JOS BAKKER

TECTONIC AND CLIMATIC CONTROLS ON LATE QUATERNARY SEDIMENTARY PROCESSES IN A NEOTECTONIC INTRAMONTANE BASIN

(The Pitalito Basin, South Colombia)

CENTRALE LANDBOUWCATALOGUS

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Cover (thesis edition)

The cover symbolizes the multi-disciplinary approach of this study. The irregular, title-bearing brown box at the top illustrates the basement morphology of the Pitalito Basin going from west to east (geology). The two block diagrams in the middle part of the cover illustrate the two different types of alluvial architecture going from west to east (geomorphology). The sides of the front cover shows a hypothetical general pollen diagram (palynology). At the back cover a part of the lithological column of section 20A is shown (sedimentology).

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LANDBOUWUNIVERSITEIT
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Author's address

J.G.M. Bakker
Department of Soil Science and Geology
Agricultural University of Wageningen
P.O. Box 37
6700 AA Wageningen
The Netherlands

Promotoren: Dr. S.B. Kroonenberg, hoogleraar in de geologie
Dr. T. van der Hammen, emeritus hoogleraar in de palynologie aan de Universiteit van Amsterdam (UvA)

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JOS BAKKER

TECTONIC AND CLIMATIC CONTROLS ON LATE QUATERNARY SEDIMENTARY PROCESSES IN A NEOTECTONIC INTRAMONTANE BASIN

(The Pitalito Basin, South Colombia)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD VAN
DOCTOR IN DE LANDBOUW- EN MILIEUWETENSCHAPPEN,
OP GEZAG VAN DE RECTOR MAGNIFICUS,
DR. H.C. VAN DER PLAS,
IN HET OPENBAAR TE VERDEDIGEN OP
WOENSDAG 20 JUNI 1990
DES NAMIDDAGS TE VIER UUR
IN DE AULA VAN DE LANDBOUWUNIVERSITEIT
TE WAGENINGEN

ISN= 521784

Ter herinnering aan mijn vader
Aan mijn moeder

NW02201,1364

ERRATA

Page 37, Photo 7: ...In Figure 23... *must be* ...in Figure 26....

Page 72, Fig. 23:... Log 88 and Log 90... *must be* .... PIT88 and PIT 90...

Figure 22 and 23 (page 71 and 72, respectively) have been placed erroneously after Figure 24 (page 70).

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STELLINGEN

*

Het belang van strike-slip bewegingen bij het ontstaan van intramontane bekkens in de Colombiaanse Andes is onvoldoende onderkend.

Dit proefschrift.

*

Het feit dat Butler (1983) geen strike-slip bewegingen kan aantonen in het Boven Magdalena dal kan worden toegeschreven aan de oriëntatie van zijn secties.

Butler, R.E. (1983) Andean-type foreland deformation: Structural development of the Neiva Basin, Upper Magdalena Valley, Colombia. *Ph.D. Thesis, University of South Carolina*, 272 p.

*

Gelet op de verspreiding van anastomoserende rivierpatronen over een breed scala van klimaatsregimes is niet het klimaat maar vermoedelijk de geringe gradiënt de bepalende faktor voor de aanwezigheid van dergelijke riviersystemen.

Dit proefschrift.

Smith, D.G. (1983) Anastomosed fluvial deposits: modern examples from Western Canada. In: *Modern and Ancient Fluvial Systems*, (Ed. by J.D. Collinson & J. Lewin), *Spec. Publ. int. Ass. Sediment.* **6**, p. 155-168.

*

Het gebrek aan inzicht of bewijsmateriaal omtrent de invloed van autochtone dan wel allochtone faktoren op de lithologische omslagen in een sedimentpakket kunnen tot de onjuiste konklusie leiden dat één van deze faktoren van ondergeschikt belang is geweest bij een dergelijke omslag.

*

De veen/kool index hoeft geen indikator van hoge temperaturen of hoge regenval te zijn.

Dit proefschrift

McCabe, P.J. (1984) Depositional environments of coal and coal-bearing strata. In: *Sedimentology of Coal and Coal-bearing Sequences*, (Ed. by R.A. Rahmani & R.M. Flores), *Spec. Publ. int. Ass. Sediment.* **7**, p. 13-43.

*

De temperatuursgradiënt in het tropisch Andiene gebied was tijdens het Laat Glaciaal slechts in geringe mate steiler dan tegenwoordig.

Dit proefschrift

*

De schijnbare tegenspraak tussen het gehomogeniseerde karakter van Andosolen en de aanwezigheid van een gedetailleerde pollenstratigrafie daarin dient nader bestudeerd te worden.

Salomons, J.B. (1986) Paleoecology of volcanic soils in the Colombian Cordillera Central (Parque Nacional Natural de los Nevados). *Ph.D. Thesis, University of Amsterdam*. Also published as *Dissertationes Botanicae* **95**, 212 p.

Bakker, J.G.M. & Salomons, J.B. (1989) A palaeoecological record of a volcanic soil sequence in the Nevado del Ruiz area, Colombia. *Rev. of Palaeobot. Palynol.* **60**, p. 149-163.

*

Aangezien de pollenconcentratie nauw samenhangt met de sedimentatiesnelheid is het sterk af te raden de fluctuaties in de pollenconcentratie te gebruiken als indicator voor een veranderend vegetatiepatroon.

Roth, L. & Lorscheitter, M.L. (1989) Palynology of a peat in Parque Nacional de Aparados da Serr, Río Grande do Sul, Brazil. International Symposium on Global Changes in South America during the Quaternary: Past-Present-Future, São Paulo (Brazil), May 8-12, 1989.

*

Bij het vaststellen van referentiewaarden voor zware metalen in de Nederlandse bodem dient ook rekening te worden gehouden met de samenstelling van de geologische ondergrond.

*

'Platteland, dat is wat er langs komt als je van de ene stad naar de andere reist.'

'Het verleden is het enige wat we hebben, daar staan we bovenop, iedere dag een stukje hoger.'

Jules Deelder

*

Gezien de geringe salariëring van de AIO's en OIO's dient reklame in proefschriften te worden toegestaan.

*

'Die Wahrheit ist das kostbarste aller Güter und soll gehandhabt werden mit Sparsamkeit und Zurückhaltung.'

In: Nächstes Jahr in Jerusalem door André Kaminski

Stellingen behorende bij het proefschrift van Jos Bakker:
Tectonic and climatic controls on Late Quaternary sedimentary processes in a neotectonic intramontane basin (The Pitalito Basin, Colombia)
Wageningen, 20 juni 1990.

This study was carried out at the Department of Soil Science and Geology,
Wageningen Agricultural University, The Netherlands.

The palynological analysis was carried out at the Hugo de Vries Laboratory,
Department of Palynology and Paleo/Actuo-Ecology,
University of Amsterdam, The Netherlands.

CURRICULUM VITAE

Jos Bakker werd geboren op 11 mei 1957 te Haarlem. In 1973 behaalde hij het MAVO diploma aan de Jeroen Mavo te Haarlem. In 1975 werd het HAVO diploma behaald en in 1977 het Atheneum-B diploma, beide aan het Mendel College te Haarlem. Hij studeerde vanaf 1977 biologie aan de Universiteit van Amsterdam. In 1980 werd het kandidaatsexamen Algemene Biologie gehaald (B1) en in 1981 het kandidaatsexamen Geologische Biologie (B5). Tijdens zijn doktoraalexamen werd een tweetal hoofdvakken en een tweetal bijvakken gevolgd, te weten:

- Een eerste hoofdvak in de palynologie. Tijdens dit onderzoek werd een vulkanische bodem afkomstig uit Colombia palynologisch geanalyseerd.
- Een tweede (extra) hoofvak in de geomorfologie werd uitgevoerd in het noordelijk deel van de hoogvlakte van Bogotá in Colombia.
- Een bijvak geomorfologie. Tijdens dit bijvak werd een post-vergletsjerd gebied in Vorarlberg (Oostenrijk) in kaart gebracht.
- Een tweede bijvak is gevolgd aan de Vrije Universiteit van Amsterdam waarbij delen uit de kursus 'Laaglandgenese' werden gevolgd.

In januari 1985 studeerde hij af als doctorandus in de biologie. Vanaf februari 1985 tot mei 1988 was hij als onderzoeksassistent werkzaam aan de Vakgroep Bodemkunde & Geologie van de Landbouwuniversiteit te Wageningen.

CONTENTS

ABSTRACT

The present study deals with the influence of tectonics and climatic changes on sedimentation patterns in the Quaternary Pitalito Basin (lat. 1°52'N, long. 76°02'W). This intramontane sedimentary basin is 15 km in width and 20 km in length and is located in the Eastern Cordillera of the southern Colombian Andes at an altitude of c. 1300 m.

Chapter 1 discusses the scope of this study.

In Chapter 2 general information is given about the regional tectonic setting of the basin, its morphological features, the present climate around the basin and the present-day zonal vegetation of the Colombian Andean region.

Chapter 3 starts with the description of a gravity survey carried out to examine basin geometry. The basin consists of a shallow western part in which the basement is situated at 300-400 m depth, and a deep eastern part with the basement at 1000-1200 m depth. This geometry is controlled by tectonic structures: (1) an active strike-slip fault with right lateral displacement along the northern boundary of the basin, (2) a relatively passive southern fault system and (3) a NW/SE-oriented fault which separates the shallow western part from the deep eastern part and which is situated at the present course of the Guarapas river. By extrapolation from the known sedimentation rates for the last 60,000 years, subsidence could have started c. 4.5 Ma ago. The second part of Chapter 3 discusses the geoelectric characteristics of the upper 200 m of the non-exposed sedimentary infill. Coarse to medium clastics (cobbles, gravel and sand) are restricted to the shallow western part, whereas clay and peat predominate in the deep eastern part. The lateral transition between the two lithofacies is situated along the present course of the Guarapas river.

In Chapter 4 the near-surface and surface basin sediments are described. These sediments have been studied in exposures and borings. They represent the last stage of sedimentary infill by a northeastward-heading fluvial system. The top of these sediments has an approximate age of 17,000 years B.P. The sediments in the shallow western region of the basin represent the proximal part of the fluvial system. They consist of widely scattered lenticular channel deposits enveloped by extensive and thick inorganic overbank deposits. The sediments in the east form the distal part of the fluvial system. They are made up by laterally fixed sandbodies whereas the adjacent flood basin deposits are characterized by the occurrence of organic strata which are several meters thick. These different types of alluvial architecture are ascribed to different fluvial dynamics. The western, proximal component is intermediate between braided and anastomosing river types and is characterized by channels migrating laterally due to avulsion. The distal, eastern part shows an anastomosing pattern and is characterized by laterally fixed channels. The transition between the two river types is equally situated along the line of the Guarapas river.

In Chapter 5 two pollen records are described which were obtained from the organic-rich deposits in the eastern part of the basin. The records revealed the paleoecological and especially climatic conditions during the last 60,000 years around the Pitalito Basin. From ≈60,000 years B.P. to ≈20,000 years B.P. mean annual temperature fluctuated considerably and decreased, in comparison with modern temperatures, with about 3°C during the relatively warm periods (interstadials) to about 6°-8°C during the coldest periods (stadials). There is no evidence for a significant variation in climatic humidity during the registered period. The changing temperatures led to a downward displacement of the vegetation belts amounting c. 500 m during the interstadials and c. 1500 m during stadial times, in comparison with the present position. In spite of these environmental changes, the fluvial system present in the Pitalito Basin probably did not change significantly until c. 20,000 years B.P. At that time the eastern flood basins became

choked with clastic sediments and peat growth ended. Sedimentological and palynological data suggest a change from relatively humid climatic conditions to a semi-arid climate which is characterized by torrential rains and a sparse vegetation cover. These semi-arid conditions lasted from ≈20,000 to at least ≈17,000 years B.P. Somewhere between 17,000 and 7000 years B.P. basin infill came to an end and the rivers incised in their own sediments. The river changed its course 90° to an outlet in the NW. There is no palynological record of this interval. Around 7000 years B.P. peat began to develop again in the northeastern part of the basin due to tilting of the basin floor. The palynological record registers the prevalence of present-day climatic conditions from 7000 years B.P. onward (mean annual temperature c. 20°C; annual rainfall c. 1200 mm). A somewhat warmer period is recorded around 5000 years B.P.; mean annual temperature was about 2°-3°C higher than today.

In Chapter 6 a synthesis of the data of the former chapters is given.

The following conclusions may be drawn:
- The intramontane Pitalito Basin developed as a result of extension along a fault wedge which forms part of the Garzón-Suaza fault. It is the first extensively described pull-apart basin in the Colombian Andean region. Tectonics played a decisive role in the differentation of the fluvial system
- The effects of climatic changes during the depositional history are superimposed upon those caused by tectonic activity. During a considerable part of the Last Glacial (60,000-20,000 years B.P.) the fluvial system in the Pitalito Basin was not affectedby the vertical shifts of the zonal vegetation belts. The start of especially dry climatic conditions around 20,000 years B.P., possibly combined with tectonic activity, caused a dramatic change in the sedimentation patterns. The maximal estimated temperature decline during the Pleniglacial at the altitude of the Pitalito Basin is possibly in the order of 6°-8°C compared to modern temperatures.
- The thick pile of sedimentary infill and the presence of well-preserved pollen at this elevation make the Pitalito Basin a very suitable site for deep borings to record Late Cenozoic vegetational changes at low elevation.

Parts of this thesis are or will be published as papers with the following titles:

Bakker, J.G.M. (1989) Neotectonic control on sedimentology of Pleistocene fluvial deposits in the intramontane Pitalito Basin, Colombia. Int. Symposium on global changes in South America during the Quaternary, 8-12 May, São Paulo, Brazil, p. 145-150.

Bakker, J.G.M. (1989) Historia de la tectónica y de la sedimentación en una cuenca Cenozoica intramontana (Cuenca de Pitalito). *Revista Colombia Geográfica* XV/2.

Bakker, J.G.M. (in prep.) Late Pleistocene vegetational and climatic history of the Pitalito Basin, Eastern Cordillera, Colombia.

Bakker, J.G.M., Kleinendorst, Th. and Geirnaerts, W.A. (in prep.) Tectonic and sedimentary history of a Late Cenozoic intramontane basin (Pitalito Basin, Colombia).

SAMENVATTING

In dit proefschrift wordt de invloed beschreven van tektoniek en klimatologische veranderingen op sedimentatiepatronen in het Kwartaire Pitalito Bekken (1°52' noorderbreedte, 76°02' westerlengte). Dit intramontaan sedimentair bekken is 15x20 km^2 en bevindt zich in de Oostelijke Cordillera in het zuidelijke deel van de Colombiaanse Andes op een hoogte van ongeveer 1300 m.

In hoofdstuk 1 wordt het doel van de studie beschreven.

In hoofdstuk 2 wordt algemene informatie verstrekt over: de regionale structureel geologische omgeving van het bekken, de morfologische verschijnselen binnen het bekken, het huidige klimaat ter hoogte van het bekken en de huidige zonale vegetatie van het Colombiaanse Andiene gebied.

In hoofdstuk 3 wordt allereerst het gravimetrisch onderzoek besproken dat werd uitgevoerd om de vorm van de bekkenondergrond te bestuderen. Het bekken bestaat uit een ondiep westelijk deel met een diepte van 300-400 m, en een diep oostelijke deel dat ongeveer 1000-1200 m diep is. Deze vorm wordt bepaald door de volgende tektonische strukturen: (1) een aktieve breuk met een dextrale 'strike-slip' bewegingskomponent langs de noordelijke bekkenrand, (2) een relatief passief breuksysteem langs de zuidrand van het bekken en (3) een NW/ZO-gerichte breuk die het ondiepe westelijke deel van het diepe oostelijke deel scheidt. Deze valt samen met de huidige loop van de Guarapas rivier. Op grond van extrapolatie van de sedimentatiesnelheden berekend over de laatse 60.000 jaar, zou de bekkendaling ongeveer 4,5 miljoen jaar geleden aangevangen kunnen zijn. In het tweede deel van hoofdstuk 3 worden de geoelectrische eigenschappen besproken van de bovenste 200 meter van de niet-ontsloten bekkenopvulling. Grof- tot middenkorrelige sedimenten (keien, grind en zand) zijn voornamelijk in het ondiepe westelijke deel aanwezig, terwijl klei en veen in het diepe oostelijke deel overheersen. De overgang tussen deze twee lithofacies ligt ter hoogte van de Guarapas rivier.

In hoofdstuk 4 worden de sedimenten beschreven die aan het oppervlak van het bekken liggen. Deze sedimenten konden aan de hand van boringen en ontsluitingen worden bestudeerd. Zij vertegenwoordigen de laatste fase van bekkenopvulling door een noordoostwaarts-stromend rivier systeem. De top van deze sedimenten heeft een ouderdom van ongeveer 17.000 jaar. De proximale sedimenten in het ondiepe westelijke deel van het bekken, bestaan uit relatief brede, dunne, lensvormige geulafzettingen die zijn ingebed in dikke pakketten komkleien. Deze afzettingen hebben een grote laterale verbreiding. De distale sedimenten in het oostelijk gebied van het bekken bestaan uit relatief smalle, dikke zandlichamen en de aangrenzende overstromingsgebieden worden gekenmerkt door de aanwezigheid van metersdikke organische afzettingen. De veranderingen in de rivierpatronen worden toegeschreven aan verschillen in de fluviatiele dynamiek. De westelijke, proximale komponent vertoont een tussenvorm van een vlechtend en anastomoserend rivierpatroon en wordt gekenmerkt door migrerende geulen door de afwezigheid van goed ontwikkelde oeverwallen. De distale, oostelijke komponent van het riviersysteem vertoont een anastomoserend patroon van een groot aantal met elkaar verbonden geulen die, door de aanwezigheid van goed ontwikkelde oeverwallen, zijn gestabiliseerd. De overgang tussen beide riviertypen ligt eveneens ter hoogte van de Guarapas rivier.

In hoofdstuk 5 worden de resultaten besproken van een palynologisch onderzoek van twee boorkernen die afkomstig zijn van de organisch rijke afzettingen in het oostelijk deel van het bekken. Aan de hand van de pollen inhoud is een rekonstruktie gemaakt van de paleoecologische en vooral klimatologische omstandigheden rond het Pitalito Bekken gedurende de laatste 60.000 jaar. Tussen de 60.000 en 20.000 jaar geleden varieerde de gemiddelde jaartemperatuur ter hoogte van het bekken sterk en was lager in vergelijking tot de huidige jaartemperatuur: Gedurende de relatief warme perioden (interstadialen) was de temperatuur ongeveer 3°C lager terwijl dit verschil toenam tot 6°-8°C gedurende de koudste perioden (stadialen). Er zijn geen aanwijzingen gevonden voor grote verschillen in de vochtigheid van het klimaat gedurende deze periode. De wisselende temperaturen hebben tot neerwaartse verschuivingen geleid van de vegetatiegordels: Tijdens de interstadialen lagen de vegetatiegordels ongeveer 500 m lager in vergelijking met hun huidige positie, terwijl de gordels gedurende de stadialen ongeveer 1500 m daalden. Ondanks deze variaties lijkt het erop dat het fluviatiele systeem in het Pitalito Bekken geen significante veranderingen onderging tot ongeveer 20.000 jaar geleden. Rond die tijd werd in de oostelijke komgebieden veel klastisch materiaal afgezet waardoor de veengroei ophield. Zowel de sedimentologische als palynologische gegevens duiden op een verandering van een relatief humide klimaat naar een semi-aride klimaat dat gekenmerkt wordt door hevige stortbuien en een geringe vegetatiebedekking. Dit semi-aride klimaat heerste van ongeveer 20.000 jaar tot ten minste 17.000 jaar geleden. De bekkeninvulling stopte ergens tussen 17.000 en 7000 jaar geleden waarna de rivieren hun eigen sedimenten versneden. De rivierloop verlegde zich 90° en stroomt sindsdien naar het huidige NW afvoerpunt. Van deze periode ontbreken palynologische gegevens. Rond 7000 jaar geleden begon de veenontwikkeling opnieuw in het noordoostelijke deel van het bekken ten gevolge van kanteling van de bekkenondergrond. Uit de pollen inhoud van dit organisch materiaal blijkt dat de klimaatsomstandigheden ongeveer 7000 jaar geleden vergelijkbaar waren met de huidige (gemiddelde jaartemperatuur c. 20°C; jaarlijkse neerslag c. 1200 mm). Rond 5000 jaar geleden was de temperatuur ongeveer 2-3°C hoger dan tegenwoordig.

In hoofdstuk 6 volgt de synthese van de data die in de voorafgaande hoofdstukken zijn besproken.

De volgende konklusies kunnen worden getrokken:

- Het intramontane Pitalito Bekken heeft zich ontwikkeld ter hoogte van een breuksplitsing die deel uitmaakt van de Garzón-Suaza breuk. Langs deze breuksplitsing trad divergentie op ten gevolge van laterale bewegingen. Het Pitalito Bekken is het eerste 'pull-apart' bekken in het Colombiaanse Andiene gebied dat uitvoerig beschreven wordt. In het Pitalito Bekken draagt tektoniek in belangrijke mate bij tot de differentiatie van het fluviatiele systeem.
- Het effekt van klimaatsveranderingen op de sedimenten is gesuperponeerd op de effekten veroorzaakt door tektoniek. Ondanks de aanzienlijke verschuivingen van de zonale vegetatie blijft het fluviatiele systeem in het Pitalito Bekken gedurende een groot deel van het Laatste Glaciaal (60.000-20.000 jaar geleden) onveranderd. Pas ongeveer 20.000 jaar geleden heeft vooral de toenemende droogte, mogelijkerwijs in kombinatie met tektoniek, geleid tot grote veranderingen in het sedimentatiepatroon van het Pitalito Bekken. De temperatuur ter hoogte van het Pitalito Bekken was gedurende het Pleniglaciaal maximaal 8°C lager dan tegenwoordig.
- Het dikke sedimentpakket en de voor deze hoogte goed bewaarde pollen maken het Pitalito Bekken zeer geschikt voor de registratie in diepe boringen van veranderingen in de vegetatie gedurende het Laat Cenozoicum op geringe hoogte.

Delen van dit proefschrift zijn of zullen worden gepubliceerd als artikelen met de volgende titels:

Bakker, J.G.M. (1989) Neotectonic control on sedimentology of Pleistocene fluvial deposits in the intramontane Pitalito Basin, Colombia. Int. Symposium on global changes in South Ameria during the Quaternary, 8-12 May, São Paulo, Brazil, p. *145-150.*

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Bakker, J.G.M., Kleinendorst, Th. and Geirnaerts, W.A. (in prep.) Tectonic and sedimentary history of a Late Cenozoic intramontane basin (Pitalito Basin, Colombia).

RESUMEN

El presente estudio trata de la influencia de la tectónica y de los cambios climáticos en los patrones de sedimentación en la cuenca Cuaternaria de Pitalito (lat. 1°52'N, long. 76°02'W). Esta cuenca intramontana sedimentaria de 20 Km de largo y 15 Km de ancho se encuentra en la parte sur de la cordillera Oriental en los Andes colombianos, a una altitud de aproximadamente 1300 m.

En el Capítulo 1 se discute el enfoque del estudio.

En el Capítulo 2 se suministra información general acerca del contexto tectónico regional de la cuenca, sus características morfológicas, su clima actual y la vegetación zonal actual de la region andina colombiana.

