) in Tropopsamments involves dilation of the sandy deposits by incorporation of organic matter and formation of structure and biopores, without detectable gains or losses of elements. In 2 to 5 ka old sandy Hapludands, primary minerals are still abundant, but up to 20% of the mineral soil consists of X-ray amorphous materials. Dilation continues and losses of Mg, Ca, Na and K, and to a lesser degree Si, have become measurable. In a < 18 ka old, stony Melanudand primary minerals and, especially, volcanic glass, become depleted in the fine-earth fraction. A mixture of short-range order material, metal-humus complexes, gibbsite and kaolin minerals dominate in the A horizon, while gibbsite, halloysite, and short- range order material are the most important secondary minerals at greater depths. Dilation of the soil mass prevails in the A horizon, while collapse has occurred in the B horizon. Considerable amounts of mobile elements are lost: 50 to 85% of Si, Mg, Ca, Na, and K have been leached from the soil profile. The strongly collapsed Haploperox, > 50 to 450 ka old, thought to have formed from Andisols, are devoid of primary minerals (except opaques) in the upper meters, and are dominated by gibbsite, kaolin minerals and iron (hydr)oxides. Under the prevailing environmental conditions, weathering and neoformation of primary volcanic minerals lead to almost complete losses of basic cations, probably in a time period between 20 and 50 ka. Si and P mineral reserves are depleted considerably, but part is still present after such long time periods.

SAMENVATTING

In dit proefschrift wordt de invloed van vulkanisme op landschapsvorming en bodemvormingsprocessen op vulkanisch moedermateriaal bestudeerd in het Atlantische laagland van Costa Rica. Dit laagland is een tektonisch randbekken gevormd als gevolg van de subduktie van de Cocos plaat onder de Carribische plaat en is opgevuld met dikke alluviale en mariene sedimenten. In het zuidwesten wordt het laagland begrensd door verschillende aktieve vulkanen. Het klimaat is gedurende het hele jaar warm en vochtig, en wordt gekenmerkt door een gemiddelde luchttemperatuur van ongeveer 25°C en een gelijkmatig verdeelde jaarlijkse neerslag van 3500 tot 5500 mm.

Landschapsvormende processen in het gebied blijken sterk beinvloed te worden door vulkanisme. In dit proefschrift heb ik twee landschapsvormende verschijnselen in meer detail onderzocht, te weten (1) de vorming van strandwallen, en (2) de dynamiek van het rivierenlandschap in de midden- en benedenloop van rivieren.

De vorming van 100 tot 5000 jaar oude strandwallen langs de Caribische kust lijkt gerelateerd te zijn met onregelmatige zandaanvoer ten gevolge van uitbarstingen van de vulkanen die het studiegebied begrenzen. Aanwijzing hiervoor zijn kleine, maar consequente en statistisch signifikante verschillen in de chemische samenstelling van strandwalsedimenten tussen de verschillende wallen. Deze verschillen blijken niet het gevolg te zijn van verschillen in textuur tussen sedimenten van de verschillende strandwallen. Het lijkt erop dat deze verschillen een gevolg zijn van variaties in de magma samenstelling tussen de verschillende erupties van de vulkanen. Het SiO_2 gehalte van de strandwalsedimenten varieert tussen de 52 en 58%, en andere elementgehalten vertonen variaties die overeenkomen met verschillen t.g.v. magma differentiatie. Petrografisch uiten deze verschillen zich in een varierende hoeveelheid andesitische gesteentefragmentjes en pyroxeen. De chemische samenstelling van strandwalsedimenten vertoont een sterke gelijkenis met die van de eruptieprodukten. Dit ondanks het feit dat een deel van de mobiele elementen verdwenen is uit de strandwalsedimenten door verwering en het mengen van vers eruptiemateriaal met oude sedimenten tijdens transport.

In het centrale deel van het studiegebied hebben laat-Holocene zandafzettingen voormalige rivierbeddingen opgevuld en bedekken ze bovendien aangrenzende voormalige oeverwallen. Twee verschillende facies kunnen worden onderscheiden. De eerste omvat de facies die de voormalige rivierbedding heeft opgevuld, en die verschillende kilometers lang, 3 tot 10 meter diep en 10 tot 80 meter breed kan zijn. De korrelgrootte van de sedimenten van deze facies neemt af van beneden naar boven en ook sedimentaire strukturen worden kleiner in deze richting. De basis van de afzetting is grindrijk, het middendeel bestaat uit zand en vertoont trogvormige scheve gelaagdheid en het bovendeel van de afzettingen is zandig en vertoont scheve gelaagdheid. De tweede facies is een minder dan 1.5 meter dik zandpakket dat is afgezet op voormalige oeverwallen en dat zich minder dan 300 m van de oude bedding uitstrekt. Deze afzettingen wijzen op een onregelmatige aanvoer van grote hoeveelheden zand die kort na erupties aangevoerd worden vanaf de bovenhellingen van de vulkanen die aan het studiegebied grenzen. Het volledig opgevuld raken van een rivierbedding en het als gevolg daarvan gelijktijdig verleggen van van de rivierloop lijkt alleen op te treden tijdens buitengewoon hevige regenval. Van afzettingen die in totaal ongeveer 300 km^2 omvatten en die gerelateerd lijken te zijn aan erupties van de Turrialba vulkaan ongeveer 2000 jaar geleden, zijn twee deelgebieden beschreven in dit proefschrift. Bovendien is een vergelijkbare, maar veel jongere afzetting beschreven die een gevolg is van de erupties van 1963-1965 van de Irazú volkaan. In de benedenloop van rivieren, tot meer dan 1 m dikke sedimentwaaiers ("crevasse-afzettingen") lijken eveneens geinitieerd te worden door vulkanische aktiviteit. Eén zo'n afzetting, mogelijk gerelateerd aan de 1864 - 1866 uitbarstingen van de Turrialba vulkaan, is beschreven in dit proefschrift. In het studiegebied wisselen korte perioden waarin veel sediment wordt afzet en het landschap ingrijpende verandering ondergaat, af met lange perioden waarin het landschap weinig verandert en stabiel lijkt te zijn.