El Capítulo 3 se inicia con la descripción del estudio gravimétrico que se llevó a cabo con el objeto de analizar la geometría de la cuenca. Esta consta de una parte somera occidental en la cual el basamento se encuentra a una profundidad de 300-400 m y de una parte oriental más profunda en la cual el basamento está a 1000-1200 m. La geometría está controlada por las siguientes estructuras tectónicas: (1) una falla transcurrente activa con desplazamiento dextrolateral a lo largo del límite norte de la cuenca, (2) un sistema de fallas relativamente pasivas en el sur y (3) una falla de orientación NW/SE que separa la parte somera en el oeste de la parte profunda oriental y se sitúa en el curso actual del río Guarapas. Mediante extrapolación de las tasas de sedimentación conocidas para los últimos 60.000 años se puede deducir que el proceso de subsidencia pudo haber empezado hace cerca de 4.5 millones de años. En la segunda parte del capítulo 3 se discuten las características geoeléctricas de los 200 m superiores del relleno sedimentario no expuesto. Los clastos de textura gruesa a media (cantos, gravillas y arenas) se encuentran restringidos a la parte somera oeste, mientras que las arcillas y turbas predominan en la parte profunda este. La transición lateral entre las dos facies litológicas se sitúa a lo largo del curso actual del río Guarapas.

En el Capítulo 4 se describen los sedimentos superficiales y subsuperficiales, los cuales se estudiaron en afloramientos y mediante sondeos. Estos sedimentos representan el último estadio de relleno sedimentario de la cuenca mediante un sistema fluvial de orientación noreste. El tope de estos sedimentos tiene una edad aproximada de 17,000 años A.P. Los sedimentos en la región somera de la cuenca representan la parte proximal del sistema fluvial. Constan de depósitos dispersos lenticulares de canal, envueltos por espesos depósitos inorgánicos de desborde. Los sedimentos en el este conforman la parte distal del sistema fluvial. Estan hechos de cuerpos arenosos fijos lateralmente, mientras que los depósitos adyacentes de la llanura de inundación se caracterizan por la presencia de capas orgánicas con un espesor de varios metros. Estos diferentes tipos de arquitectura aluvial tienen su orígen en diferencias en la dinámica fluvial. El componente occidental proximal es intermedio entre el tipo de ríos trenzados y anastomosados y se caracteriza por la presencia de canales con migración lateral debido a procesos de avulsión. La parte distal este muestra un patrón anastomosado y se caracteriza por la presencia de canales fijos lateralmente. La transición entre los dos tipos de ríos se sitúa igualmente a lo largo de la línea del río Guarapas.

En el Capítulo 5 se describen dos secuencias de polen obtenidas de los depósitos ricos en materia orgánica de la parte este de la cuenca. Estos registros de polen revelan las condiciones paleoecológicas y especialmente paleoclimáticas durante los últimos 60.000 años en la cuenca de Pitalito y sus alrededores. Entre 60.000 y 20.000 años A.P. la temperatura media anual fluctuó considerablemente, siendo más baja en general, en comparación con la temperatura de los tiempos modernos. La temperatura bajó unos 3°C durante los períodos relativamente calientes (interestadiales) hasta cerca de 6°-8°C durante los períodos más fríos (estadiales). No hay evidencias de variaciones significativas en la

humedad climática durante el período registrado. Los cambios de temperatura llevaron al desplazamiento hacia abajo de los cinturones de vegetación. Este descenso altitudinal fue de cerca de 500 m (tomando la posición actual como referencia) durante los interestadiales y de cerca de 1500 m durante los estadiales. A pesar de estos cambios ambientales, el sistema fluvial existente en la cuenca de Pitalito probablemente no cambió significativamente hasta hace 20.000 años A.P. Durante esta época las llanuras de inundación del este se vieron obstruidas por sedimentos clásticos y la formación de turbas llegó a su fin. Los datos sedimentológicos y palinológicos sugieren un cambio de condiciones climáticas relativamente húmedas hacia un clima semi árido que se caracterizó por la presencia de lluvias torrenciales y por una cobertura vegetal escasa. Estas condiciones de semi-aridez se mantuvieron desde 20.000 años hasta al menos 17.000 años A.P. Entre 17.000 y 7.000 años A.P. el relleno de la cuenca llegó a su fin, y los ríos se encajonaron en sus propios sedimentos. El río que llenaba la cuenca cambió su curso en 90° con el desagüe hacia el noroeste. No hay registros palinológicos de este intervalo. Alrededor de 7.000 años A.P. el desarrollo de las turbas en la parte noreste de la cuenca empezó de nuevo, debido a un basculamiento. Los registros palinológicos muestran la dominancia de las condiciones climáticas actuales a partir de 7000 años A.P. (temperatura media anual c. 20°C, precipitación media mensual c. 1200 mm). Se registraron indicios de la existencia de un período un poco más caliente hacia 5.000 años A.P. con una temperatura media anual 2°-3°C más alta que la de hoy.

En el Capítulo 6 se presenta la síntesis de los datos de los capítulos anteriores.

Se pueden esbozar las siguientes conclusiones:

- La cuenca intramontana de Pitalito se desarrolló como consecuencia de un régimen extensional en una cuña creada por la intersección de dos fallas que forman parte de la falla Garzón-Suaza. Esta es la primera cuenca de tracción descrita para la región andina colombiana. La tectónica jugó un papel primordial en la diferenciación del sistema fluvial.
- Los efectos de los cambios climáticos durante la historia de la depositación sedimentaria se sobreponen a aquellos causados por la actividad tectónica. Durante una parte considerable del Ultimo Glacial (60.000-20.000 años A.P.) el sistema fluvial de la cuenca de Pitalito no se vió afectado por los desplazamientos verticales de los cinturones de vegetación zonal. El inicio de condiciones particularmente secas hacia 20.000 años A.P., posiblemente en combinación con la actividad tectónica produjó un cambio dramático en los patrones de sedimentación. El descenso máximo estimado de la temperatura durante el Pleniglacial a la altura de Pitalito es posiblemente del orden de 6°-8°C en comparación con las temperaturas modernas.
- El espeso relleno sedimentario y la presencia de polen bien preservado, hacen de la cuenca de Pitalito una localidad muy favorable para realizar sondeos profundos con el fin de estudiar los cambios de la vegetación durante el Cenozoico Tardío a esta altitud.

Algunos apartes de esta tesis se han publicado o serán publicados en los siguientes artículos:

Bakker, J.G.M. (1989) Neotectonic control on sedimentology of Pleistocene fluvial deposits in the intramontane Pitalito Basin, Colombia. Int. Symposium on global changes in South Ameria during the Quaternary, 8-12 May, São Paulo, Brazil, p. 145-150.

Bakker, J.G.M. (1989) Historia de la tectónica y de la sedimentación en una cuenca Cenozoica intramontana (Cuenca de Pitalito). *Revista Colombia Geográfica* XV/2.

Bakker, J.G.M. (in prep.) Late Pleistocene vegetational and climatic history of the Pitalito Basin, Eastern Cordillera, Colombia.

Bakker, J.G.M., Kleinendorst, Th. and Geirnaerts, W.A. (in prep.) Tectonic and sedimentary history of a Late Cenozoic intramontane basin (Pitalito Basin, Colombia).

F

CHAPTER 1

INTRODUCING THE SUBJECT

The study of the ecosystems of the Colombian Andean region started already some thirty years ago. Initially, special interest was paid to the paleoecology of the Eastern Cordillera by means of palynological studies of peats and lake sediments of which the publications of Van der Hammen & González (*e.g.* 1960, 1963) are the most important. These studies were supported by the studies on the actuoecology of the Andean vegetation by Cuatrecasas (*e.g.* 1934, 1958) and additional studies carried out by the aforementioned autors. The study of the Quaternary of Colombia soon grew out to a multi-disciplinary project including among others palynology, geomorphology, climatology, geology and archeology.

In 1980 the ECOANDES-project was started which led to a profound and well coordinated study of the montane regions of Colombia by means of a total of seven transects with an E-W orientation which, until now, are partly completed. The aim of the project is the study of the structure, function and evolution of the Colombian Andean ecosystems (Van der Hammen 1973). The results of this project and former studies are published in two series, *viz.*, 'The Quaternary of Colombia' of which until now 15 volumes have been published and 'Studies on Tropical Andean Ecosystems' with three volumes and another three volumes in preparation.

This study was carried out in the compass of the ECOANDES-project and might be integrated in the planned transect in the southern part of the Colombian Andean region.

The present study is focused on the transitional zone between the low tropical forests (c. 300-1000 m altitude) and the low montane, sub-Andean forest (c. 1000-2300 m altitude). In the dry and warm tropical zone there is extensive cattle-breeding and cultivation of rice and cotton. At the altitude of the relatively wet and colder low montane zone, agricultural activity is dominant: coffee, cane sugar, bananas and maize are important products. The area between c. 1000 m and 1500 m altitude forms in a certain way the climatic transition of these zones and is suitable for different types of land use. Climatic changes may have possible consequences on land use in this zone. Palynological studies from higher montane regions of the Colombian Andes point to the altitudinal shifting of the zonal forest belts in time due to climatic changes. This study investigates the role of climatic changes in the paste on natural processes in low montane areas and its effects on sedimentation processes.

The study area is confined to the Pitalito Basin (Fig. 1; 1320 m a.s.l.; lat. 1°51'N, long. 76°02'W), a sedimentary basin in the Eastern Cordillera in the southern part of Colombia (Huila Department). This basin was choosen for the following reasons:

- The basin is situated in the transitional zone of which very little paleoclimatic information was available up to now.
- The thick sedimentary deposits in the basin were expected to reveal information on

geomorphological, sedimentological and erosional processes that took place in this zone.
- The relatively high sedimentation rate in this basin enables a detailed reconstruction of sedimentary processes.
- The expected absence of large hiatuses in the sedimentary infill would make an almost complete palynological record possible.
- The presence of extensive organic-rich, swampy areas in the basin generally ensures a good pollen preservation, often a problem in these warm climates (Salomons 1986).

The initial objective of the present study was to look if the variation in the types of sediments could be related with climatic changes. Therefore, the study was concentrated on palynological and lithological analysis of some boreholes to register climatic fluctuations by means of paleoecological data. The idea was that variations in climate could be related to changes in sedimentary and erosive processes.

It was thought that changes in the sediment type were principally dominated by climatic control. However, it appeared that tectonic activity also plays an important role on the types of sediments and their distribution in the Pitalito Basin. Several authors already recognized the influence of both climatic and, especially, tectonic control on alluvial architecture and their respective sedimentological deposits (*e.g.* Schumm 1968; Miall 1978, 1980; Ballance & Reading 1980; Read & Dean 1976, 1982; Alexander & Leeder 1987). Therefore, the ultimate goal of this study became to identify and differentiate between the effects of both external factors (climate and tectonics) on sedimentological patterns in this sedimentary basin.

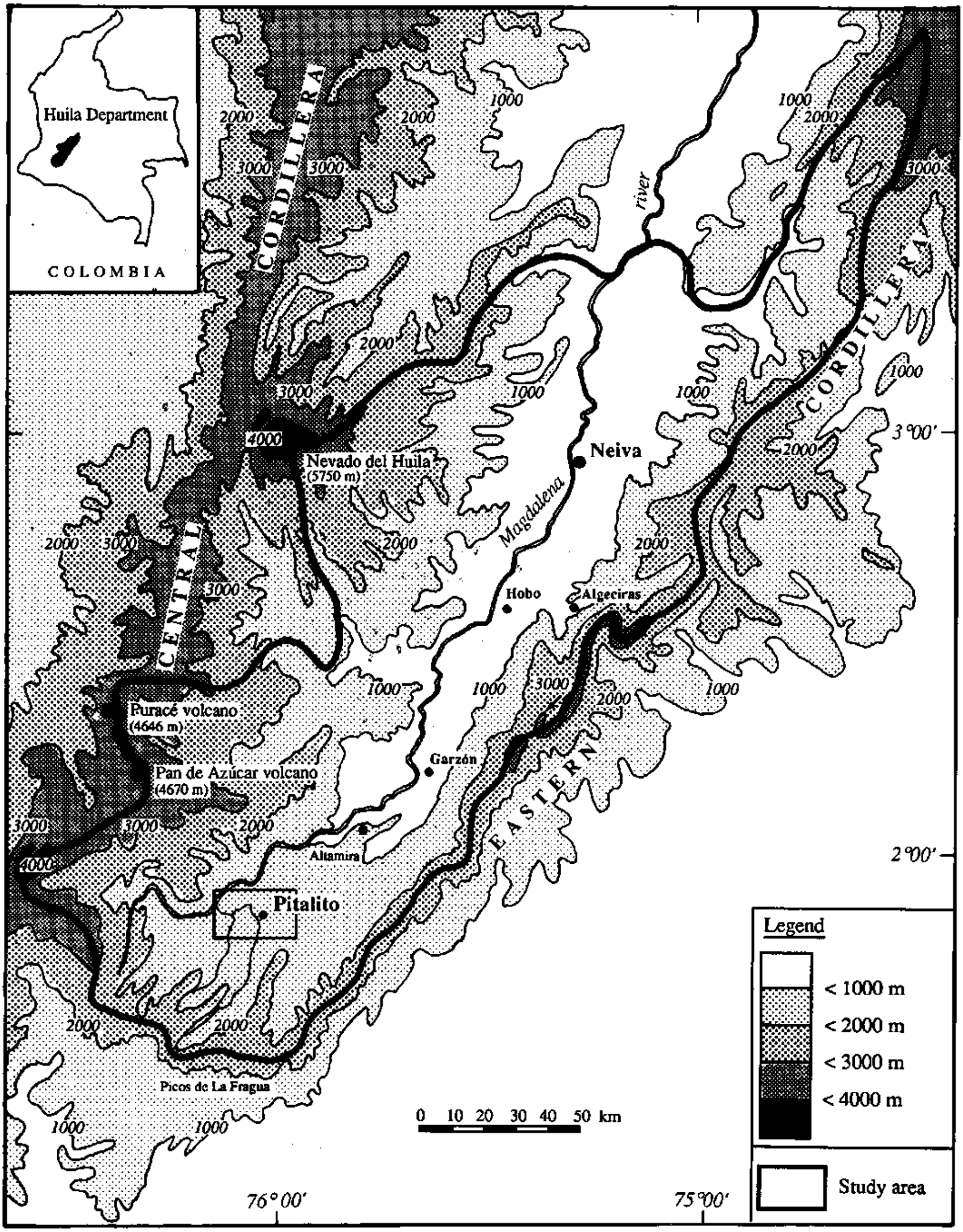


Fig. 1.
Geographical position of the study area: the Pitalito Basin. The solid line indicates the contours of the Huila Department.

CHAPTER 2

GENERAL INFORMATION ABOUT THE STUDY AREA

2.1 Regional geology

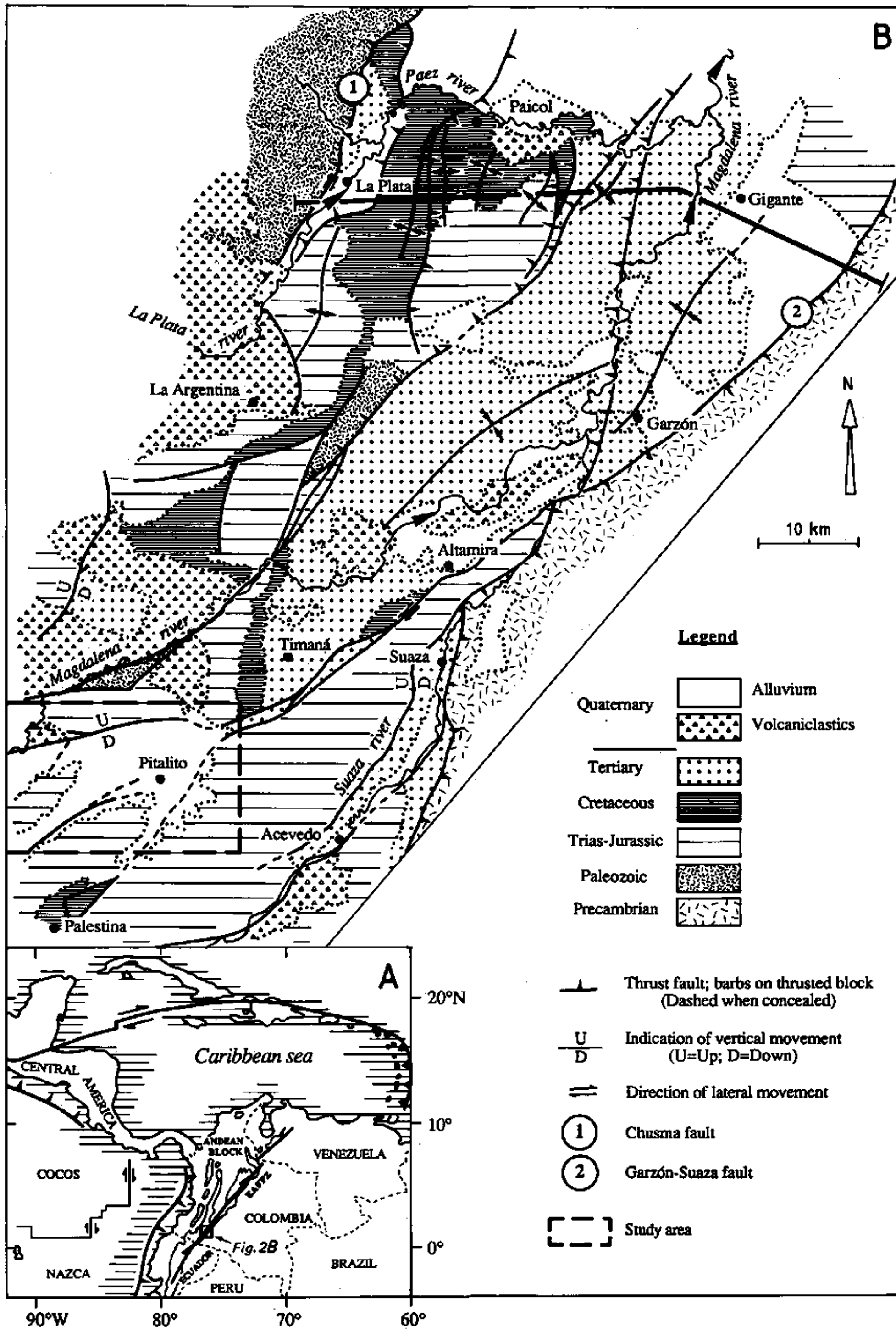
2.1.1 GEOLOGICAL SETTING

The northern Andes evolved due to the active thrusting of the east-dipping Nazca plate under the South American Plate. The underthrusting of this oceanic plate leads to deformation and crustal shortening along the western margin of South America and results in crustal uplift. Many studies concerning the evolution of the northern Andes have been published, among others by Bonini *et al.* (1984). In the scope of this thesis attention will be paid especially to the regional geology of the Andean mountain chain in southern Colombia.

At the latitude of the Pitalito Basin the Eastern Cordillera bifurcates from the Central Cordillera (Fig. 1). These Colombian cordilleras are separated by the Upper Magdalena Valley: a Neogene foredeep trough. Both Cordilleras, together with the Upper Magdalena Valley, are the main morphostructural elements at the latitude of the study area.

The Central Cordillera is the northward extension of the Cordillera Real and forms a regional crustal uplift of Paleozoic crystalline rocks (Irving 1975; Butler 1983; Butler and Schamel 1988). This mountain chain forms a positive feature since the Caledonian Orogeny (c. 400 m.y. ago) which is evidenced by metamorphosed Lower Paleozoic rocks which predominantly comprise marine sediments (Kroonenberg & Diederix 1982). The eastern side of the Eastern Cordillera is delineated by the 'Eastern Andean Frontal Fault Zone' (EAFFZ; Fig. 2A) which separates the Andean block from the South American craton (Guiana shield; Pennington 1981). Along this NE-striking fault zone the Andean block is moving NNE with respect to the craton and is predominated by dextral strike-slip movements. In southern Colombia, at the latitude of the study area, a branch of the Frontal Fault Zone crosses the Eastern Cordillera to its western side forming the Garzón-Suaza fault system (no. 2; Fig. 2B). The Eastern Cordillera is a region of divergent Neogene east- and west-verging crustal uplifts which are bounded by moderately steep to high-angle reverse faults at this latitude (Butler 1983; Butler & Schamel 1988). This mountain chain is underlain by Precambrian and Triassic-Jurassic crystalline rocks as shown by the extensive outcrops in upthrusted blocks in the area (Fig. 2B; Kroonenberg & Diederix 1982; Fuquen & Nuñez 1989).

The Upper Magdalena Valley is a marginal foreland basin with a similar age of development as the uplifts in the Eastern Cordillera. Two Neogene basins constitute the Upper Magdalena Valley: the Girardot Basin in the north and the Neiva Basin in the south (Van Houten & Travis 1968). Both basins developed during the uplift of the Eastern


B
Paez river
Paicol
Magdalena river
Gigante
La Plata
La Plata river
La Argentina
Garzón
N
10 km
Altamira
Magdalena river
Timaná
Suaza
Pitalito
Suaza river
Acevedo
Palestina
U
D
Legend
Quaternary
Alluvium
Volcaniclastics
Tertiary
Cretaceous
Trias-Jurassic
Paleozoic
Precambrian
Thrust fault; barbs on thrusted block (Dashed when concealed)
Indication of vertical movement (U=Up; D=Down)
Direction of lateral movement
1 Chusma fault
2 Garzón-Suaza fault
Study area
A
Caribbean sea
CENTRAL AMERICA
ANDEAN BLOCK
VENEZUELA
COCOS
COLOMBIA
Fig. 2B
NAZCA
ECUADOR
PERU
BRAZIL
20°N
10°
0°
90°W
80°
70°
60°

Cordillera and were filled in by Cretaceous and Neogene sediments. The latter originated from the erosion of the Central Cordillera. The Neiva Basin ceases to exist at the latitude of the Pitalito Basin, south of the town of Timaná, where it is bounded by upthrusted blocks of Triassic-Jurassic age (Fig. 2B).

2.1.2 GEOLOGICAL HISTORY

Little is known about the Precambrian and Paleozoic times. Due to the relatively high metamorphic grade of the Precambrian and Paleozoic sediments it is assumed that during pre-Cretaceous times several orogenetic phases lead to the metamorphism of these sediments. The high-grade metamorphic Precambrian rocks (cf. the Garzón Massif) may be explained by a model for continental collision during the Nickerie orogeny around 1200-1300 Ma ago (Kroonenberg 1982, 1983; Priem *et al.* 1989). The metamorphic Lower Paleozoic rocks represent epicontinental deposition and subsequent metamorphism as a result of an orogenetic event, possibly of Caledonian age. The Central Cordillera emerged as a positive feature, not to be submerged again (Kroonenberg & Diederix 1982). The Upper Paleozoic rocks consist of unmetamorphosed marine sediments which were deposited in a shallow shelf sea, east of the Central Cordillera.

After the upheaval of the Central Cordillera the Neiva Basin was the site where eastward thinning sediments were deposited. During Triassic-Jurassic time volcaniclastic sediments were deposited in a continental to shallow marine environment east of the Central Cordillera (Saldaña Formation; Table 1). This important magmatic event is related to a subduction system with an active volcanic arc (Butler 1983). Jurassic intrusion followed extrusion which is evidenced by the related grantitoid intrusive bodies. During the Cretaceous shallow marine to paralic sediments were deposited east of the Central Cordillera. The Cretaceous sediments range from coarse-grained quartz sandstones to fissile shales and multi-coloured mudstones (Caballos, Villeta, Guadalupe and Guaduas Formations; Table 1). Further uplift of the Central Cordillera caused the regression of the seas and the Cretaceous sediments were partially eroded. Clastics were shed eastwards from the Central Cordillera forming coalescent alluvial fans (Van Houten & Travis 1968) which make up the Gualanday Formation and lasted to Upper Oligocene or Lower Miocene (Van der Hammen 1958, 1961).

The Neogene sequence starts with the Honda Formation (Middle Miocene) and is made up of conglomerates of western origin which were deposited by eastward flowing alluvial systems (Wellman 1970; Van der Wiel, in prep.). With the initiation of the uplift of the Eastern Cordillera all the pre-Neogene rocks were deformed. Renewed uplift and increasing volcanism in the Central Cordillera is reflected by the Neiva Formation (Upper Miocene?) and the Gigante Formation (Late Miocene/Pliocene). The upper part of the latter consists of conglomerates of eastern provenance suggesting that from the Pliocene onward the Eastern Cordillera forms a topographic barrier (Kroonenberg & Diederix

Fig. 2.
A/ Regional structural setting of the Eastern Cordillera of Colombia and the Eastern Andean Frontal Fault Zone (EAFFZ; redrawn from Pennington 1982).
B/ Geological map of the southernmost part of the Upper Magdalena Valley including the Pitalito Basin (modified from Kroonenberg & Diederix 1982). The line between La Plata and Gigante illustrates the position of the cross-section depicted in Fig. 3.

1982; Van der Wiel 1989). According to Howe (1974) the direction of sediment transport was still towards the east.

During the Pliocene/Quaternary uplift, deformation continued and was accompanied by an increase in volcanic activity in the Central Cordillera. Enormous amounts of volcaniclastic sediments (lahars and ignimbrites) filled in the valleys and predominantly determine, together with the upthrusted blocks of Triassic-Jurassic rocks, the main morphology of the southern part of the Upper Magdalena Valley.

Table 1.
General stratigraphy of the exposed lithological formations around the Pitalito Basin.

ERA	PERIOD	EPOCH	FORMATION	ENVIRONMENT & EVENTS		
CENOZOIC	QUATERNARY	Holocene	Infill of the Pitalito Basin	*Fluvial/lacustric*		VOLCANISM
		Pleistocene				
	—2 Ma—					
	NEOGENE	Pliocene	Lahars / Fm. Guacacallo	*Continental Volcanic*	UPLIFT E. CORD.	
		Miocene	(not exposed)	*Fluvial*	INTERMITTENT UPLIFT C. CORD.	
	—22.5 Ma—					
	PALEOGENE	Oligocene	Fm. Gualanday	*Fluvial*		
		Eocene			EROSION	
		Paleocene	Fm. Guaduas	*Lagoonal*	REGRESSION	
	—65 Ma—					
MESOZOIC	CRETACEOUS	Upper	Fm. Guadalupe	*Coastal*		
			Fm. Villeta	*Shallow marine*		
		Lower	Fm. Caballos	*Coastal*	TRANSGRESSION	
	—141 Ma—					
	JURASSIC		granitoid intrusions	*Shallow marine*	ARC VOLCANISM	
	—195 Ma—					
	TRIASSIC		Fm. Saldaña			
	—230 Ma—					
PALAEOZOIC	Upper		(not exposed)	*Marine*	EMERGENCE CENTRAL CORDILLERA	
	Lower		metamorphic shales and conglomerates	*Epicontinental*		
	—570 Ma—					
PRECAMBRIAN			(not exposed)	*Continental*	CONTINENTAL COLLSION (?)	

2.1.3 STRATIGRAPHICAL RECORD

The oldest type of rock that is exposed in the study area is possibly of Lower Paleozoic age. The surrounding mountains of the Pitalito Basin are completely dominated by rocks of Triassic-Jurassic age. Cretaceous and Tertiary sediments are present in the neighbouring northern and southern valleys near Timaná and Palestina (Fig. 2B). The sedimentary rocks of Neogene age are not exposed in the study area. Quaternary deposits consist of volcaniclastic sediments and are exposed in the northern part of the basin. Unconsolidated fluviolacustrine and fluvial sediments constitute the sedimentary infill of the Pitalito Basin and are extensively described in this study (Chapters 3, 4 & 5). Below a brief lithological description is given for the rock units which occur in the study area.

For more detailed information concerning the lithological formations which are exposed in the Upper Magdalena Valley the reader is referred to: Julivert (1968), De Porta (1974), Wellman (1970), Howe (1974), Cediel *et al.* (1981), Kroonenberg *et al.* (1981), Kroonenberg & Diederix (1982) and Van der Wiel (in prep.).

<u>Lower (?)Paleozoic</u>

Outcrops of quartzitic metaconglomerates and low-grade metamorphic shales enriched in pyrite were found during the present study in the most northwestern part of the region near the confluence of the Guarapas river with the Magdalena river. According to Kroonenberg (oral information) this might be the oldest type of rock in the study area. The nature of its contact with the other rock units is unknown.

<u>Triassic-Jurassic</u>

-Saldaña Formation The Pitalito Basin is almost completely enclosed by this formation together with the Jurassic intrusive rocks. The main constituents of this formation around the Pitalito Basin are acid to intermediate volcanic rocks and dacitic-andacitic lavas which according to recent studies belong to the Prado Member (Upper Saldaña Formation; *e.g.* Mojica & Llinás 1984). In the northern mountain fringe of the basin this formation consists of, besides aforementioned types of rocks, cherty intercalations and dark-coloured conglomerates in a calcareous matrix both belonging to the Chicalá Member (Lower Saldaña Formation). The small convex shaped hills in the basin itself are outcrops of light-coloured intermediate volcanic rocks with a very breccious character.

-Jurassic intrusions In the northern mountain range and at the easternmost side of the basin some outcrops of this rock type are known. Acid to intermediate intrusives (granodioritic to andesitic) in the form of plutons and small stocks predominate and are related to the extrusives of the Saldaña Formation. The Saldaña xenoliths in the igneous rocks confirm that the intrusions succeeded the Saldaña volcanics. Cretaceous sediments were not affected.

<u>Quaternary</u>

-Lahar deposits (Upper Pliocene?) Large angular clasts of dark andesite which are embedded in a strongly cemented dark-coloured matrix. The lahar may reach a thickness of several tens of metres and is affected by the fault which delineates the northern margin of the Pitalito Basin (Appendix I). Most probably this occurrence in the northwestern part of the area is the equivalent of the Altamira lahar: a volcanic breccia of Pliocene/Pleistocene(?) age that can be followed over tens of kilometers coming from the Central Cordillera (Kroonenberg & Diederix 1982).

-Guacacallo Formation (Pliocene?) The formation is made up of rhyolitic ignimbrites with intercalations of intensely weathered fanglomeratic deposits (Kroonenberg *et al.* 1981; Van der Wiel, in prep.). These volcaniclastic sediments originate from several volcanic eruptions and have an estimated volume of 100 km^3 (Kroonenberg *et al.* 1981). Whether the sediments are of younger, older or similar age as the lahars is unknown. East of the study area the sediments form a dissected plateau extending over 1000 km^2. Outcrops are found at the NE side of the study area near the Guacacallo village and form the easternmost exposures of this formation in the area.

-Sedimentary infill of the Pitalito Basin (≈Middle Pliocene-Holocene) The basin infill consists of unconsolidated fluvial and fluvio-lacustrine sediments. In the Chapters 3, 4 & 5 these unconsolidated sediments will be discussed in detail.

2.1.4 GEOLOGICAL STRUCTURES

The Neiva Basin is related structurally to the uplift of the Eastern Cordillera which started around Early Miocene (Butler 1983). During this orogenic event the Neogene sediments were folded and formed broad synclinoria and small anticlinoria. NE/SW-oriented fault systems appear and in time the sediments show an increase in eastern components indicating the increasing relief of the Eastern Cordillera (Van der Wiel, in prep.). Butler (1983) and Butler & Schamel (1988) describe two main fault systems which dominate the southern part of the Neiva Basin (Fig. 3):

(1) The Chusma fault system delineating the eastern side of the Central Cordillera and characterized by *en echelon* west-dipping thrust faults and

(2) The Garzón-Suaza fault system which bounds the western side of the Eastern Cordillera and characterized by *en echelon* east-dipping reverse faults.

(1) The Chusma fault system The subducting Nazca plate lead to a compressive system and crustal shortening occur deep beneath the Central Cordillera. This compression resulted in a fault system which consists of numerous *en echelon* east-verging thrust faults which developed sequentially from west to east during the Middle to Late Oligocene: the Chusma fault system (Fig. 3). During the eastward migration of the system the throw on the faults diminishes. The western margin of the system is bounded by the Chusma-thrust: a major west-dipping basement fault (no. 1; Figs. 2 & 3). To the south the system is truncated by the younger Garzón-Suaza fault system. The Chusma fault system is approximately 250 kms in length and 50 kms in width and narrows towards the north. It became inactive when the Garzón-Suaza fault system developed (Butler 1983).

(2) The Garzón-Suaza fault system During the Early Miocene the shallow subduction of the Nazca plate lead to crustal upheaval east of the Chusma fault system. The uplift of the Eastern Cordillera initiated and a west-verging, high-angle, reverse fault system developed: the Garzón-Suaza fault system. This fault system borders a.o. the Garzón Massif, a Precambrian high-grade metamorphic complex cropping out in the Eastern Cordillera along the Garzón-Suaza thrust, yielded an age of 1180 Ma (Alvarez & Cordani 1980; Alvarez 1981; Priem *et al.* 1989). This Precambrian rock unit is considered to form the uplifted part of the western border of the Guiana shield (Kroonenberg 1982). According to preliminary fission-track datings the massif was uplifted some 12 to 9 Ma ago (Van der Wiel 1989). The Garzón-Suaza fault system dominates in the Pitalito Basin where it diverges in an active northern fault and an inactive southern fault. More detailed information about the structural setting of the Pitalito Basin itself will be discussed in Chapter 3.

Although Butler (1983) could not detect strike-slip movement along the Garzón-Suaza fault system, right-lateral displacements along faults are very common in view of the regional structural setting. This is evidenced by the northern continuation of the 'Eastern Andean Frontal Fault Zone' (Pennington 1981) which includes, among others, the

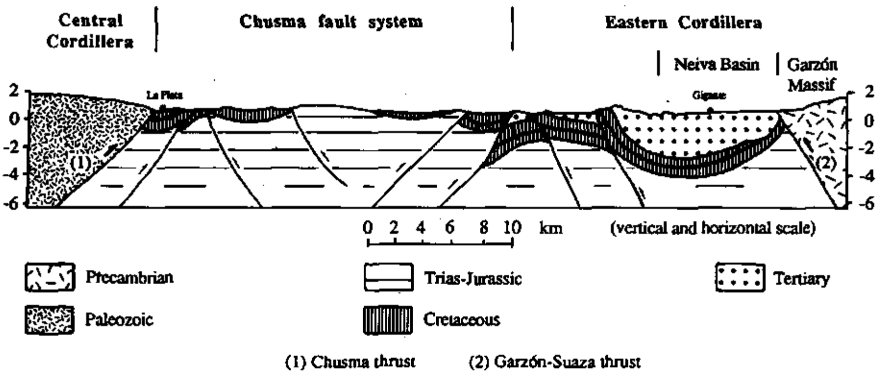


Fig. 3.
Cross-section through the Upper Magdalena Valley at lat. 2°30'N showing the principal geological structures (modified from Butler 1983).

Boconó fault. This fault of Pliocene-Pleistocene age is situated in the Merida Andes (Venezuela) and shows a lateral offset of 40-50 km which is ascribed to dextral strike-slip displacements (Schubert 1983). Boinet *et al.* (1985) suggest for both, the Colombian Eastern Cordillera and the Merida Andes, a synchronous evolution with similar structural features starting from the Upper Miocene onward.

Chorowicz *et al.* (1987) postulated that, based on SPOT images, the Garzón-Suaza fault does show dextral strike-slip movements in the Upper Magdalena Valley. They consider the Algeciras Valley which is situated c. 100 km northeast of the Pitalito Basin (Fig. 1) a small pull-apart basin with right-lateral displacements. This observation is supported by observations of Diederix (oral information) who found lateral dislocated alluvial fans near El Hobo village which lies c. 15 km west of the Algeciras Valley and in the Timaná Valley where outcrops show evidence for slip movements. Dextral strike-slip movement in the Upper Magdalena Valley is also reported by Guillande (in prep.) and Vergara (in prep.). This present study (Chapter 3) points to the same conclusion.

2.2 Geomorphology of the Pitalito Basin

The landforms of the area of investigation have been studied by making geomorphological maps which finally resulted in a coloured map, scale 1: 40,000 (Appendix I). The interpretation and delineation of the respective land units were studied with the aid of aerial photographs. The lithological content of the individual landforms was studied in the field with the help of boreholes or exposures, *e.g.* along the incisions of both main tributaries or clay pits. The latter method was especially applied in the area west of the Guarapas river where numerous excavated pits were exploited to extract clay. In this part of the basin it was impossible to make boreholes due to the compact character of clay and silt and the local presence of very coarse material. The genesis of the land units was deduced from shape, material and their geomorphological setting.

The basis of the geomorphological map (Appendix I) is formed by the isohypses which have an accuracy of ±5 m and are depicted with a contour interval of 10 m for altitudes below 1350 m and a contour interval of 100 m for the altitudes above 1400 m. The altitudinal data were obtained by digitizing the areal photographs and applying the required photogrammetrical corrections.

2.2.1 DRAINAGE

The basin is drained by two main rivers, the Guarapas river and the Guachicos river. Both rivers have their source in the 'Picos de La Fragua' area at an approximate elevation of 2500-2800 m (Fig. 1). This NW-SE oriented transversal ridge connects the Eastern Cordillera with the Central Cordillera. This means that nowadays the entire course of both rivers is located in the mountainous forest belt.

The most conspicuous morphological feature in the Pitalito Basin is the deep incision of the Guarapas river (C; Photo 1) which traverses the flat topography from south to north and dissects the basin plain into a western and an eastern part. At the northern side of the basin the river abuts onto the northern hills, changes its course towards the west and flows parallel to the northern basin margin. Relatively low-sinuosity reaches alternate with more sinuous ones which is apparent from the upstream part of the Guarapas river where sinuosity is high compared with the more downstream area. The valley floor is about 200-250 m wide and consists of several terraces (not indicated in the map). The river incises into the unconsolidated Quaternary basin fill to a maximum depth of about 40 m forming very steep and high valley walls. These walls are very unstable due to their steepness and the material which they are made of (predominantly massive clay; Photo 4). These clays show vertical cracks parallel to the wall which make these outcrops dangerous to study. Undercutting by the river frequently results in a collapse of the entire wall and large masses of material slump into the river.

Another important contributor to the drainage system is formed by the Guachicos river which is nowadays restricted to the western side of the basin. The incision of this river merely exceeds 15 m in the upstream part and the banks consist of coarse material (pebbles and cobbles; Photo 5). Bank recession especially occurs along the right side of the river by undercutting. Both rivers are single-channeled. The maximum grain size transported by both rivers at the latitude of the basin differs considerably; the Guachicos river transports pebbles and cobbles along its entire course whereas the most important sediment load for the Guarapas river consists of the sand-size fraction. The rivers converge nearby the NW outlet of the basin after which they cut down through the hard rock and join the Magdalena river at an altitude of c. 1200 m.

Western part

The area west of the Guarapas river is considered as the western part of the basin. The density of drainage is low and a distinct pattern in the drainage system is lacking. Numerous small depressions are present which are seasonally ponded. These depressions seemed to be connected to each other by indistinct gullies which are only visible in the areal photographs during the wet season (March-July; Fig. 7). The percolation of water is very slow due to the high content of clay in the soils which makes them highly impermeable (Espinal and Petrelli 1966). Therefore, water remains on the surface for several weeks which is aggravated by the flat topography of the area. On the whole, the western drainage pattern is of a deranged character: a pattern which testifies that up to now insufficient time was available for drainage integretation.

The cardinal direction of the incipient drainage system is heading NE towards the Guarapas river. This leads to deep incisions along the western valley side as the difference in elevation between the Guarapas river and the western drainage system amounts about 25-40 m (Fig. 4B).

Eastern part

The eastern part comprises the area east of the Guarapas river. Drainage is very poor especially in the northern region where swamps dominate (C; Photo 2). This area called La Coneca is the only site in the basin where peat of Holocene age can be found (Chapter 4). The intermittent streams coming from the northern mountain range debouch into the swamps. Canals are constructed to drain the terrain artificially. The only perennial stream is coming from the outermost NE-part of the basin (Regueros rivulet; Appendix I). This stream dissects the relative easily erodable volcaniclastic sediments belonging to the Guacacallo Formation and shed its load in the form of an extended alluvial fan into the Pitalito Basin (D; Photo 2). The rivulet traverses the eastern part of the basin to join the Guarapas river at the latitude of Pitalito town.

The southern area is better drained but still poor of character. The rectangular-like drainage pattern which is formed by the small intermittent streams is predominatly determined by the presence of NE/SW-oriented sandy ridges (b; Photos 1, 2 & 3). Initially, these ridges may have formed a kind of divide between the lower lying areas which they enclose entirely. They force the small streams (C; Photo 3) in the same direction until the streams are able to break trough the ridges and can change their course (A; Photo 3). Such breaches through the ridge might be the result of for instance later compaction of the ridges or seepage (Louwe Kooijmans 1974).

The small intermittent streams show an anomalous flow direction south of Pitalito town (*e.g.* El Higueron area; Appendix I); initially they run towards the NE whereas the cardinal direction of the actual drainage system is NW (C & A, respectively; Photo 3). Further downstream these small streams display sharp-angle turns and the flow direction inverts towards the NW. Furthermore, the streams belonging to the actual NW heading drainage system incise into the basin fill which is in contrast to the older generation of streams. Apparently, these small intermittent streams are a relict of an older fluvial system with a NE paleocurrent in times when the local base level was at least as high as the surface level of the basin infill (c. 1300 m altitude). Nowadays, the base level is about 100 m lower and is formed by the Magdalena river at c. 1200 m altitude. The changing flow direction and the contemporaneous incision of the actual main tributaries must be ascribed to an extrabasinal (allocyclic) control which forced the streams to displace their course from a NE to a NW direction in combination with the lowering of the local base level. In summary, then, whereas in former times the subsiding basin formed the locus of deposition and was filled in by a NE-flowing fluvial system, erosion by a NW-flowing drainage system is the dominant proces of today.

Chapter 4 indeed shows that a former fluvial system with a NE paleocurrent characterizes the last stage of sedimentary infill of the basin.

2.2.2 BASIN PLAIN RELIEF

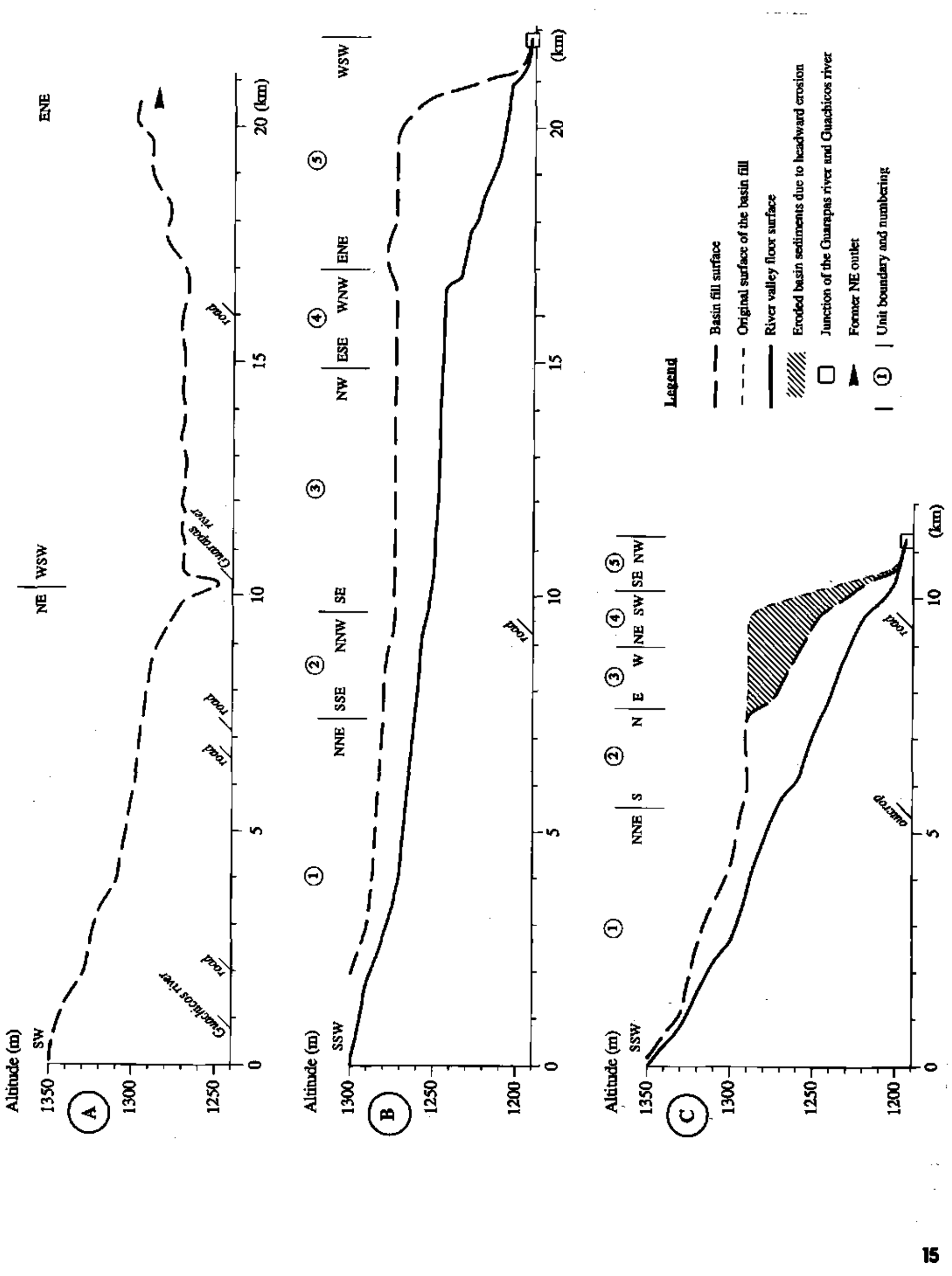
The topographic framework of the area of investigation is in many aspects determined by the sediments which fill in the basin. The flat topography of the Pitalito Basin is in strong contrast with the surrounding mountainous regions which enclose the basin (Photo 10). The northern hills reach an altitude of 1800 m and the crests of the southern mountains range between 2000-2500 m. West of Pitalito town some convex-shaped isolated hills raise about 30-50 m from the flat topography of the basin fill. These hills are made up of intensely fractured acid to intermediate volcanic rock (Saldaña Formation) overlain by a thick weather mantle (5-10 m). Locally some closed depressions are present at the sub-rounded tops. Similar features are found near Garzón in metamorphic rock (Ruiz, 1977) and ascribed to pseudo-karstic activity, *viz.*, physical and chemical processes leading to the solution of minerals. This process is intensified by the presence of multiple joints that facilitate the infiltration of water into the underlying hard rock.

The top of the lahar deposits around the outlet of the actual drainage system of the Pitalito Basinin lies at an altitude of c. 1400 m. This volcaniclastic material is dissected by the Guarapas river to a depth of c. 1190 m. In the outermost NE of the study area a pass forms the divide between the Pitalito Basin and the Timaná Valley. The pass lies at an altitude of c. 1295 m (Fig. 4A) and is structurally related with an assemblage of NE/SW-oriented faults (Fig. 2B). This pass might have been the outlet of the NE-flowing fluvial system before c. 17,000 years B.P., *viz.* before the 90° turn of the river system to the actual outlet (sections 2.2.1 & 4.2).