Bodemvorming in de tijd is bestudeerd in een minder dan 5000 jaar oude chronosekwentie op strandwallen bestaande uit andesitisch zand langs de Caribische kust. Alle bestudeerde bodems liggen onder een tropisch regenwoud vegetatie. De bodems op de 3 jongste strandwallen zijn geklassificeerd als Tropopsamments, terwijl de bodems op de oudere strandwallen Hapludands zijn. De drainage varieert onder invloed van tectonische bodemdaling van zeer snel in de twee jongste bodems tot matig gedraineerd in de twee oudste bodems. Het moedermateriaal van alle bodems is zand met een vergelijkbare mineralogische samenstelling: andesitische gesteentefragmentsjes, plagioklaas en pyroxeen domineren, terwijl opake mineralen in mindere mate voorkomen. Geen van de moedermaterialen bevatte meer dan 13% vulkanisch glas (op volumebasis). Het blijkt dat onder deze omstandigheden Andisolen gevormd kunnen worden binnen 2000 jaar. Matige drainage veroorzaakt vlekking en aanrijking van ijzer in de vorm van huidjes, alsook de vorming van de dunne O-horizon in de 2 oudste bodems. Een toename in fijn materiaal en in het organische stof gehalte veroorzaken een toename van de kationen uitwisselcapaciteit en andische eigenschappen, en een afname in bulkdichtheid en pH met het ouder worden van de bodem. De diepte tot waarop sporen van dierlijke aktiviteit worden waargenomen neemt toe met stijgende bodemouderdom. In de twee oudste bodems veroorzaakt de matige drainage echter dat deze aktiviteit geremd wordt in de diepere lagen. Als gevolg van de extreme uitloging van deze bodems door regenwater is de som van uitwisselbare kationen minder dan 2 $cmol+.kg^{-1}$ in de B horizon van de oudere bodems. Dit ondanks het feit dat deze horizonten nog aanzienlijke hoeveelheden verweerbare primare mineralen bevatten.

Gebruik makend van mikromorfologische, mineralogische en chemische analyse technieken zijn in deze chronosekwentie ook verwering en nieuwvorming van mineralen onderzocht. Met het toenemen van de leeftijd van de bodems zijn verwering en nieuwvorming van mineralen gekenmerkt door: (1) een toename in de mate waarin zandkorrels een "gerafeld" uiterlijk hebben door verwering van de randen en (2) een afname in de hoeveelheid zand en een gelijktijdige toename van het kleigehalte. Andesitische gesteentefragmentjes blijken sneller te verweren dan plagioklaas en pyroxeen mineralen. De snelheid waarmee deze laatste twee worden aangetast door verwering is ongeveer gelijk. Het kleigehalte in een ongeveer 2000 jaar oude bodem in deze chronosekwentie is verschillende malen hoger dan dat in ongeveer even oude bodems ontwikkeld op rhyolitisch moedermateriaal in Nieuw Zeeland. Allofaan vorming vindt vooral plaats in de B horizont, en de Si/Al verhouding van deze allofaan varieert van 1.9 tot 3.8. Complexen van Al met organische stof, allofaan, Al-oxiden en -hydoxiden worden gevormd in de A horizonten, terwijl kleine hoeveelheden gibbsiet aanwezig zijn in bodems ouder dan 2000 jaar. Kleine hoeveelheden 2:1 en 1:1 zijn aanwezig in de kleifraktie van alle bodems. Waarschijnlijk zijn deze mineralen afkomstig van het moedermateriaal, waarin bolletjes bestaande uit deze mineralen aanwezig zijn in de zandfraktie. Tevens bevatten de andesitische gesteentefragmentjes vaak insluitsels bestaande uit deze kleimineralen.