A SW/NE-profile along the basin axis (Fig. 4A) displays topographical phenomena which show evidence that the subdivision of the basin in a western and eastern part is not only based on differences in the drainage pattern, but is also expressed in the topography. The western part generally slopes eastward with an angle ranging between 0°-1° (Fig. 4A). At the latitude of the Guarapas river a vertical offset of c. 10 m is reached over a relatively short horizontal distance. This step in the elevation reflects the possible presence of a fault with an eastern downwarped block. The existence of such a fault is indeed supported by the gravimetrical data and will be discussed in more detail in Chapter 3.

East of the Guarapas river the topography is almost horizontal and fluctuates around 1270 m altitude. The micro relief is characterized by a variety of ridges (Photos 5 & 10) which enclose moderately to poorly drained depressions. The ridges are straight to low-sinuous with a convergent and divergent character (Appendix I). The overall pattern however, is a convergent one towards the east. The interconnection of these ridges is high and some of them end in badly drained areas (*e.g.* La Coneca). The width of

Fig. 4. (next page)
Schematic relief profiles: A/ SW/NE-oriented profile along the basin axis of the Pitalito Basin. Note the jump at the position of the Guarapas river which is situated at the break of the eastward-sloping western part and the (sub)-horizontal eastern part. B & C / Longitudinal profiles along the Guarapas river and the Guachicos river, respectively. The surface of the basin sediments is taken as abney level. The dashed lines represent assumed elevations of the abney level (see text).


A
Altitude (m)
1350
1300
1250
0
5
10
15
20 (km)
SW
NE
WSW
ENE
Guachicos river
road
road
road
Guarapas river
road
B
Altitude (m)
1300
1250
1200
SSW
NNE
SSE
NNW
SE
NW
ESE
WNW
ENE
WSW
1
2
3
4
5
road
(km)
C
Altitude (m)
1350
1300
1250
1200
SSW
NNE
S
N
E
W
NE
SW
SE
NW
outcrop
road
(km)
Legend
Basin fill surface
Original surface of the basin fill
River valley floor surface
Eroded basin sediments due to headward erosion
Junction of the Guarapas river and Guachicos river
Former NE outlet
1 | Unit boundary and numbering

the ridges varies between 20-45 m and they have a maximum relief of about 5 m above the adjacent areas. The poorly drained, lower-lying areas represent c. 60-70 % of the eastern half of the basin. The origin of these morphological features will be discussed in more detail in Chapter 4.

The relative flat topography of the basin fill is disturbed by the incision of two main distributaries of which especially the Guarapas river forms steep and high valley walls. Two longitudinal profiles along both the Guarapas river and the Guachicos river reveal more aspects about their gradient. Based on the orientation of the river course both profiles can be divided into 5 units (Figs. 4B and C).

-Guarapas river (Fig. 4B) At the entrance in the wide plain of the Pitalito Basin, the stream have incised into the basin fill to a depth of about 10-15 m (unit 1). This SSW/NNE-oriented part of the stream shows a relative low gradient (0°-1°) and a concave slope. The depth of the incision gradually increases to c. 30 m (units 2, 3 & 4). The gradient changes when the river abuts onto the northern hills and alters its course from a WNW-direction towards the WSW (transition from unit 4 to 5). This part of the river displays a gradient of 1°-2° and the valley depth increases from c. 30 m to 70 m over a distance of about 5 km.

-Guachicos river (Fig. 4C) At the entrance into the Pitalito Basin the stream beds are cut down into the basin sediments to a depth of c. 5 m (unit 1) which gradually increase to a depth of c. 20 m until the river changes its course towards the north, perpendicular to the adjacent NW/SE-oriented hills (unit 2). In this part the gradient increases to 1°-2° and the valley reaches a maximum depth of 45 m at the transition to unit 3. From this part further downstream, the surface of the sedimentary infill gradually decreases to an altitude of c. 1200 m.

The sudden increase in valley depth in both rivers can in fact be explained by the same cause: where the rivers are no more obstructed by a topographic barrier and can run freely towards the NW outlet, they will take the shortest possible course thereby creating a steeper gradient.

For the Guarapas river this barrier is formed by the updoming basement in the NW-part of the basin (Fig. 11; section B-B') and the east-dipping western part of the basin (Fig. 4A). Both features force the river to the east. Otherwise, one would expect the river to cross the basin more diagonally: heading more directly towards the NW outlet. The relatively straight NW course of the Guarapas river (unit 3; Fig. 4B) is also affected by the fault-bounded step in the topography (Fig. 4A).

Obviously, the outcrops north of the Guachicos river form a topographic barrier and initially force the river to flow towards the NE. From unit 2 (Fig. 4C) further downstream the stream is able to alter its course more directly towards the outlet and cuts down deep into the basin sediments. In the last three units, the surface of the basin fill is lowered by headward erosion from the junction of the Guarapas river with the Guachicos river (Fig. 4C; shaded area). The original surface must have resembled the steep valley walls along the Guarapas river.

2.3 Present climate

The distribution of precipitation, cloudiness, temperature, *etc.* varies considerably from one place to another in Colombia and depends on various factors like relief, altitude and latitude. Besides, the climate is dominated by the presence and location of the Inter Tropical Convergence Zone (ITCZ). The ITCZ forms the asymptotic confluence zone of the northern and southern trade winds and agrees with a cyclonic character of the weather, *viz.*, higher percentages of cloudiness, higher mean precipitation and higher wind velocities (Hastenrath 1985). The axis of the zone is at its equatormost position during the northern winter (February/March) and at its most northern position (lat. 5-10°N) during northern summer (July/August) which means that most parts of Colombia are influenced by the displacement of the axis. In a coarse spatial resolution this would imply a comparatively dry season during the period around January/February for the northern part of Colombia whereas during the passage of the ITCZ (May/June) there would be a relatively wet period. The reverse would apply to the southern part of Colombia.

The regional climatic differences in the Andean area are primarily dominated by the altitude and the position of the respective sites relative to the mountain chains. The high chains of the Andes with their N/S orientation form a threshold for the NE to SE oscillating trade winds ('easterlies') and the very humid western winds ('westerlies') coming from the Pacific. The precipitation regime at the outer sides of the cordilleras differs therefore considerably from the regime of the intra-Andean regions. With the exception of the western side of the Central Cordillera the exterior slopes of the Colombian Andes are more humid than its inner parts (Fig. 5). At the Pacific side of the Colombian Andes the pluviometric optimum lies between 50 m and 500-1000 m altitude (not shown) and at the eastern outerside between 400 m and 1000 m altitude (Fig. 5 and Oster 1979). As the Western Cordillera at the latitude of the Pitalito Basin is relatively low with summits ranging between 2500 m and 3000 m altitude, very humid westerlies can pass this mountain range and subsequently are met by the higher Central Cordillera. This explains the asymmetric pluviometric character of the Central Cordillera: the western side has a pluviometric optimum of c. 3000-4000 mm (Fig. 5) and the eastern side of c. 1500 mm. As the Magdalena Valley at the latitude of the study area is narrow its influence on regional differences in precipitation is apparently negligible. Oster (1979) assumes that the precipitation regime at the eastern side of the Central Cordillera does not change significantly above c. 2000 m altitude at the latitude of the Pitalito Basin (curve 1A; Fig. 6). This is not consistent with the available data obtained from the HIMAT (Servicio Colombiano de Meteorología e Hidrología: Boletín Climatológico Mensual y Anuario Meteorológico). These more recent data might indicate an increasing humidity at higher elevations (curve 1B; Fig. 6). A similar, so called secondary maximum is also described by Vis (1989) for the western side of the Eastern Cordillera at lat. 5°N.

The intra-Andean regions generally show a precipitation regime with a bimodal character determined by the position of the ITCZ. There are two comparatively dry seasons (Dec.-Febr. and June-Sept.) and two wet seasons (March-June and Oct.-Dec.) which might be

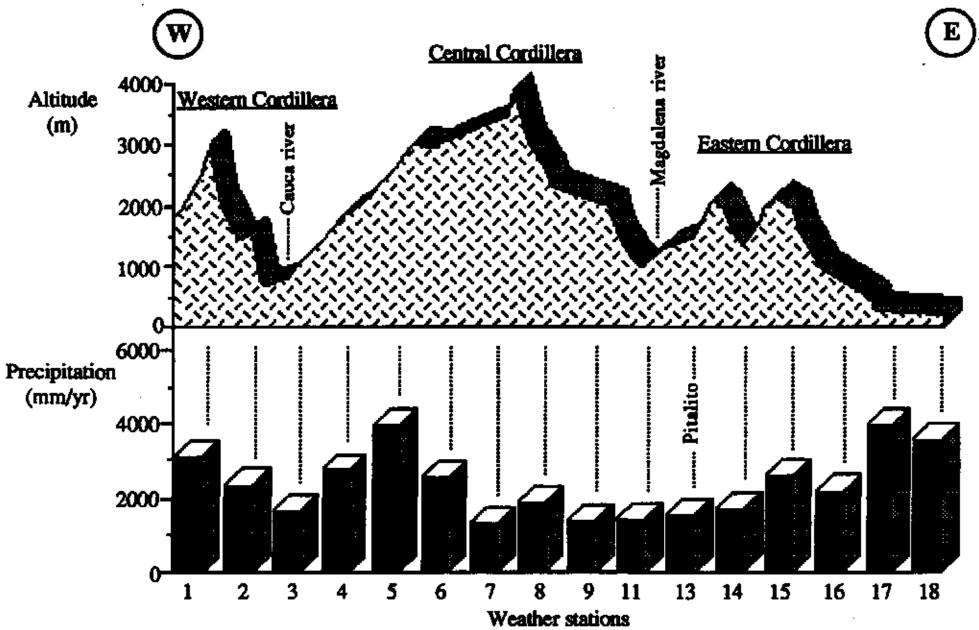


Fig. 5.
Mean annual precipitation totals along a schematic cross-section of the Andean region at the latitude of the study area. For the numbering and location of the weather stations see Fig. 8.

expected in the Pitalito Basin too. However, whereas the first dry and the first wet period is registered in the Pitalito Basin, the second dry period around June-Sept. is nearby absent and the amount of precipitation is almost equal to that of the second wet period Oct.-Dec. (Fig. 7). This can be partly explained by the intensified easterlies during the period May-August (Oster 1979) especially around the mid-troposphere (c. 4000-5000 m altitude; Hastenrath & Lamb 1977, 1978). During this period the easterlies can extend its influence into the Upper Magdalena Valley. The passage of these eastern instable air masses into the intra-Andean regions is facilitated by the low local relief of the Eastern Cordillera (Fig. 5).

The mean monthly temperature in the Pitalito area is about 20°C without significant fluctuations (Fig. 7). So is the difference between monthly minimum and maximum values. Trojer (1959) found for southern Colombia a temperature gradient of 0.60°C to 0.65°C per 100 m in the 'coffee-zone' (c. 1100-2000 m altitude) whereas in higher regions the gradient declines to 0.55°C/100 m.

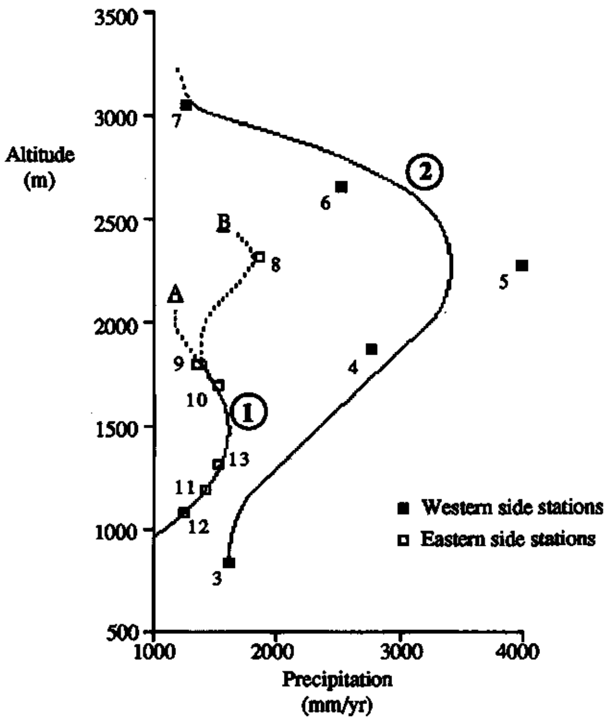


Fig. 6.
The mean annual precipitation along two transects at the eastern and western side of the Central Cordillera (curve 1 and 2, respectively). Data from Oster (1979; curve 1A) and from HIMAT (curves 1B and 2). For numbering and location of the weather stations see Fig. 8.

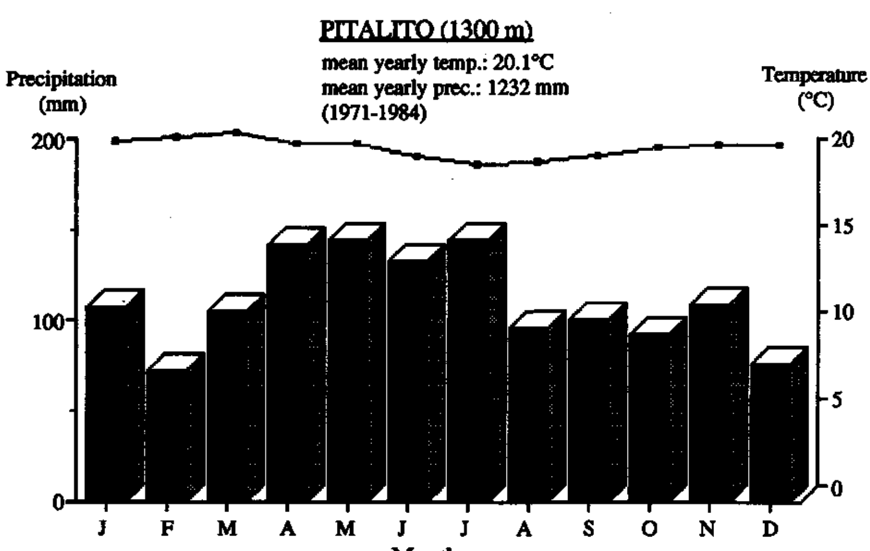


Fig. 7.
Monthly temperature (line) and rainfall conditions (bars) in the Pitalito Basin.

2.4 Present vegetation

The vegetation of the Colombian Andes has been described by several authors, among others: Cuatrecasas (1934, 1958), Espinal (1977), Cleef (1981), Cleef *et al.* (1983), Cleef & Hooghiemstra (1984), Cleef *et al.* (1984), Sturm & Rangel (1985), Rangel & Franco (1985), Rangel & Lozano (1986 and in prep.) and Rangel *et al.* (in prep.). Except for the publications of Rangel and co-autors (see also Fig. 8), detailed descriptions of the vegetation for the southern part of Colombia are scarce. Most of the studies are concentrated on more northern locations, especially around lat. 5°N. At this latitude the Eastern and Central Cordillera are separated by a wide Magdalena Valley (c. 50 km wide, altitude c. 300 m) so that the altitudinal vegetation belts of both mountain chains are clearly distinguishable and separated by a dry tropical woodland vegetation in the valley. This is in sharp contrast with the width of the Upper Magdalena Valley at lat. 2°N which lies in the order of several hundreds of meters at an altitude of c. 1200 m (Fig. 9). This implies that at this lower latitude the transition from the western flank of the Eastern Cordillera to the eastern side of the Central Cordillera is indistinct in terms of topography,

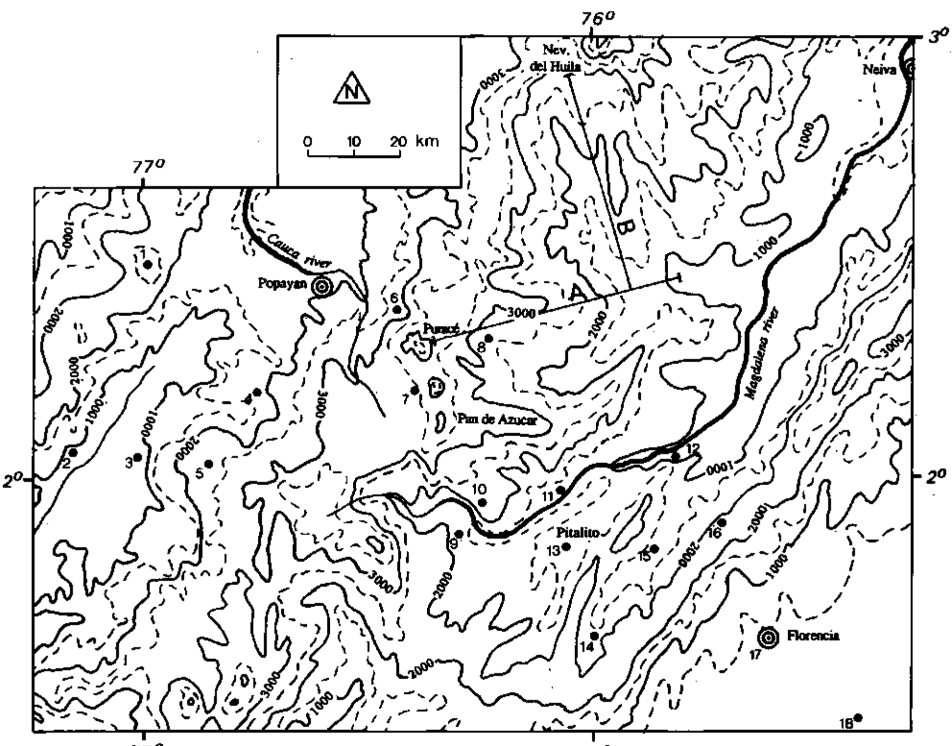


Fig. 8.
Approximate location of weather stations (1-18) and vegetation transects based on the studies of Rangel & Franco (1985) and Rangel & Lozano (1986). A/ Transect along relatively dry slopes; B/ Transect along humid slopes.

climatic condition and type of vegetation. In addition, both mountain chains have a different elevation: at lat. 5°N the Central Cordillera on the whole is higher than the Eastern Cordillera whereas at lat. 2°N both cordilleras have about the same altitude.

The dissimilarities of the regional topography between the two areas make it questionable whether above-mentioned studies are applicable for southern Colombia. They may lead to differences in mean annual rainfall which may effect the position and composition of the zonal vegetation types. Mean annual temperatures are mainly determined by altitude and, besides some diferences in average lapse rate, can be considered to be the same in both areas. As the mean annual temperature is the main discriminating factor for the distribution and composition of the zonal vegetation types it is assumed that aforementioned studies in the more northern areas can be utilized for the more southern parts of Colombia as well.

2.4.1 ZONAL VEGETATION

The aforementioned studies show that slopes facing the slopes along the Magdalena Valley are relatively dry and those facing the Amazon basin are humid (Fig. 5). Accordingly, we may expect to find the more humid zonal vegetation types at the eastern slopes of the Eastern Cordillera, whereas the more arid equivalents are supposed to prevail at the internal valley sides. A concise overview of the present zonal vegetation types around the study area will be presented here on the basis of the altitudinal zonation of the Colombian Andean vegetation as proposed by Cuatrecasas (1958) and Cleef (1981). From low to high altitudes the following vegetation belts are distinguished (Fig. 9):

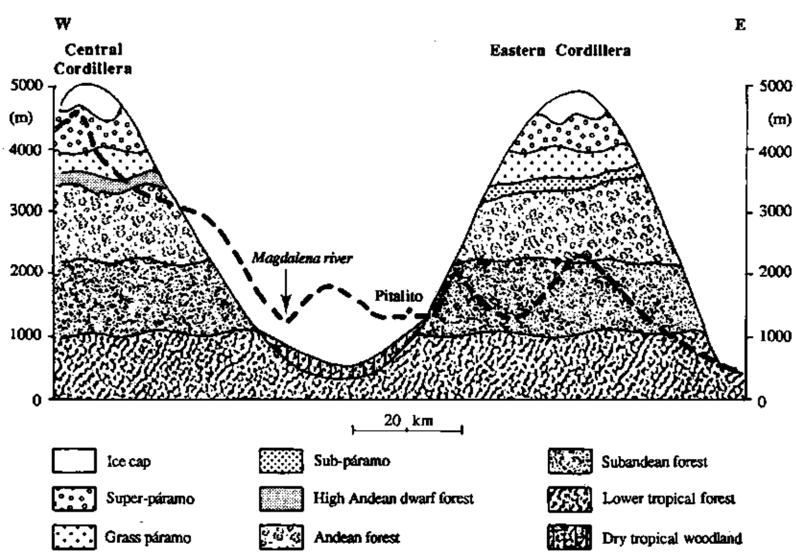


Fig. 9.
Modern altitudinal distribution of the vegetation belts of the Central and Eastern Cordillera in Colombia (after Van der Hammen 1974). The dashed bold line illustrates the mountain topography at the latitude of the Pitalito Basin.

A. The neotropical forest	B. The transitional zone	C. The páramo
1. The lower tropical forest	4a. High Andean dwarf forest	5. The grass páramo (or páramo proper)
2. The sub-Andean forest	4b. The subpáramo	6. The superpáramo
3. The Andean forest		

A. The neotropical forest (0-3500 m)

1. The lower tropical forest belt (c. 300-1000 m)
In this lower tropical vegetation belt a (very) humid and a dry type of vegetation may be distinguished. Average annual temperature is higher than 24°C for both types.
-The (very) humid tropical forest belt can be found at the foot of the eastern side of the Eastern Cordillera. Mean annual precipitation ranges between 4000 and 8000 mm. The diversity in species is high and the internal structure of these forests is complex and consists of several strata. No dominant taxa are present. The upper stratum is dominated by tall trees (up to 50 m). These forests are characterized by the presence of *Bauhinia, Cecropia, Cedrela, Clusia, Croton, Ficus, Inga* and representatives of the Euphorbiaceae, Melastomataceae, Moraceae, Palmae and Rubiaceae. The type of forest is rich in epiphytes (orchids, bromeliads and ferns).
-The dry tropical woodlands are present along the eastern slopes of the Magdalena Valley. The average annual rainfall lies between 1000 and 2000 mm. Prominent taxa in this open type of vegetation are *Bursera, Eugenia, Guazuma, Hirtella, Miconia, Scheelea, Solanum, Spondias* and *Anthurium*. Other important contributors are representatives of the Amaranthaceae, Cactaceae, Euphorbiaceae and Gramineae. Bryophyta are almost absent.
Rangel & Lozano (1986) describe a community at an altitude of c. 900-1200 m with high trees of *Guarea guidonia, Perebea* sp. and *Randia spinosa* at the transition with the sub-Andean forest. In the lower arborescent story *Myrcia* aff. *paivae* is present. The opulent herbaceous layer covers c. 75%.
North of Neiva (Fig. 1) in the area named desierto La Tatacoa, climatic conditions lead to the existence of an open xerophytic vegetation (Rangel & Franco 1985). Average annual rainfall varies between 500 and 1000 mm. The community consists of a thorn scrub of columnar cacti and thin shrubs. Vegetation in this zone shows a very patchy distribution. Main constituents of this xerophytic vegetation are: *Pithecellobium, Randia aculeata, Croton, Solanum, Lemaireocerus* and other representatives of the Cactaceae.