In dit proefschrift is ook geprobeerd om bepaalde aspekten van bodemvorming te kwantificeren in acht verschillende bodems op vulkanische moedermaterialen. Onder de aanname dat het element Ti immobiel is, zijn verliezen en aanrijkingen van de hoofdelementen in de bodems berekend uit gemeten gehalten van deze elementen. Bodemvorming in een minder dan 500 jaar oude Tropopsamment omvat vooral volume verdunning (uitzetting) van het zandige moedermateriaal door aanrijking van organische stof en de vorming van struktuur en bioporien. Aanrijkingen of verliezen van elementen werden niet gemeten. 2000 tot 5000 jaar oude Hapludands bevatten nog steeds grote hoeveelheden primare mineralen, maar tot 20% van het bodemmateriaal bestaat reeds uit niet kristallijn materiaal. Volume verdunning (uitzetting?) is verder gevorderd en verliezen van Mg, Ca, Na en K, en in minder duidelijk ook van Si, worden meetbaar. De hoeveelheid primare mineralen, en met name vulkanisch glas, raakt uitgeput in de fijne fraktie van een minder dan 18000 jaar oude, stenige Melanudand. Een mengsel van niet-kristallijne mineralen, complexen van organische stof en Al, gibbsiet en kaoliniet/halloysiet zijn het belangrijkste in de fijne fraktie in de A horizont, terwijl gibbsiet, halloysiet en niet-kristallijne mineralen het belangrijkste zijn in de diepere lagen. In de A horizont is de grondmassa uitgezet, maar de B horizont is ingestort onder het gewicht van de bovenliggende bodemlagen. Aanzienlijke hoeveelheden mobiele elementen zijn verdwenen uit de bodemmassa: 50 to 85% van de oorspronkelijk aanwezige hoeveelheden Si, Mg, Ca, Na, en K zijn verloren gegaan door uitspoeling. De grondmassa van oudere Haploperox is ingestort over de hele profieldiepte. Deze bodems worden verondersteld te zijn gevormd uit Andisolen en varieren in leeftijd tussen >50000 en 450000 jaar oud. Behalve een geringe hoeveelheid opake mineralen zijn alle primaire mineralen verweerd. Mineralogie van deze oude bodems wordt gedomineerd door gibbsiet, kaoliniet en halloysiet, en ijzeroxiden en -hydroxiden. Onder de heersende omgevingsfactoren leiden verwerings- en nieuwvormingsprocessen in vulkanisch moedermateriaal tot uitspoeling van bijna alle oorspronkelijk aanwezige basische kationen. Ook minerale reserves van Si en P raken sterk uitgeput, maar zelfs na dergelijk lange tijdsperioden is nog altijd een deel aanwezig.

NASCHRIFT

Eind 1986 vertrok ik enigszins morrend naar Costa Rica. Een stage in Colombia was niet doorgegaan wegens de onveilige situatie in het dorp waar ik heen zou gaan, en andere mogelijkheden waren er eigenlijk niet. Tot mijn verbazing bleek het werken op een LU steunpunt lang niet zo beklemmend als ik had gevreesd en was het Atlantische laagland een fascinerend gebied. Toen ik eind 1988 de mogelijkheid kreeg om als AIO verder onderzoek te doen in dit gebied, greep ik deze kans dan oók met beide handen aan.

Nu, ruim 7 jaar later, ligt er een proefschrift voor U met daarin een deel van de resultaten van het onderzoek dat ik gedurende dit AIO-schap heb verricht. Veel mensen hebben mij bij de uitvoering ervan geholpen, en wil hierbij bedanken.

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Wim Wielemaker heeft een grote invloed gehad op mijn vorming in de veldbodemkunde, en het was tijdens veldtrips met hem dat we voor het eerst op het idee kwamen dat "er wel een leuk promotieonderzoek zat in het bestuderen van de bodemvorming in het gebied".

Toine Jongmans heeft aan het grootste deel van het bodemkundige onderzoek in dit proefschrift aktief bijgedragen heeft zijn co-auteurschap in enkele hoofdstukken meer dan verdiend. Daarnaast was en is hij mijn steun en toeverlaat bij het regelen van Wageningse perikelen.

Jetse Stoorvogel heb ik als student leren kennen in het destijds nog volop in ontginning zijnde noordoosten van het Atlantiche laagland. Tot mijn verrassing kwam hij in 1990 als collega op het steunpunt werken. Sindsdien hebben we vele stimulerende gesprekken over bodemkundige en andere zaken achter de rug, en was nooit te beroerd om me in te wijden in wat voor mij komputer"geheimen" waren.

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CURRICULUM VITAE

André Nieuwenhuyse werd op 2 november 1962 geboren te 's-Heer Hendrikskinderen. In september 1981 begon met de studie Bodemkunde en Bemestingsleer aan de Landbouwuniversiteit Wageningen. Na het voltooien van afstudeervakken in de regionale bodemkunde, bodemkunde en plantenvoeding, geomorfologie en tropische bodemkunde, studeerde hij in september 1988 af. Vanaf november 1988 tot augustus 1993 was hij werkzaam als AIO op het steunpunt van de Landbouwuniversiteit in Costa Rica aan het onderzoek waarvan de resultaten vermeld staan in deze thesis. Sindsdien heeft hij zich in Costa Rica beziggehouden met consultancies, het uitvoeren van een bodemkartering van de Costaricaanse provincie Guanacaste en de praktische veehouderij.