2. The sub-Andean forest belt (c. 1000-2300 m)
The descriptions of this zonal forest belt are mainly based on unpublished relevés of Van der Hammen, Jaramillo and Murillo for the Eastern Cordillera and of Rangel, Cleef and Salamanca (in prep.) for the Central Cordillera. These relevés are concentrated around lat. 5°N. Mean annual temperature ranges between 19°-23°C for the lower zones and 14°-19°C for the higher ones.
-Annual precipitation in the humid sub-Andean forest belt varies between 1500 and 2700 mm. The most important elements are: *Ladenbergia, Hedyosmum* and *Morus* in

association with o.a. *Cassia, Cecropia, Cleome, Clusia, Cyathea, Miconia, Tournefortia,* Bombacaceae and Sapotaceae. At the upper limit (c. 2000-2300 m.) a *Clusia-Hieronima-Weinmannia* forest is present.

•*Nectandretum* forest dominates in less humid areas (annual rainfall varies from 1000 to 2700 mm). They are principally associated with *Hedyosmum* and *Chrysoclamys.* Other woody taxa are *Acalypha, Alchornea, Clusia, Piper, Rapanea* and *Trichilia.*

-Annual precipitation of the more arid sub-Andean forest ranges between 1000-2000 mm. Floristically, two types of sub-Andean *Inga-Cecropia* forests can be distinguished: an altitudinal lower *Inga-Cecropia* forest with *Vismia* and *Hieronima* and an upper *Inga-Cecropia* forest with *Quercus* (Kuhry 1988).

•*Inga-Cecropia* forest with *Vismia* and *Hieronima* (1200-1850 m). Dominant species are lacking. Characteristic taxa are *Acalypha, Alchornea, Croton, Hieronima* and *Vismia.* Characteristic lianas species belong to the families of Amaranthaceae, Bignoniaceae, Leguminosae, Malpighiaceae and Marcgraviaceae. Treeferns (Cyatheaceae) may attain a coverage of 20%. Except for the low sub-Andean regions (1000-1300m), Palmae hardly attain 10% of the coverage. Terrestrial Bryophyta are almost absent.

•*Inga-Cecropia* forest with *Quercus* (2050-2300 m). In this type of forest *Quercus humboldtii* may become the dominant taxon in the tree layer. Other important taxa are *Billia* and *Hedysomum.* The coverage of Palmae and Cyatheceae is low (c. 10%).

3. The Andean forest belt (c. 2300-3500 m).

In the region of the study area this zonal vegetation type together with the higher situated páramo belt is mainly present in the Central Cordillera (Fig. 9). Locally, only the lower zones of the Andean forest belt may occur in the Eastern Cordillera as the highest tops do not exceed 2600 m altitude. Based on the relevés of Rangel & Franco (1985) and Rangel & Lozano (1986) the Andean forest at the eastern side of the Central Cordillera could be subdivided into a low, middle and high altitudinal zone. The transition between the Andean forest and the páramo belt lies considerably higher on the humid slopes (c. 3600 m) than on the drier ones (c. 3300 m). Mean annual temperature ranges between 8°C and 16°C.

-At the relatively dry slopes the lower zone consists of a *Quercus humboldtii* forest and a community with *Hedyosmum* and *Billia* which is recorded at a higher elevation than the oak forests. The middle zone is characterized by a *Brunellia-Weinmannia pubescens* community. The high-Andean zone is dominated by *Weinmannia* forests. Mean annual precipitation varies between 1000-1500 mm.

•*Quercus humboldtii* forest (c. 2200-2700 m). *Quercus humboldtii* is the absolute dominant taxon in the arboreal stratum (coverage c. 40%) accompanied by *Brunellia,Clusia, Ladenbergia* sp., *Miconia* and *Prunus.* In the understory appear *Alchornea, Ardisia, Eugenia, Hedyosmum, Weinmannia glabra* and *Viburnum.* Treeferns (Cyatheaceae) are most common. *Quercus* trees are usually covered with numerous epiphytes, bryophytes and lichens. Also numerous are parasitical elements belonging to the family of the Loranthaceae.

Oak forests are described by Lozano & Torres (1974) and Kappelle (1987). They reported the occurrence of *Quercus* down to 1000 m altitude at the eastern slopes of the Magdalena Valley.

•*Hedyosmum* community (c. 2600-2700 m). Besides *Hedyosmum huilense* other important contributors are *Billia, Clethra fagifolia, Clusia, Persea, Weinmannia glabra* and various species of Lauraceae. In the understory several species of the Palmae are dominant. Bark-growing briophytes are abundant.

•*Brunellia-Weinmannia pubescens* forest (c. 2600-3000 m). The most important arborescent taxa are *Brunellia macrophylla,Weinmannia pubescens, Clethra* aff. *revoluta* and *Miconia stipularis* and among the smaller trees *Hedyosmum* cf. *bonplandianum* and *Saurauia*. The presence of epiphytes (*e.g. Elaphoglossum*) is abundant. Coverage of treeferns is low (c. 10%).

•*Myrica-Weinmannia subvelutina* forest (c. 3000-3300 m). The coverage of arborescent taxa is high (c. 75%) with *Myrica pubescens* and *Weinmannia subveltina* as characteristic elements. Other prominent taxa are *Weinmannia vegasana, Clusia multiflora* and *Miconia latifolia*. In the understory several species of the Ericaceae are present. In the herb layer (c. 10%) *Begonia, Blechnum* and *Pilea* are recorded.

-At the more humid slopes the forest line reaches higher altitudes. This phenomenon is described by Van der Hammen & Cleef (1985) and is in accordance with the findings of Rangel & Franco (1985). Three different communities could be discerned. The lowest zone is dominated by a *Hieronima-Solanum* community, the community of the middle zone is very well comparable with the *Brunellia-Weinmannia pubescens* forest of the drier slopes. The upper zone is dominated by *Weinmannia* forests. Mean annual precipitation lies between 2000 and 2500 mm.

•*Hieronima-Solanum* forest (c. 2300-2600 m). The most important arborescent taxa are *Solanum inopinum, Hieronima colombiana, Miconia* sp. High percentages (c. 70%) of the understory coverage with, among others, *Meriania, Sapium, Saurauia* and Cyatheaceae. Epiphytes coverage is c. 20%.

•The *Brunellia-Weinmannia pubescens* forest ranges between c. 2800-3000 m altitude.

•*Weinmannia mariquitae* forest (c. 3200-3500 m). Besides *Weinmannia mariquitae* this forest is characterized by the presence of *Rapanea dependens*. Other taxa are *Diplostephium* and occasionally *Miconia* sp. The coverage of Gramineae - *Chusquea, Neurolepis, Rhynchospora* - is low (10-15%).

B.The transitional zone (c. 3400-3900 m)

This zone forms the transition from a closed vegetation type towards a more open one. Two types of zonal vegetation can be distinguished: the High Andean dwarf forest (HAD-forest) and the subpáramo. The HAD-forest is specific for the Central Cordillera whereas the subpáramo vegetation is typical for the Eastern Cordillera. Both types differ in their physiognomy and floristics. In view of the dynamics of the floristic system it must be emphasized that both vegetation types may replace one another locally. Mean annual temperatures fluctuate from 10°C to 5°C and mean annual precipitation between 1000-1500 mm.

4a. The High Andean dwarf forest belt (HAD-forest)
This forest is dominated by small trees between 3 and 5m in height which only form one story. The cover by bryophytes and epiphytes is generally higher than in the Andean forest belts.
•The *Weinmannia cohensis* forest (c. 3400-3700 m) is a typical high-Andean forest which grade into the páramo belt. The forest is dominated by small trees, scrubs and high herbs. Floristically, it contains elements from the páramo as well as from the Andean forest belt. Besides *Weinmannia cohensis* (c. 50%) other characteristic taxa are *Ilex*, *Diplostephium*, *Gynoxis*. Important contributors of the lower stratum are *Vaccinium* and *Neurolepis* and the herb *Rhynchospora* (c. 10%). The coverage of terrestrial bryophyta is high (c. 30%).
•An *Ilex-Gynoxys* community is described for the upper zone (c. 3600-3700 m: Rangel & Franco 1985; Sturm & Rangel 1985). Other important contributors are *Diplostephium*, *Senecio* and *Miconia*. At poorly drained places *Sphagnum* sp. was recognized. The bryophytic ground layer attains a coverage of 60-70%.
Forest line stands may contain a number of taxa characteristic for the upper Andean forest belt such as *Brunellia*, *Ilex*, *Hesperomeles*, *Oreopanax*, *Rapanea* and small trees of *Weinmannia*.

4b. The subpáramo vegetation belt
Based on floristics and physiognomy Cleef (1981) recognized a lower subpáramo dominated by dwarf trees and shrubs and an upper subpáramo dominated by dwarf shrubs. Most characteristic taxa for the subpáramo are *Hypericum*, *Monnina*, *Myrica*, *Polylepis* and representatives of the Compositae, Ericaceae, Melastomataceae and Rosaceae. The herb layer may cover 15-50% with species such as *Calamagrostis effusa*. Terrestrial bryophytes may cover as much as 70-100%. On (per)humid mountain sides bamboo species prevail, *e.g.* *Swallenochloa tessalata*, and the bryophytic ground layer is dominated by mosses and liverworts.

C. The páramo (3400-4700 m)

5. The grass páramo vegetation belt (c. 3400-3900 m).
The grass páramo forms an open type of vegetation in which the Gramineae and the Compositae form the most important taxa. Towards the upper limit the vegetation cover is usually not completely closed. Mean annual temperature ranges between 3°-6°C and the mean annual precipitation lies around 1500 mm in the humid areas and around 1000 mm in the drier ones. *Calamagrostis effusa* is the most prominent páramo bunchgrass and cover 50-80%. A conspicuous feature is *Espeletia hartwegiana* ssp. *centroandina* (Compositae) which forms high stemrosettes. Other important contributors are *Castilleja fissifolia*, *Satureja nubigena*, *Halenia*, *Cerastium*, *Hypericum*, *Geranium*, *Senecio*, *Oreobolus* and *Jamesonia*. In humid regions the grasses may be completely replaced by bamboo (*Swallenochloa* sp.).

6. The superpáramo vegetation belt (c. 4000-4700 m).
The superpáramo reaches up to the nival belt and is characterized by a sparse and

irregular distributed vegetation cover. Due to its high elevation the zone is daily exposed to nightfrost. Especially in the upper part of the superpáramo plant growth is restricted to sheltered habitats. The mean annual temperature is below 3°C and the mean annual precipitation usually do not exceed 1000 mm. Most important taxa are *Calandrina*, *Cerastium*, *Draba*, *Lachemilla*, *Loricaria*, *Senecio* and *Valeriana*. The coverage of bryophytes and lichens is relatively high.

2.4.2 AZONAL VEGETATION

The azonal vegetation types depend largely on local edaphic or climatic conditions and are often not bound to a single belt but may extend over several altitudinal zones. Van Straaten (1989) produced a map of the local vegetation present in the La Coneca area, a badly drained region in the NE part of the Pitalito Basin. He distinguished four vegetation types in this area which are here briefly described.

-*Vegetation type 1* This vegetation type is characterized by a dense cover (c. 50-90%) of tall (2-3 m high) *Tibouchina* shrubs (Melastomataceae) and a low cover of Cyperaceae (c. 5-30%). Other characteristic shrubs are *Piper* sp., *Ludwigia* cf. *octovalvis*, and *Baccharis nitida*. The herbcover varies from 0% under dense shrub stands up to 100% in an open vegetation type. Some herbaceous elements are Cyperaceae (a.o. *Fuirena incompleta* and *Cyperus* cf. *haspan*), Gramineae (*e.g. Panicum helobium*, *Leersia hexandra* and *Erianthus trinii*) and *Begonia* sp. The fern cover is generally low (to c. 30%) but on well-drained sites where the peatlayer is thin or lacking *Thelypteris* sp. is abundantly present. This vegetation type is generally present at the rim of the moor where a thin peat layer (mostly less than 150 cm thick) lies on top of clayey material and where eutrophic conditions prevail.

-*Vegetation type 2* Characteristic for this type of vegetation is the abundance in the herbcover of Cyperaceae (c. 20-50%; mainly *Fimbristilis complanata*) in association with Gramineae (c. 20-50%; mainly *Eleocharis nodulosa* and *Coelorhachis aurita*). The dense cover of *Tibouchina* sp. in some stands is explained by the nutrient supply from the underlying clayey layer. In these cases the covering peat layer is relatively thin (< 80 cm). Other characteristic shrubs are *Ludwigia* cf. *octovalvis* and *Baccharis nitida*. *Eriochrysis*, *Xyris* and *Lipocarpha humboldtii* are characteristic elements in the herbcover. *Begonia patula* and *Phyllanthus stipulatus* are also present. This vegetation type is generally found in the central part of the moor where the peat layer is thick (c. 100-250 cm thick) and where oligotrophic conditions prevail.

-*Vegetation type 3* Dense stands of tall (2-6 m high) *Baccharis nitida* scrubs are characteristic for this type of vegetation (total cover 40-60%). Another, less dominant element is *Tibouchina* sp. (20-40%). Cyperaceae (mainly *Fimbristilis complanata*), Gramineae (mainly *Erianthus trinii*) and *Relbunium hypocarpium* (Rubiaceae) are characteristic elements in the undercover. This vegetation type was found in the central part of the moor at moderately drained sites with mesotrophic to eutrophic conditions. The thickness of the peat layer varies between 30 and 300 cm.

-*Vegetation type 4* Characteristic is the low cover (total of c. 15%) of the poorly growing scrub layer: *Tibouchina* scrubs are less than 1.5 m tall. The herblayer covers 75 to 90% and is dominated by *Fimbristilis complanata* and *Hydrocotyle bonplandi*. *Sobralia* *rosea* and *Epidendrum secundum* (both Orchideaceae), *Begonia patula* and *Xyris* sp. are also present. The cover by mosses is high (20-40%). This open vegetation type is present along the northern mountain front and has a raised bog character in which oligotrophic conditions prevail. The thickness of the quivering peat layer varies between 2 and 4 m.

Descriptions of other azonal vegetation types in more or less similar ill-drained areas at higher altitudes are given by Cleef & Hooghiemstra (1984), Rangel & Aguirre (1986) and Wijninga & Vink (1986).

Important are the largely azonal *Alnus jorullensis* forests which nowadays are mainly present between c. 2550 m and c. 3100 m altitude (Cleef & Hooghiemstra 1984). These azonal forests are often found along water courses. Further upslope they grade into the zonal *Weinmannia tomentosae* forests. In the past they may have covered large, flat-lying, ill-drained areas of the highplain of Bogotá and also might have dominated on the alluvial plain of the Pitalito Basin.

2.5 Archeology

The abovementioned vegetation types refer to undisturbed natural stands of vegetation. However, human influence may interfere with the natural development of zonal as well as azonal vegetation types.

Due to its climate and flat topography the Pitalito Basin (also known as the Valley of Laboyos) forms an exquisite site for early human occupation. Man could easily cultivate land under relatively mild climatic conditions. The only problem man had to cope with was the irrigation of the ill-drained eastern areas. Archeological studies along the lower Cauca and Magdalena rivers in northern Colombia (Plazas *et al.* 1988) show that man was capable to deal with such problems already some 2500 years ago. Early men constructed canals perpendicular to the natural course of the river. Smaller ditches of prehispanic age are found in the El Dorado Valley in the Central Cordillera of Colombia (Bray *et al.* 1985). These ditches, which form a grid, are designed to control subsurface water; during dry seasons ground water flow into the valley from ill-drained areas and during wet seasons the water table is kept as low as possible. These ditches were constructed around 1200 A.D. Ditches with a typical rectangular pattern are also visible in the areal photographs of 1963 of the Pitalito Basin. They possibly form a relict of an early irrigation or drainage system and their presence suggests that man occupied the basin centuries ago.

Early occupation in the Pitalito Basin is confirmed by the work of Llanos (1989) who found evidence for first human activity around 3000 years B.P. He studied a settlement of prehispanic age just north of Pitalito along the Cálamo rivulet (Appendix I). The classification he used was based on the work of Gomez & Cubillos (1979) who studied the famous National Archeological Park of San Agustin. This park is located just about 25 kms east from the Pitalito Basin. The following is based on the study of Llanos (1989). He studied a 135 cm deep excavation in a terrace. Based on the findings of numerous ceramics he distinguished three periods of prehispanic occupation and classified them as follows:

Período Formativo (≈3000 to ≈2200 years B.P.)
In a buried organic-rich soil many artefacts of this period were found. The soil at 120 cm depth yields a ^{14}C-age of 2200±120 years B.P. After this period the site was left. Llanos suggests that men went to another place outside the Pitalito Basin. This occupation is partly contemporaneous with an early phase of the San Agustin culture of which the oldest date is about 2500 years B.P.

Período Clásico Regional (≈1600 to ≈1100 years B.P.)
During this period the site was occupied three times. The terrace was frequently inundated due to flooding of the Cálamo rivulet and men were forced to leave the terrace. The sediments were made up of sand and sandy clay and contain material of volcanic origin. The 95 cm level yields a ^{14}C-age of 1430±370 years B.P. This level is considered to form the oldest part of this period.

Período Reciente (≈900 to≈300 years B.P.)
The last period of occupation has a relative age of 600 years B.P. Llanos based this dating on the type of ceramics and a presumed amelioration of the climate around this time (Piñeros 1988). Around this period inundation came to an end and man habitated the terrace again after a long period of abandonment.

Human influence on natural vegetation stands
About 2000 years ago human interference with natural vegetation stand became important and is registered in pollen diagrams (*e.g.* Van der Hammen 1962; Kuhry 1988; Piñeros 1988). Man deforested the hills and a more open secondary forest type occupied the affected sites. The hills surrounding the Pitalito Basin are nowadays covered by grasslands and cropland and only small patches of sub-Andean forest with *Cecropia* and *Inga* can be recognized.

The edaphic conditions in the basin are largely determined by the ill-drained character of several sites, especially in the eastern region (Espinal & Petrelli 1966). Therefore, the majority of natural stands in the basin itself are of an azonal character which is stated by the findings of Van Straaten (1989). Nowadays large areas are drained artificially through small channels leading to a complete change of edaphic conditions.

The most important cultivation products around and in the Pitalito Basin are: coffee, several types of fruit (bananas, oranges, guyaba), mais, millet and cacao. The poorly drained regions are used as grasslands for cattle breeding.

PHOTOS

Photo 1
Stereopair of flight M-1379/no.39773-39774 (1966): the area around Pitalito. A/ the re-entrants of the southern mountain front; b/ the paleochannels which now appear as sandy ridges in the eastern part of the basin; C/ Guarapas valley floor; 2/ the approximate location of the fault which borders the southern basin margin (see also Fig. 12).

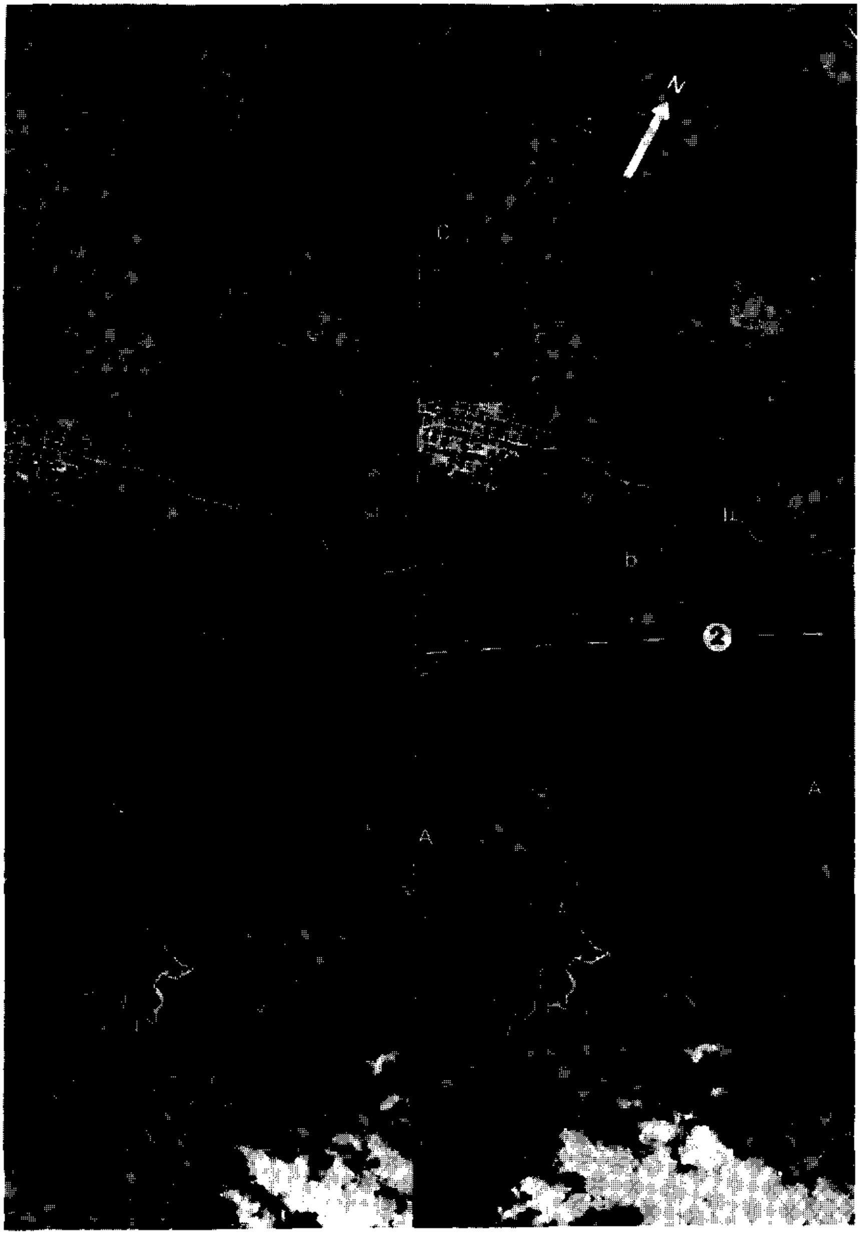

N
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Photo 2
Stereopair of flight M-1261/no.25759-25758 (1963): the area around La Coneca. A/ the straight morphology of the northern mountain front with small colluvial cones; b/ the paleochannels which now appear as sandy ridges in the eastern part of the basin; C/ The ill-drained La Coneca area; D/ the large alluvial fan made up by reworked material from the Guacacallo Formation and deposited by the Regueros rivulet 1/ the approximate location of the dextral strike-slip fault which borders the northern basin margin (see also Fig. 12); 2/ location of core PIT 2.

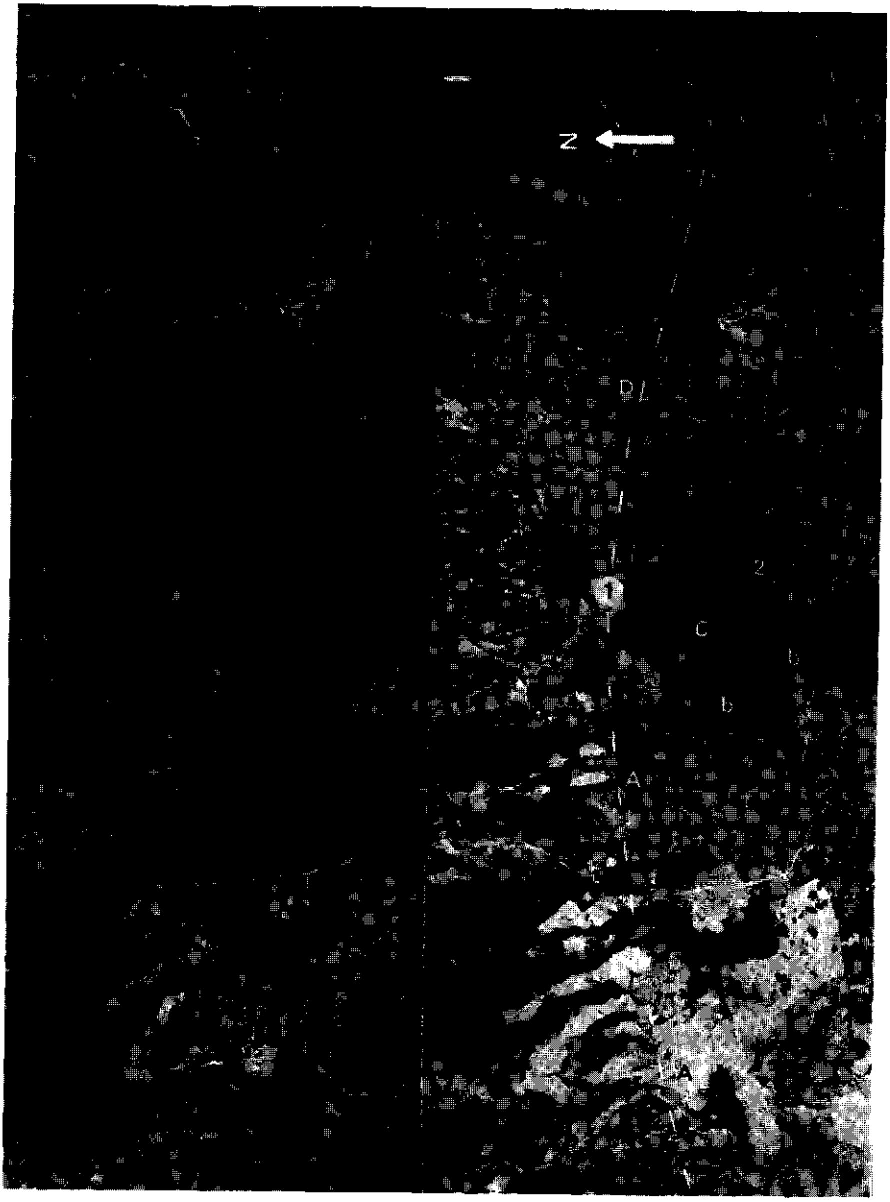
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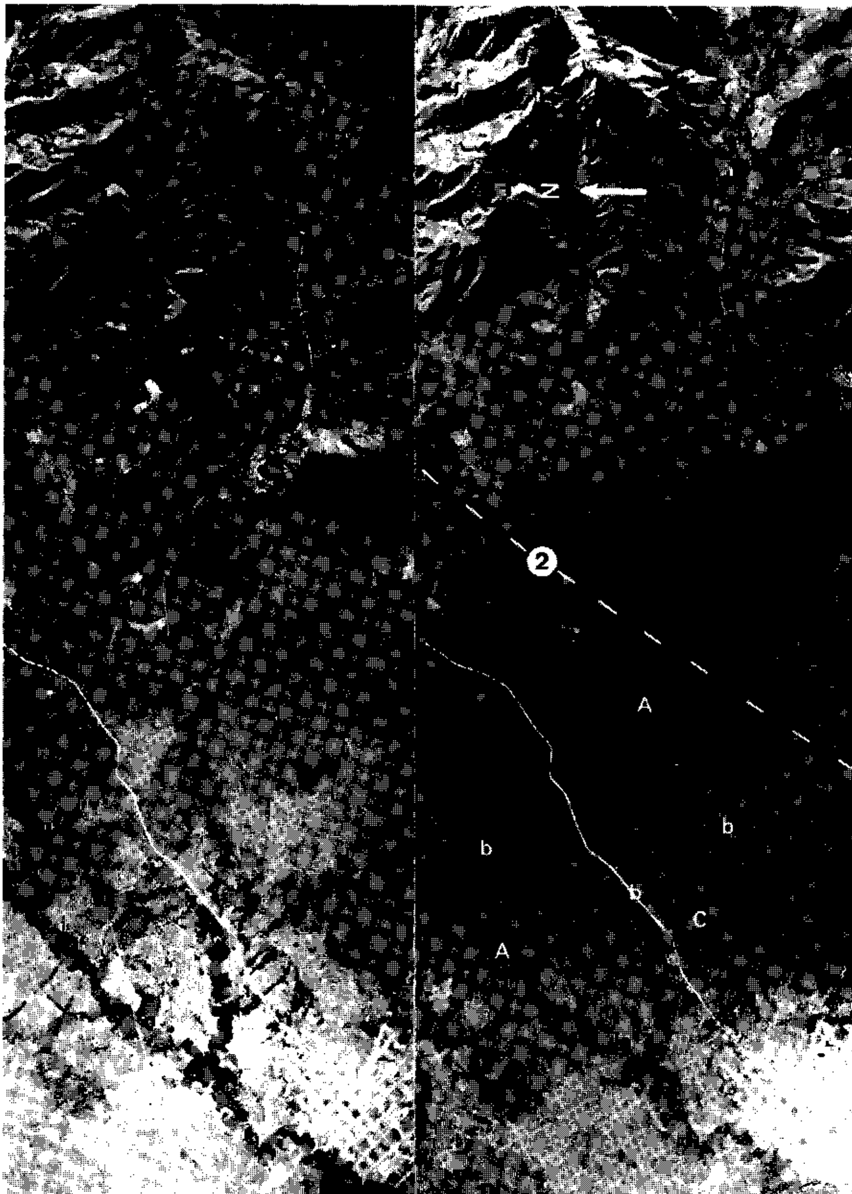
Photo 3
Stereopair of flight C-1308/no.116-115 (1970): the area NE of Pitalito. A/ W to SW-heading rivulets which make part of the actual drainage system and slightly dissect the basin infill; b/ the paleochannels which now appear as sandy ridges in the eastern part of the basin; C/ N-heading rivulets belonging to a fossil drainage system, just north of C this rivulet turns towards the west by means of a sharp angle; 2/ the approximate location of the fault which delineates the southern basin margin (see also Fig. 12).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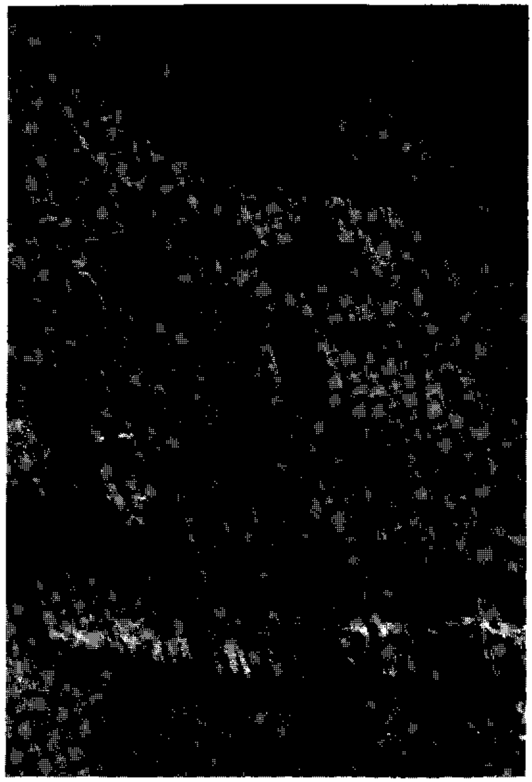


Photo 4
The steep valley walls along the Guarapas river which are predominantly made up by massive layers of clay. Each layer is about 150 cm thick and is seperated by a sandy or organic-rich stratum.

Photo 5
The coarse-clastic banks along the Guachicos river.

Photo 6
Tilted basin sediments along fault no. 4 (Fig. 12). At the left-hand of the picture lignitic material dip towards the NW whereas the clayey sediments at the right hand dip towards the SE. On top, coarse fluvial sediments are deposited by the Guachicos river.

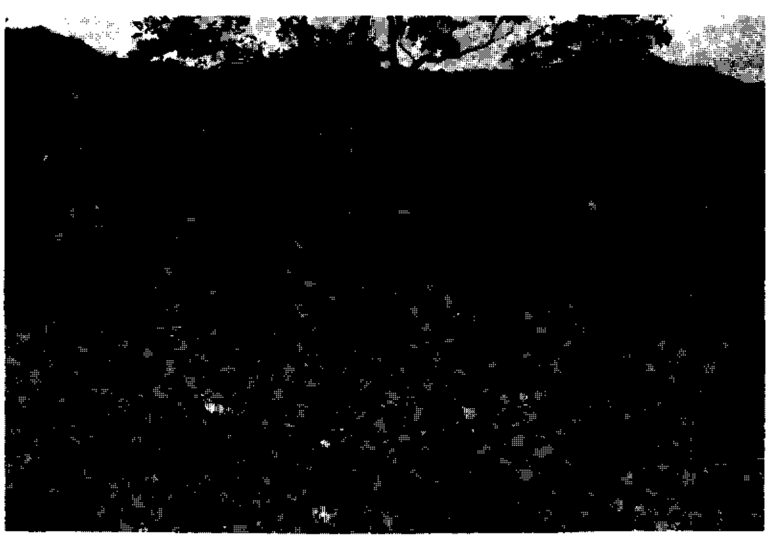


Photo 7
Parallel view with a coarse-grained ridge at the left gently grading to the lower lying peat-rich flood basin at the right. In Figure 23 this morphological feature is outlined in terms of grain size and lithofacies.

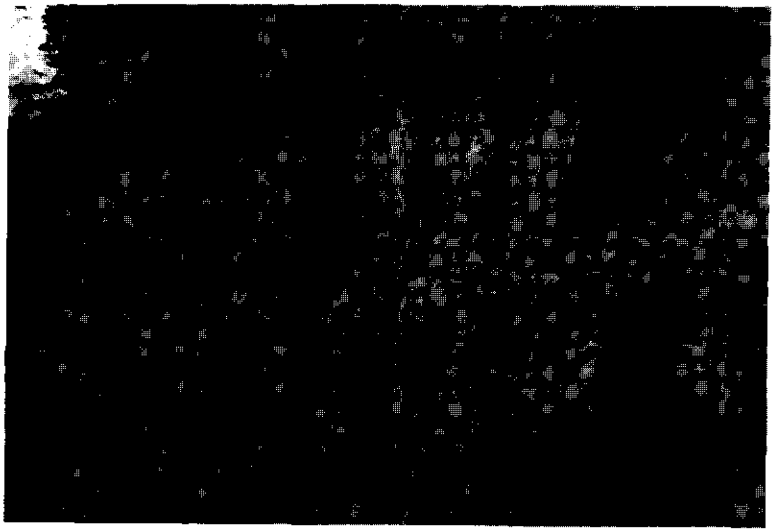


Photo 8
Massive clayey/silty basin deposits along the Guarapas river (Section 20B). Note the intercalations of dark-coloured, thin but continous layers of organic debris which are interpreted as distal crevasse splay deposits. The stratigraphical position is between 17 m and 20 m depth. The shovel is c. 1.50 m high.

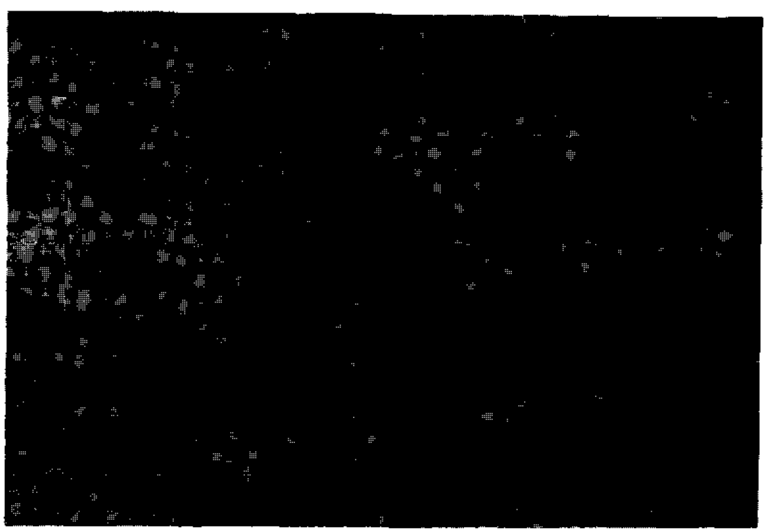


Photo 9
A 10 to 20 cm thick volcanic ash layer which served as a marker horizon in the western part of the basin. It is characterized by a dark-coloured stratum which is sandwiched between two light-coloured strata. The knife is c. 25 cm high. See also Fig. 25.

Photo 10
View from the northern mountain front over the eastern part of the Pitalito Basin in southern direction. In the foreground a part of La Coneca is visible.

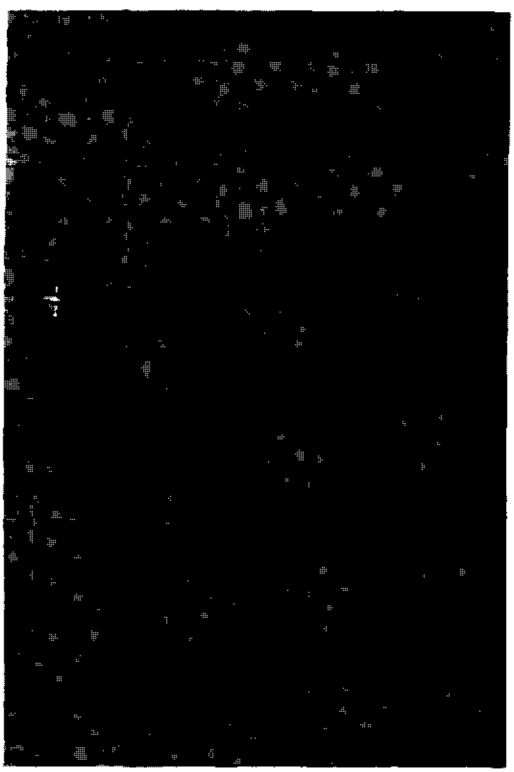


Photo 11
Large-scale, heterogenous cross-stratification with grain-size variation in the foresets ranging from fine sand to clay. The vertical-oriented structures at the lower left side evidence bioturbation. The knife is c. 25 cm high.

CHAPTER 3

BASIN FLOOR MORPHOLOGY AND CHARACTERISTICS OF THE NON-EXPOSED SEDIMENTARY INFILL

This chapter deals with the structural setting of the Pitalito Basin and the possible influence of tectonic activity on changes in the sedimentary processes. Tectonic control, *e.g.* basin subsidence and basement uplift, may play a significant role on alluvial architecture (*e.g.* Allen 1978, 1979; Kraus & Middleton 1987) and internal asymmetry in basement morphology can lead to spatial differences in sedimentological processes in the basin itself (Ballance 1980; Bridge & Leeder 1979; Read & Dean 1982).

The basement morphology and the possible presence of fault systems in the Pitalito Basin were studied with the help of a gravity survey whereas the non-exposed sediments were examined by means of geoelectrical methods. As the metric accuracy might be sometimes unsatisfactory the geophysical exploration only gives a global description of the above mentioned non-exposed features. The outcome of this study may shed some light on the differentation in the exposed surface and near-surface sediments which will be discussed in Chapter 4.

3.1 Gravimetrical survey

3.1.1 MATERIAL AND METHODS

In August 1987, a gravimetrical reconnaissance survey was carried out in order to gain an approximate picture of the geometrical structure of the basement. The gravimetrical method is based on the attraction of two masses which are considered here as points. When such a point mass is placed in the vicinity of a body it will experience an acceleration towards this body caused by the gravitational field of the last. The mass of the earth can cause the same reaction to an external point which experiences a gravitational acceleration towards the centre of the earth. This study addresses to the differences in accelerations (Δg) between an observation point relative to a base station both lying in the study area (Appendix II). Such relative measurements yield information about local changes in mass density within the earth as well as about the surfaces that delimit regions of differing mass density.

The unit of gravity anomalies, being differences in acceleration, is $\mu m\ s^{-2}$ also called gravity unit (g.u.). The most used unit in practice however, is called gal (cm s^{-2}) or milligal (=10 g.u.). The very small changes in gravity can be measured by a gravimeter:

an instrument which holds a point mass (=external point) in an unstable equilibrium. This very sensitive balance is displaced due to a change in the gravity which can be measured by the force necessary to return the element to its equilibrium position. With some assumptions (see below) this method may give outcome of the distance between the surface and basement so that estimations about basin depth and basin floor morphology can be given.

The instrument used was a Texas Instruments Worden gravimeter no. 1118, with a small dial constant of 0.05142 at 65°F. Since no accurate data on the topographical heights of the surrounding hills were available, the measurement locations were restricted to the relatively flat basin floor (Appendix II).

3.1.2 CALCULATION OF THE RELATIVE BOUGUER ANOMALIES

The relative differences in gravity must be corrected appropriately to take account of several factors: geographical latitude, elevation, local relief, *etc.* The corrected values, called relative Bouguer anomalies, generally differ from the measured relative ones.

As there was no information about regional anomalies, a local base station (lat. 1°52' N) within the study area was assigned an anomaly of 0.0 mgal (point 1; Fig. 10 and Appendix II). The instrument readings have been corrected for instrument drift and tide influences by means of a linear interpolation in time (the time interval never exceeded 2 hours). Furthermore, data were also corrected for geographical latitude, the elevation relative to the base station ('free-air correction') and the additional attraction exerted by the surplus of material relative to the base station ('Bouguer correction').

The latitude correction has been applied by using equation (1):

$$\frac{\mathrm{d}\,g_1}{\mathrm{d}\,s} = \frac{\mathrm{d}\,g_1}{R_\mathrm{e}\,\mathrm{d}\phi} \approx \frac{\mathrm{d}\,g_1}{R_\mathrm{eq}\,\mathrm{d}\phi} \approx 0.081 \sin(2\phi)\ [\mathrm{mgal/100\ m}] \tag{1}$$

where g_1 is the gravity at latitude l, s is the horizontal north-south distance from the base point, R_e is the earth radius, R_{eq} is the earth radius at the equator and ϕ the latitude. The result must be subtracted from or added to the measured gravity difference accordingly as the station is on a higher or lower latitude than the base station.

The free-air correction using equation (2):

$$\frac{\mathrm{d}\,g_{\mathrm{fa}}}{\mathrm{d}\,R_\mathrm{e}} = \frac{-2\gamma M_\mathrm{e}}{R_\mathrm{e}^3} \approx \frac{-2g}{R_\mathrm{eq}} \approx -0.0372\ \ [\mathrm{mgal/m}] \tag{2}$$

where M_e is the earth mass and γ is the universal gravitational constant.

The Bouguer-correction using equation (3):

$$\frac{\mathrm{d}\,g}{\mathrm{d}\,R_\mathrm{e}} \approx \frac{\mathrm{d}\,g_\mathrm{B}}{\mathrm{d}\,R_\mathrm{eq}} = 2\pi\gamma\sigma\ \ \mathrm{mgal/m} \approx 0.04188\sigma\ \ [\mathrm{mgal/m\ of\ elevation}] \tag{3}$$

where σ is the density of the rock material. The determination of rock density of the formations is based on commonly known values (*e.g.* Telford *et al.* 1976, Tables 2.2 to 2.5; Parasnis 1986, Table 3.1). The number of involved formations has been reduced to two: basin infill and basement. The latter is considered to be uniform without differences in density. For the basin infill (unconsolidated sediments) a mass density of 2.0 g/cm^3 is proposed and for the basement 2.5 g/cm^3 (intermediate volcanic rocks of the Saldaña Formation).

At the Department of Landsurveying, Photogrammetry and Remote Sensing, Wageningen Agricultural University, the topographical heights of the datapoints have been calculated from digitized areal photographs of the area (de Wit 1989). This resulted in a contour map with an interval of 5 m and an accuracy of ± 5 m. The maximum error in the station gravities due to the error in the estimation of the elevations is in the order of ± 2.0 mgal. Because of this relatively large error no terrain corrections have been applied and the values are given in mgals without decimals. The measured and corrected anomalies of the datapoints as well as their location and elevation are shown in Appendix II.

The final corrected relative Bouguer anomalies (g_{corr}) which are presented in Figure 11 have been calculated based on latitude 1°52' N for ϕ and 2.0 g/cm^3 for rock density at the surface. These figures together with the combined corrections yield equation (4):

$$g_{corr} = g_{obs} - 0.05274d + 0.22474h \text{ [mgal]} \qquad (4)$$

where g_{obs} is the measured gravity anomaly, d is the distance in kilometers north of the base station and h the elevation in meters above the base station.

For more detail concerning the gravitational method reference is made to Telford *et al.* (1976) and Parasnis (1986).

3.1.3 INTERPRETATION

The relative Bouguer anomalies are presented in a Bouguer anomaly map with a contour interval of 2 mgal (Fig. 10). Several sections have been modelled and projected onto a vertical plane with the polygon method according to Talwani *et al.* (1959), using the algorithm described by Won & Bevis (1987). Three of these sections are depicted in Fig. 11: section A-A' is most representative for the eastern part of the basin, section B-B' for the western part, and section C-C' illustrates the relation between the western and eastern part. The small-scale character of the survey made it unnecessary to separate the regional trend from the local variations in the models.

The most conspicuous feature is the strong negative anomaly of 20 mgals in the northeastern part of the basin. If we assume that the anomaly can be entirely ascribed to basin geometry, a minimum depth of c. 1200 m has been calculated for this part of the basin using a maximum density contrast of 0.5 g/cm^3 between the basement and the sedimentary infill (Fig. 11; section A-A'). The steepest side is found along the northern margin of the basin whereas the opposite southern margin is less steep.

The values in the western part of the basin (*viz.* west of the Guarapas river) are lower than in the eastern part (Fig. 10) and reach a total value of about 5 mgals along both the

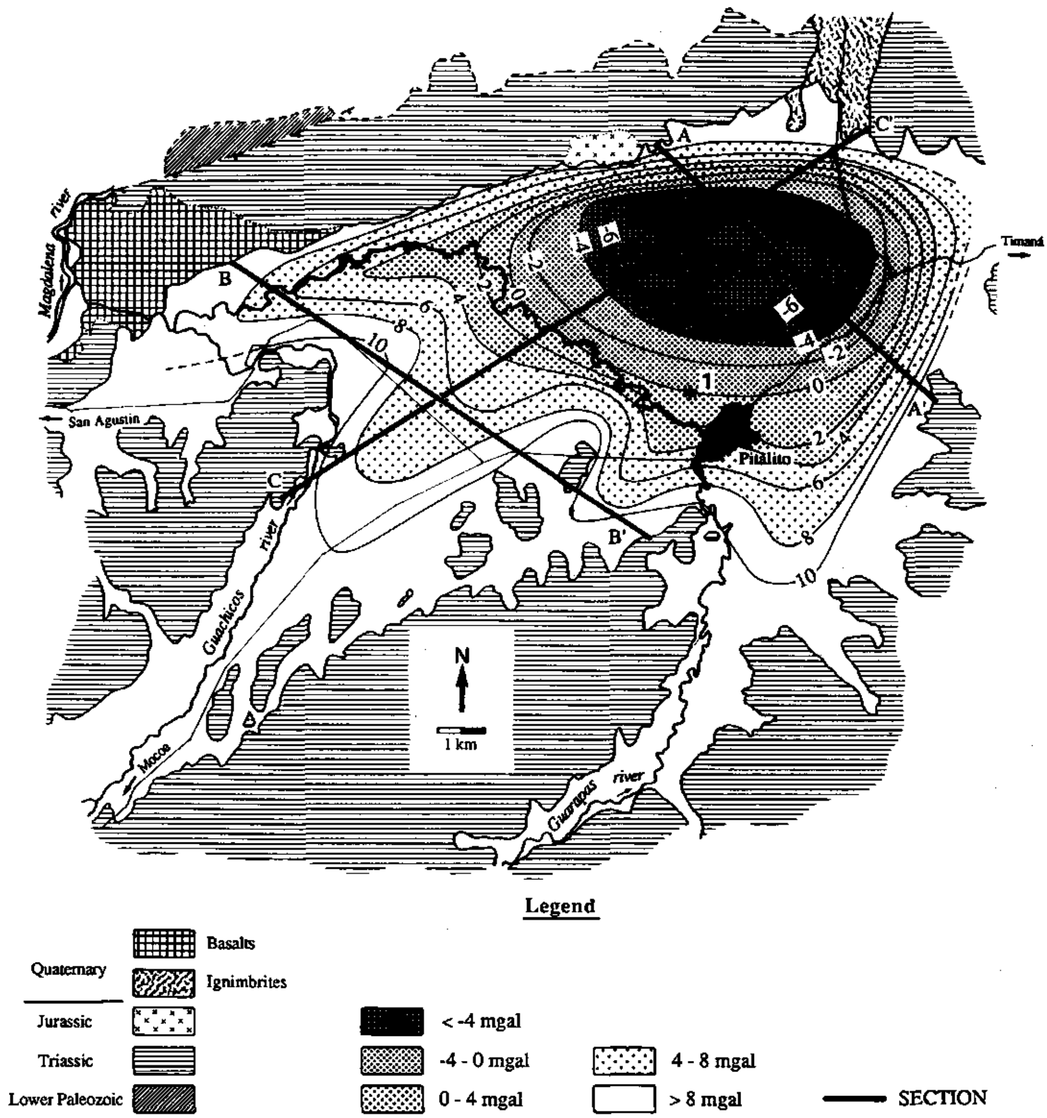
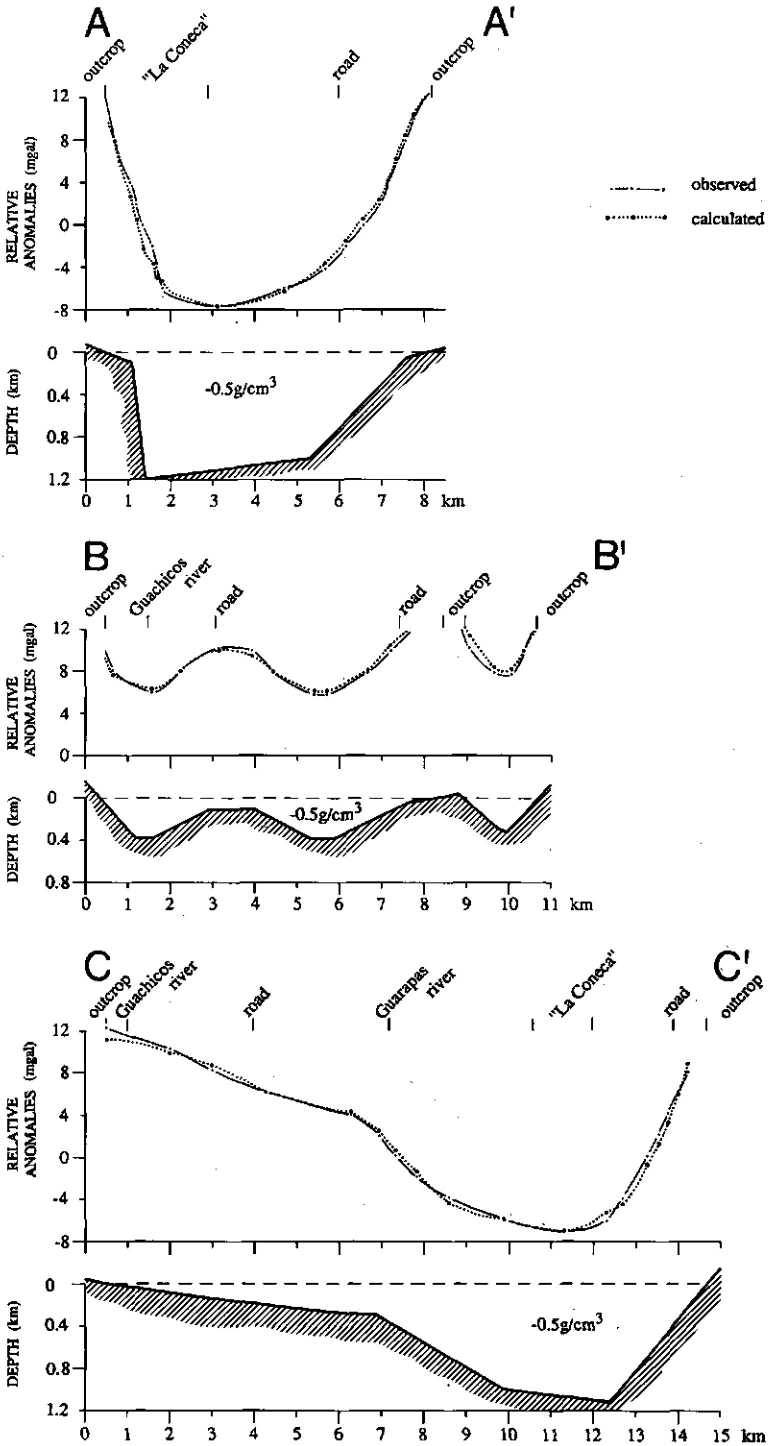


Fig. 10.
Bouguer anomaly map of the Pitalito Basin relative to the basepoint (*). Contours in milligal. The contour map is based on 100 gravity stations all restricted to the flat topography of the basin (see Appendix II).

Fig. 11. (next page)
Observed and calculated gravity profile and interpreted basement relief of the Pitalito Basin along three sections: A-A', B-B' and C-C', respectively. For the calculation a density contrast of 0.5 g/cm^3 between sedimentary infill and basement was used. See Figure 10 for the location of the sections.


A
A'
outcrop
"La Coneca"
road
outcrop
RELATIVE ANOMALIES (mgal)
DEPTH (km)
-0.5g/cm3
observed
calculated
km
B
B'
outcrop
Guachicos river
road
road
outcrop
outcrop
-0.5g/cm3
C
C'
outcrop
Guachicos river
road
Guarapas river
"La Coneca"
road
outcrop
-0.5g/cm3

northern and southern margin. A maximum depth of c. 400 m is calculated for the western part of the basin and locally the basement rises to a depth of c. 80 m below the surface (Fig. 11; section B-B').

The Bouguer contour lines tend to close in the southern tributary valleys (Fig. 10). It remains uncertain if the same holds for the eastern and the northwestern part of the study area.

From west to east the Bouguer anomalies increase and show a relatively sharp decline at the latitude of the Guarapas river. Assuming again that the anomalies are entirely ascribed to basement geometry this would imply that west of the Guarapas river the basin floor generally slopes towards the east, (Fig. 11; section C-C') while at the line of the Guarapas river the basement shows a sharp drop from 300 m to a depth of c. 1100 m along the entire NW/SE stretch. In the east the basement continues at a depth of c. 1100 m and descends further to a depth of about 1200 m in the northeast.

Based on the gravimetrical data and morphological features the following principal tectonic structures are recognized in the Pitalito Basin and outlined in Figure 12.

-Northern margin The northern margin is delineated by an active fault system with a W/E-orientation (no. 1; Fig. 12). The southern part forms the downwarped block. The vertical displacement along this northern fault system is about 1200 m in the eastern part of the basement and decreases to c. 400 m further west (Fig. 11; sections A-A' and B-B'). Its recent activity is expressed in the straight morphology of the mountain front (Photo 2), its triangular facets, the faulted colluvial cones in the west and tilting of younger alluvial sediments along the fault. Offsets of several streams draining the hills along the northern margin of the basin, suggest right-lateral movement along the fault. The eastern continuation of the northern fault remains uncertain, though interpretation of the gravimetrical data (Fig. 11; section C-C') suggests a curvature of the fault system (Fig. 12).

-Southern margin The southern basin margin is bounded by *en echelon* NE/SW-oriented fault zones (Fig. 12; no. 2 & 3). The mountain front along fault zone no. 2 shows numerous re-entrants (Photo 1) whereas the one along the western fault zone no. 3 has a straighter form. This might imply that the last-mentioned fault zone was active more recently than fault zone 2. Fault zone 2 converges with the northern fault system (no. 1) in the easternmost part of the basin and extends further east into the Timaná Valley. Along both fault zones the northern block has been downwarped. The differences in vertical displacement are considerably: along zone 2 it measures about 900 m whereas along fault zone 3 it is about 300 m (Fig. 11; section B-B').

Fieldwork revealed a third NE/SW-oriented fault (Fig. 12; no. 4) present in the NW-part of the basin. The fault branches from the northern fault zone and affects the basin infill (Photo 6). Apparently, the vertical offset along this fault was too small to be detected by the gravimetrical survey.

Where the Guarapas river diagonally crosses the Pitalito Basin, its course coincides with the abrupt drop in the basement, picked up by the gravimetrical survey (Fig. 11; section C-C') and suggests the presence of a fault with a NW/SE orientation (no. 5; Fig. 12). This fault dissects the Pitalito Basin into a shallow western part and a deep eastern part. In the north this fault abuts onto the northern fault system while in the south it delimits the *en echelon*-positioned fault zones 2 and 3. The vertical displacement along this fault is c. 700 m.

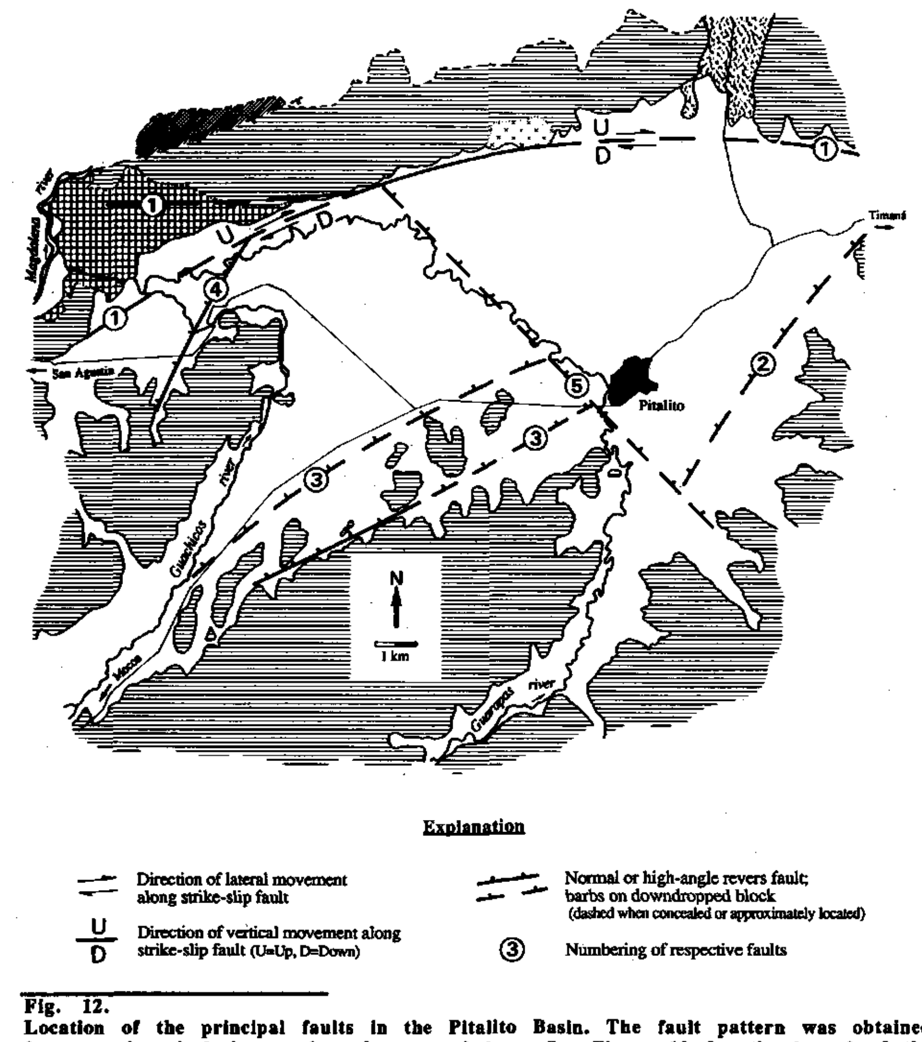


Fig. 12.
Location of the principal faults in the Pitalito Basin. The fault pattern was obtained from gravimetrical data and surface morphology. See Figure 10 for the legend of the surrounding area.

3.2 Geoelectrical survey

3.2.1 MATERIAL AND METHODS
A resistivity method based on the electrical properties of rocks was used to get an idea of the areal distribution and the type of sedimentary infill of the basin. Therefore, a total of 55 vertical electrical soundings (VES) has been executed. The object of electrical sounding is to determine the vertical resistivity distribution. The equipment used was an ABEM Terrameter with booster. The siting of the geoelectrical measurements (Appendix III) was based on the results of the gravimetrical survey.

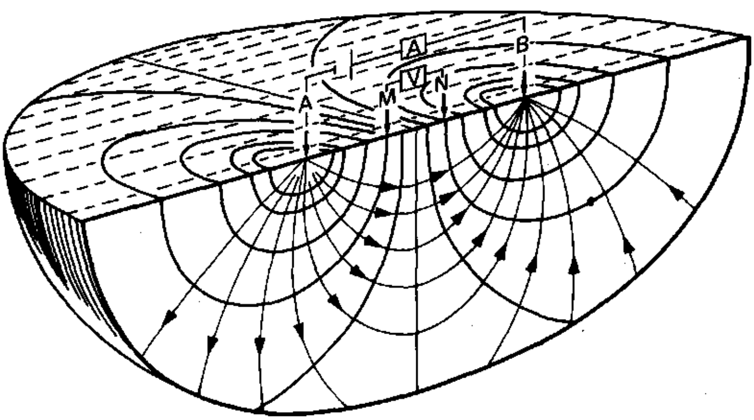


Fig. 13.
The electrode configuration according the Schlumberger array. The two current electrodes A & B are moved symmetrically outward with respect to the fixed potential electrodes M & N. With increased distance between A & B (=d(AB)) information from deeper levels are obtained.

For these soundings a Schlumberger array was used (Fig. 13). A direct or alternating current is introduced into the ground by means of two electrodes (A & B). The potential difference at the sounding point is measured between two probes (M & N) which are connected to a reading device. The essential idea behind VES is that with an increasing distance between the current (A&B) and potential electrodes (M&N), the current filament passing across the potential electrodes returns to the surface from increasingly deeper levels. With the Schlumberger array this situation is reached when the M/N-electrodes are kept fixed and the current electrodes A and B are moved symmetrically outward. In our case the maximum distance between the current electrodes was 600 m.

The unit of the electric resistivity of rocks is Ωm and is an extremely variable property as it not only depends on the type of rock but also on moisture content, ionic conductivity of the water, porosity of the sediments *etc*. The values outlined in Table 2 are applied for the sedimentary infill of the Pitalito Basin with the assumption of a constant groundwater quality. The values of peat and clay are confirmed by resistivity logging down to c. 8 m in the NE-part of the Pitalito Basin. The other values are based on the type of sediments at the surface (Chapter 4) combined with the results of the VES at that particular place.

Table 2.
The used electric resistivities (Ωm) of the sediments present in the Pitalito Basin.

Lithology	Electric resistivity (Ωm)
peat/clay	10-25 Ωm
sand (fine to medium)	25-80 Ωm
sand (medium to coarse)	50-120 Ωm
gravel & pebbles	>200 Ωm
basement	60-100(?) Ωm

The resistivity measured with a specific d(A,B) is called the apparent resistivity (ρ_a) and is plotted against AB/2 on a double logarithmic plot (Fig. 14). Subsequent curve matching with so-called master curves will lead to a model whereby one or more stratigraphical layers are involved. In this way a specific resistivity (ρ_s) can be calculated for a certain number of layers. For example, the values of the observed apparent resistivities in VES 1 (Fig. 14A) result in a four-layer case with at the top a layer about 10-15 meters thick with a very high specfic resistivity of approximately 350 Ωm. For a more detailed description of the electrical method one is referred to Telford *et al.* (1976) and Parasnis (1986).

3.2.2 INTERPRETATION

The soundings have been interpreted using a computer program based on the linear filter method (O'Neill 1975; Koefoed 1979). The basin appeared far to deep to be delineated geometrically using the available geoelectrical equipment and only in a few soundings close to the border of the basin, could the basement be reached. This is in agreement with the results from the gravimetrical survey.

In Figure 14 three VES-diagrams are shown which are representative for three regions in the Pitalito Basin. VES 1 is typical for the westernmost part of the basin and shows high to very high specific resistivities which increase towards the surface. There is a considerable decrease of the resistivities towards the east (*e.g.* VES 20) but the trend of increasing values towards the surface can still be recognized. This trend is characteristic for all the soundings west of the Guarapas river. East of the Guarapas river the calculated values show a strong decline (*e.g.* VES 50) compared with the aforementioned soundings. The resistivities do not show any significant change and remain low to very low. The measured resistivities are depicted in three contour maps of AB of 600 m, 300 m and 100 m, respectively (Fig. 15). Subsequently, the calculated specific resistivities of the sediments were interpreted and outlined in a fence diagram (Fig. 16). Due to the limited length of the VES the derived general character and distribution of the depositional facies only accounts for the upper hundreds of meters of sediment (100-200 m). As the eastern part is filled with c. 1200 m of material it is obvious that only the top part of the basin infill can be described.

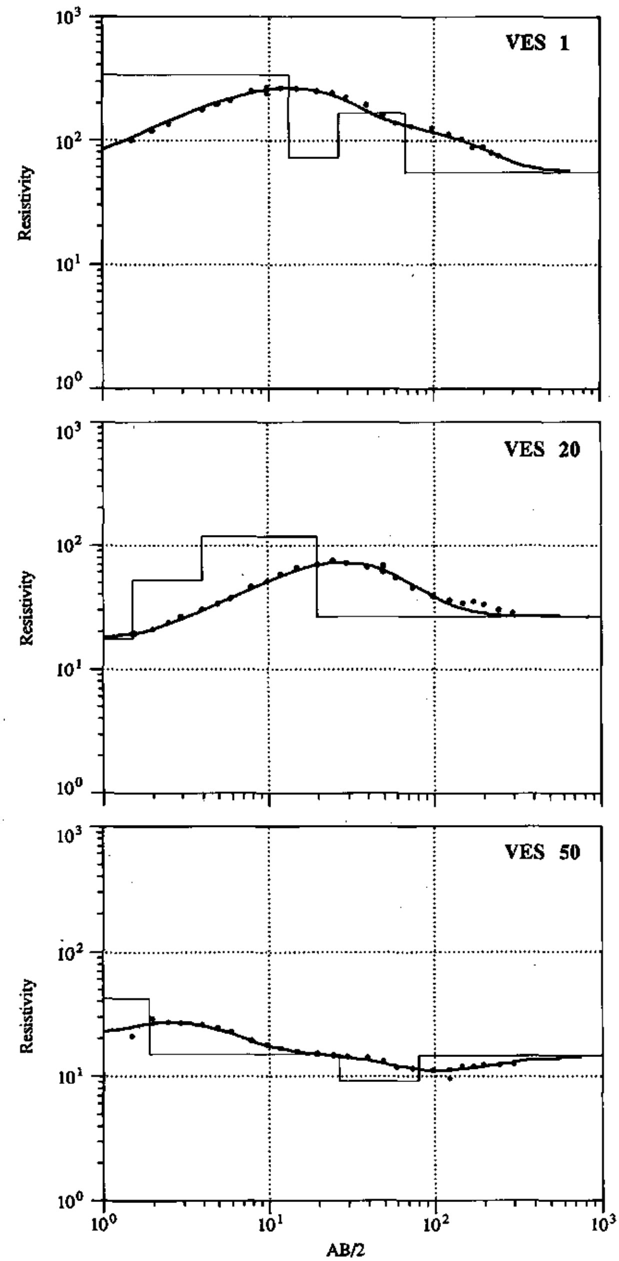


Fig. 14.

Three plots on logarithmic scale showing the ρ_a curve (thick solid line) and the obtained layered model (thin solid line). The latter reveals the number of involved layers, their thickness and their specific resistivity. Appendix III shows the location of these VES.

The contour map with an electrode distance of 600 m (Fig. 15A) shows relatively high resistivities in the western part of the basin which grade towards the east to low and very low resistivities. Comparing the distribution of these resistivities with the basement morphology it becomes apparent that low resistivities are restricted to the deep eastern part of the basin and the higher resistivities to the shallow western part. The transition between both types of resistivities lies at the latitude of the normal fault (no 5; Fig. 12).

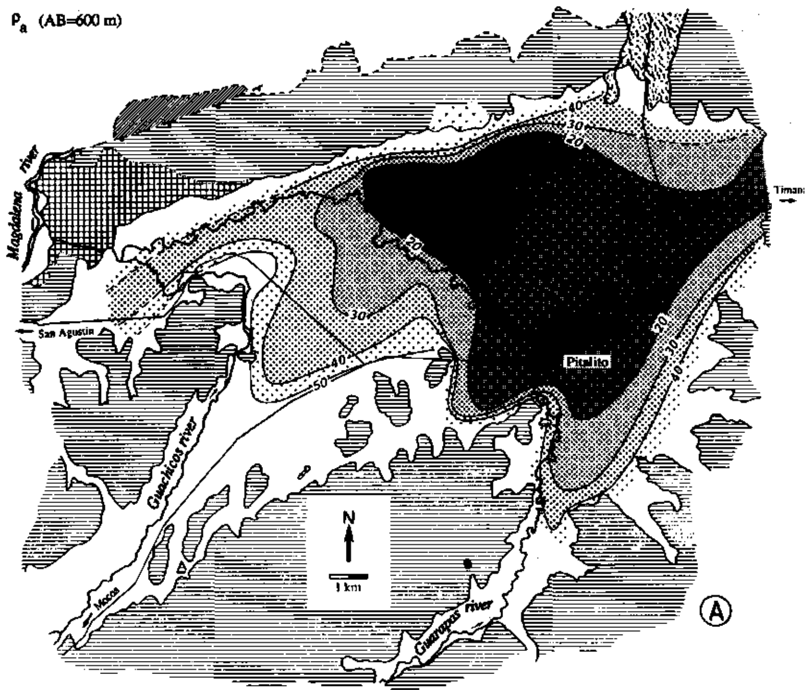
With an electrode distance of 300 m (Fig. 15B) the higher resistivities in the west have shifted somewhat to the east compared to those measured with an electrode distance of 600 m but on the whole are still restricted to the western side of the Guarapas river. Noteworthy are the high resistivities in the northeastern part of the basin.

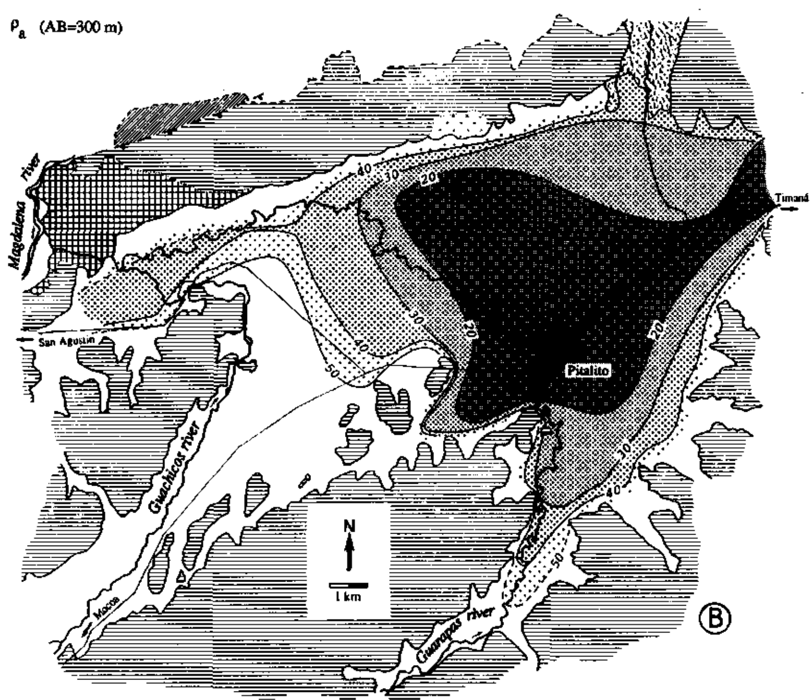
At an electrode distance of 100 m the resistivities in the west become high to very high (up to 100 Ωm or more) and again the transition to sediments with lower resistivities has shifted to the east (Fig. 15C). The areal extent of low resistivities in the eastern part shows a significant decrease.

According to the above-mentioned method the sediments which fill in the basin were classified into three groups (Fig. 16). The following conclusions can be made concerning their distribution:

- The coarse sediments principally remain restricted to the western shallow part of the basin (fences A-A' and B-B'; Fig. 16) whereas the finer sediments (clay and peat) are found in the east (fences C-C' and D-D').
- The lateral transition between both types of sediments is abrupt. Its position is stable in time and lies at the latitude of the NW/SE-oriented normal fault (fence B-B'; Fig. 16).
- The actual near-surface and surface sediments show a transgressive overlap of coarse material over fine-grained sediments from west to east (fences A-A' and B-B'). These sediments embody the last stage of sedimentary infill of the basin.
- The extension of coarse material in the northeastern area (fence D-D'; Fig. 16) is ascribed to the progradation of an alluvial fan into the basin (D; Photo 2). The material originates from the northern ignimbrites which are easily eroded and dissected by the Regueros rivulet.

On the whole, the maps of the resistivities of the basin sediments show a similar contour pattern as the Bouguer anomaly contour map (Fig. 10). The distinction between western and eastern basement morphology is not only expressed by the grain-size distribution but also confirmed by differences in alluvial architecture and surface morphology of the last-stage basin infill (see Chapter 4).


ρ_a (AB=600 m)
Magdalena river
San Agustin
Timaná
Pitalito
Guachicos river
Mocoa
Guarapas river
20
30
40
50
N
1 km
A


ρ_a (AB=300 m)
Magdalena river
San Agustin
Timaná
Pitalito
Guachicos river
Mocoa
Guarapas river
20
30
40
50
N
1 km
B

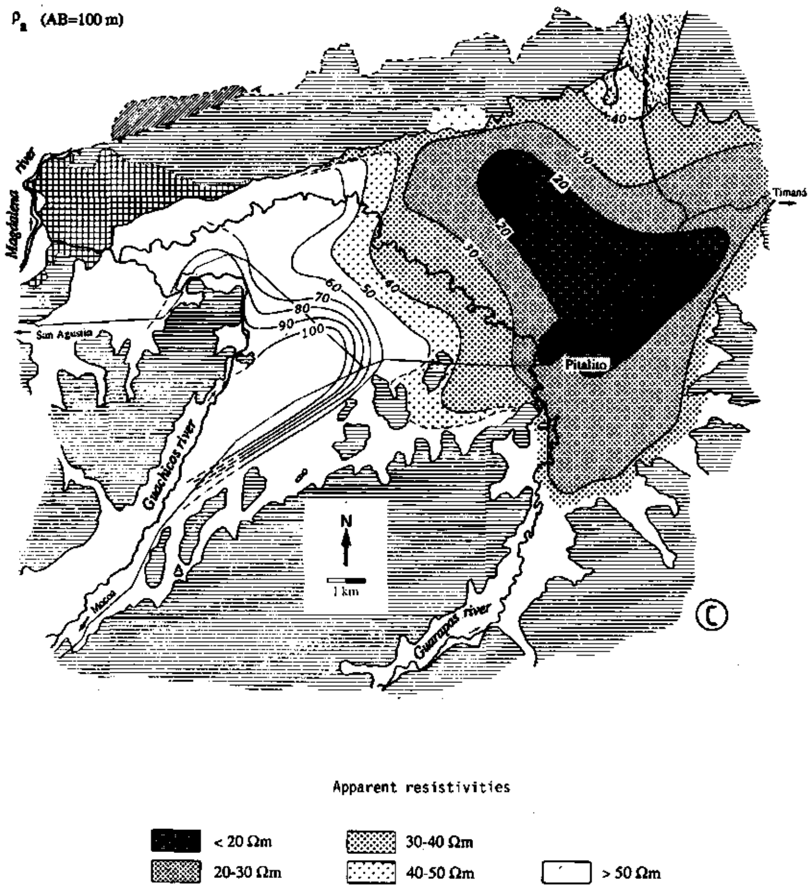
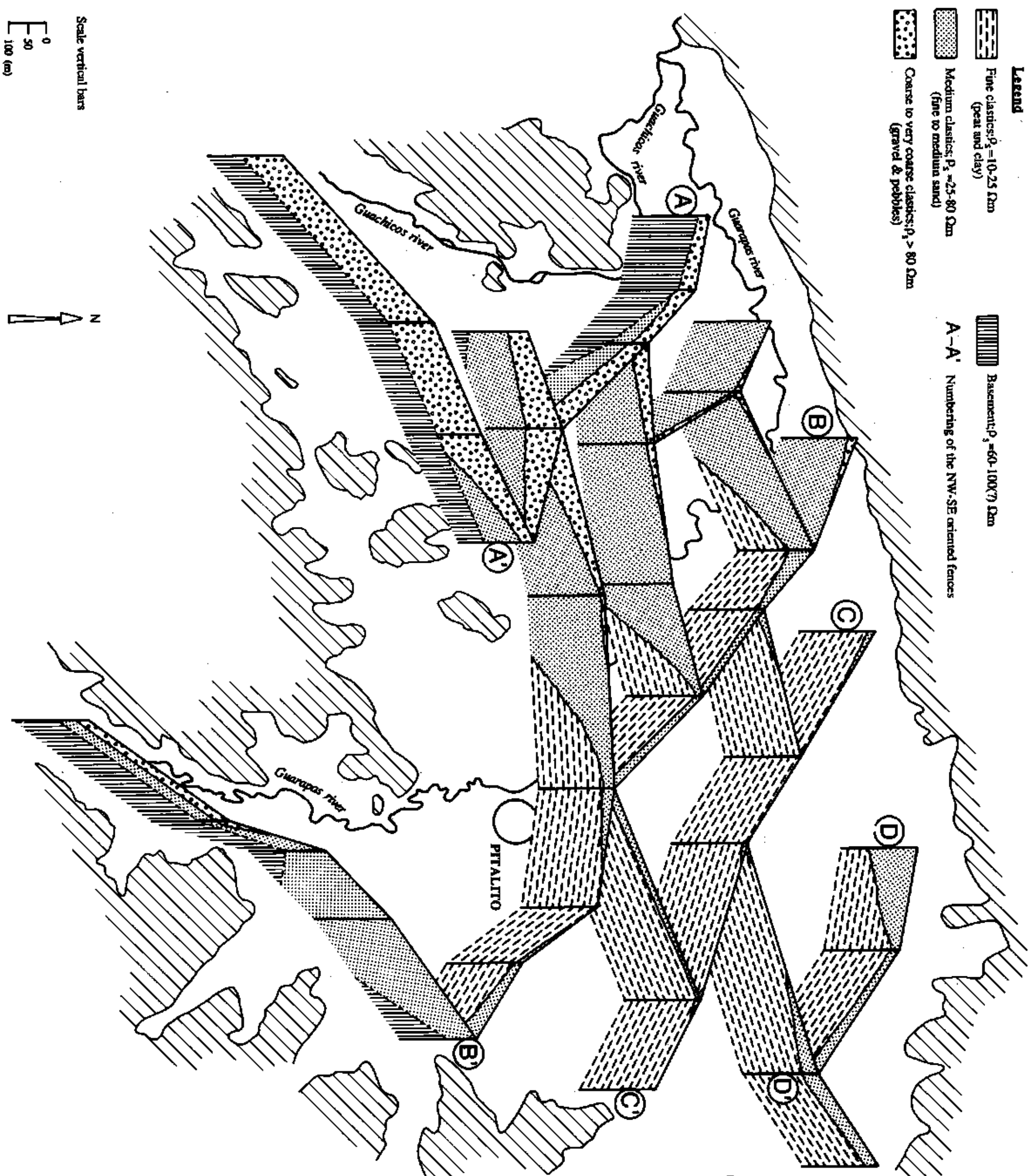


Fig. 15.
Contour maps of the apparent resistivities of the sedimentary infill of the Pitalito Basin at d(AB) of 600 m (A), 300 m (B) and 100 m (C), respectively. See Figure 10 for the legend of the surrounding area.

Fig. 16. (next page)
Schematic fence diagram showing three classes of infill of the Pitalito Basin. The classification is based on the calculated specific resistivities of these sediments. The depth of the basin floor is principally based on the gravimetrical data.

Non-exposed basin features



3.3 Conclusions

TECTONIC AND DEPOSITIONAL HISTORY

The initiation of the Pitalito Basin can possibly be explained by the horizontal displacement along the northern dextral strike-slip zone (no. 1; Fig. 12) leading to zones of transpression and transtension (Reading 1980). Such transtensional zones are characterized by subsiding basins which show evidence of large and rapid vertical movement along oblique-slip faults and become the loci of sedimentation. The general idea is that subsidence tends to occur where strike-slip is accompanied by a component of divergence, for example through extension near a fault junction which leads to a 'fault-wedge basin' (Crowell 1974). This is idea compatible with the dextral character of the strike-slip fault and the fault junction in the northeastern part of the Pitalito Basin. In that part of the basin the complex, braided fault pattern of the Garzón-Suaza fault (Fig. 2B) diverges in a northern and a southern fault (Figs. 12 & 17; no. 1 and 2, respectively).

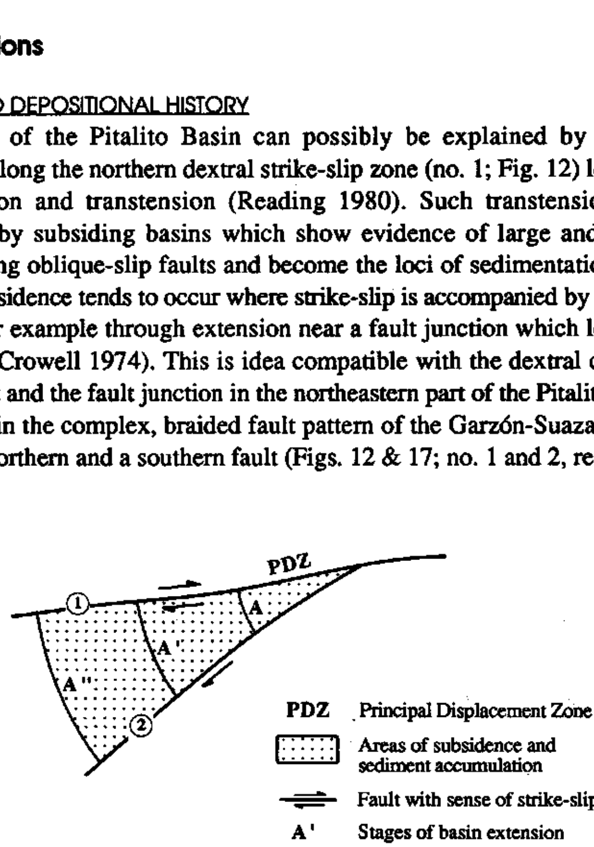


Fig. 17.
Sketch map showing subsidence of the tip of a fault wedge with divergence (Crowell 1974). The numbers 1 and 2 refer to similar faults in the Pitalito Basin (Fig. 12).

From Fig. 17 it is clear that subsidence is related to the orientation and the overall slip vectors of the geological structures. In the example the graben initially developed in the eastern part of the basin: A; and extended towards the west in time: A', A''. The three fault zones in the Pitalito Basin: zones 2, 3 & 4, respectively (Fig. 12), are supposed to branch from the northern dextral strike-slip fault as a result of lateral displacements along the latter. Such branching splays are commonly found along strike-slip faults and may form flower structures (Christie-Blick & Biddle 1985). The characteristics of these upward-diverging faults vary between normal and reverse faults. Based on the morphology of the southern basin margin and the differences in vertical offset, fault zone 2 (Fig. 12) is considered to form the 'first generation' of NE/SW-oriented faults.

In the scope of the 'fault-wedge basin' model which is schematically shown in Figure 17, three phases of tectonic and depositional history are distinguished for the Pitalito Basin. As stated before only the upper part of the c. 1200 m thick sediment pile could be described by means of the geoelectrical survey. Therefore, these sediments only describe basin fill during the later stages of the evolution of the Pitalito Basin (Fig. 18; Phase 2 and 3, respectively). A fourth phase represents the actual situation which is characterized by erosional processes. Accepting a sedimentation rate of 0.25-0.30 cm/ 1000 years (section 5.2.2.3) for the entire infill of the basin (c. 1200 m thick) implies that sedimentation in the Pitalito Basin may have started around 4.5 Ma ago.

-Phase 1 (Fig. 18A) At first, subsidence was mainly restricted to the present eastern part of the basin. The principal structures were formed by the northern strike-slip fault (no. 1; Fig. 12), the southern basin-margin fault zone (no. 2) and the NW/SE-oriented normal fault (no. 5). This normal fault possibly formed a terminal fault and developed during the earliest stage of basin subsidence. Subsidence caused the accumulation of a thick sediment pile in the area.

The applied geoelectrical method could not reach the sediments which were deposited during this initial phase of subsidence. As data about the areal facies distribution in the Pitalito Basin are lacking, a paleoenvironment is assumed which is quite similar to the one that prevailed during the succeeding phase 2: a poor sediment supply from the west into a deep or shallow lake with possible presence of organic-rich sediments, due to local emergence. According to Pitman III & Andrews (1985) a 'pull-apart' basin may be sediment starved during the initial phase of subsidence. This situation may last several millions of years and is explained by the rapid subsidence caused by rifting and cooling during this period. In time, subsidence rates decrease as only cooling is responsible for subsidence and the basin will be filled in with sediments. When the 300 m depth line is accepted as upper limit for the non-detected sediments, phase 1 would describe the period from ≈4.5 Ma to ≈1.2 Ma ago.

-Phase 2 (Fig. 18B) With continued dextral strike-slip displacement, transtension resulted in a westward growth of the basin and the development of a 'second generation' of splay faults along the southern margin further to the west (fault zone no. 3; Fig. 12). From then on, the actual dimensions (*viz.* length and width) of the present Pitalito Basin were almost established and subsidence continued both in the west and the east. The areal facies distribution of this phase is shown by the deep electrical soundings (Figs. 15A and 15B, respectively).

In the western part of the basin coarse to medium-grained clastics were deposited in a fluvio-lacustrine environment. Fine-grained, possibly organic-rich sediments were deposited in the deep eastern part of the basin. These sediments were probably deposited in a lacustrine environment (Fig. 18B). The sediment influx was from west to east and deposition occurred in a shallow-water delta with streams that migrated freely across the broad, shallow western basin floor which slopes towards the east. Due to high rates of subsidence the rivers continually fell behind to build up an equilibrium profile (McClean and Jerzykiewicz 1978). This initially resulted in a westward regression of the fluvio-lacustrine sediments compared to phase 1 and promoted sediment starvation in the eastern part of the basin.

The transition from the western fluvio-lacustrine to the eastern lacustrine sediments is situated along the line of the NW/SE-oriented fault (no. 5; Fig. 12). It is indicative for the tectonic control on the grain-size distribution. A plausible explanation for the coincidence of geological structures and the type of basin infill might be found in varying subsidence

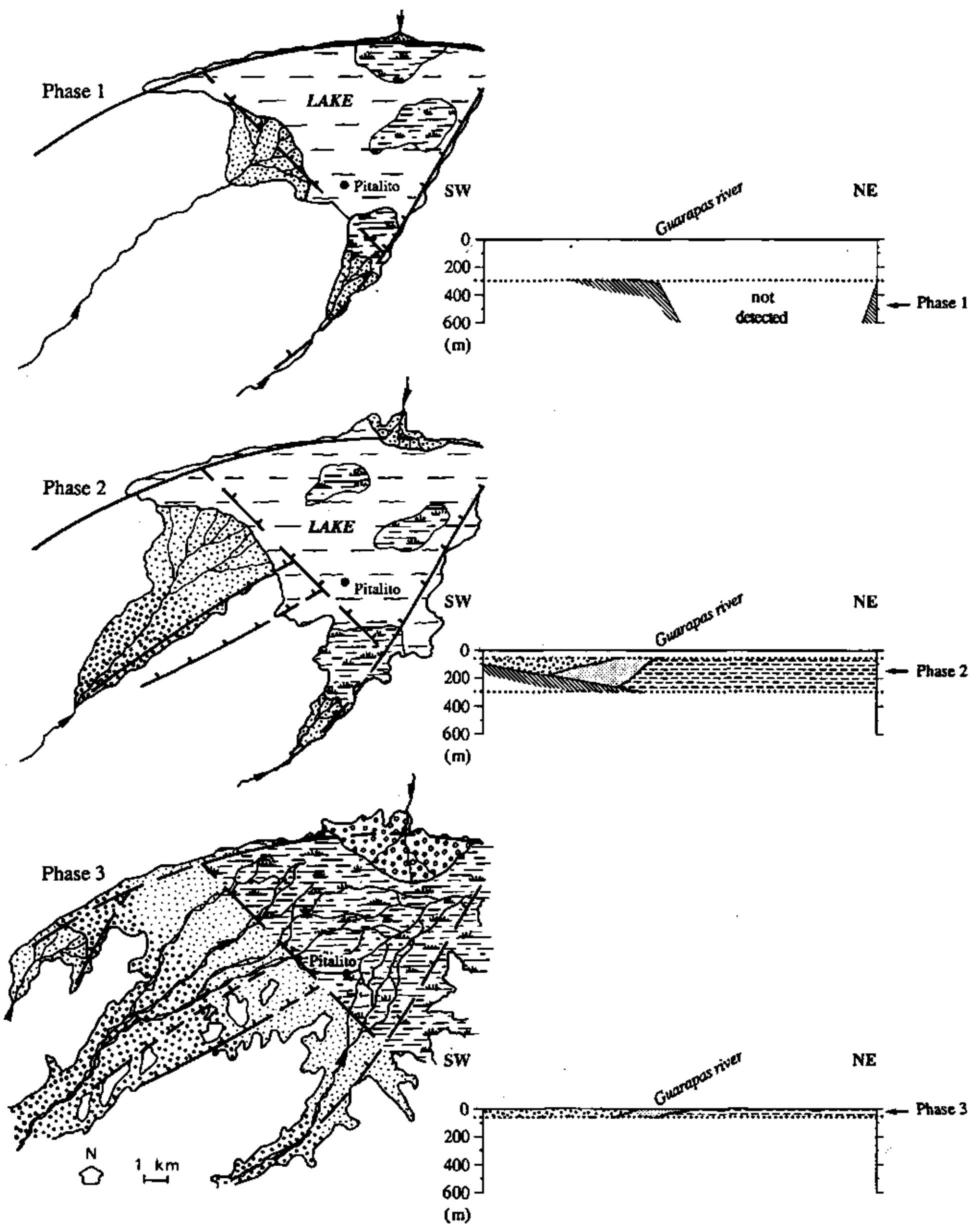


Fig. 18.
Tectonic-sedimentary evolution of the Pitalito Basin explained within the framework of a transtensional regime along a segment of the dextral strike-slip Garzón-Suaza fault system. In the left-hand figures the paleoenvironment is depicted. The right-hand figures represent profiles along the SW/NE-basin axis and correspond to the outlined phases. The arrow points to the stratigraphical position of the sediments which were deposited during the respective phases. For the legend of the sediments see Fig. 16.

rates with higher rates in the eastern part compared to the western part. Loading effects may have played a role in this differentation; the presence of the thick sediment pile in the east might have lead to higher subsidence rates in this part of the basin comapred to the western part where such a thick pile was lacking. Phase 2 lasted from ≈1.2 Ma to ≈200 Ka if the described sediments for this phase are assumed to be situated between approximately 300 and 50 m depth.

-Phase 3 (Fig. 18C) The branching splay from the northern oblique-slip fault in the outermost NW part of the study area (no. 4; Fig. 12) suggests the existence of a younger phase of westward basin extension. This normal fault may be considered as the 'third generation' of splays. Phase 3 describes the period from ≈200,000 years B.P. to 17,000 years B.P. During this phase material was deposited which constitutes the actual surface and near-surface sediments of the basin. They represent the final stage of basin fill and are of fluvial origin. Their sedimentological characteristics will be discussed in detail in Chapter 4. From the geoelectrical data the following conclusion can be drawn:

West from the actual position of the Guarapas river, coarse-grained sediments were deposited on top of the underlying fluvio-lacustrine sediments from phase 2 (coarsening-upward megacycle). This is explained by an eastward prograding fluvial system. The progradation can be ascribed to decreasing subsidence rates in the entire basin or to the increase of sediment supply, leading to a state where basin infilling outpaced subsidence. According to McClean & Jerzykiewicz (1978) such a state would produce maximum alluviation distally, and possibly erosion and reworking proximally. Similarly, the Ridge Basin of California is filled by fluvial sediments in its last stage (Crowell & Link 1982). To the east the grain size of the material deposited by this river system decreased. Except for the occasional deposition of sand in this area, detrital influx was low and enabled the accumulation of very fine material like clay and peat. The transition between the coarse sediments in the west and the fine-grained sediments in the east is located along the line of the NW/SE-oriented normal fault (no. 5; Fig. 12). This means that tectonic control still played an important role in the distribution of the sediments.

-Phase 4 (Appendix I) This phase is reflected by the modern surface morphology and the present processes that take place in the Pitalito Basin. The modern drainage system shows a NW flow direction and the river beds are incised deeply into the accumulated sediments. The 90°-turn of flow direction (section 2.2.1) and the reversion of deposition to erosion, mark the transition from phase 3 to phase 4. Such a regional change might be explained by some allocyclic control, *e.g.* tectonics or a climatic change. According the ^{14}C datings of PIT 11 and PIT 2, phase 4 started somewhere between 17,000 and 7,000 years B.P.

3.4 Discussion

The development and morphology of the Pitalito Basin is analogous to that of the Little Sulphur Creek Basins (McLauglin & Nilsen 1980, 1982; Nilsen *et al.* 1980; Nilsen & McLaughlin 1985). These similarities are outlined in Fig. 19. The Little Sulphur Creek Basins developed as a result of oblique pull-apart extension along the right-lateral Maacama fault zone. Although the lateral offset along the northern basin margin is unknown, the Pitalito Basin shows many characteristics which are applicable to basins which develop adjacent to strike-slip faults (Nilsen & McLauglin 1985):

- The Pitalito Basin is formed along the Garzón-Suaza fault zone: a zone showing dextral strike-slip movement.
- Of all the fault-controlled basin margins, the strike-slip fault along the northern one is the most prominent.
- The basin is asymmetrical, with its structurally deepest part close to the active strike-slip margin.
- The basin fill is dominated by axial infilling (from west to east), subparallel to the northern strike-slip basin margin.
- The basin fill is characterized by abrupt lateral facies changes.
- The abrupt lateral facies changes are stable in time.
- The small-sized basin contains a very thick sedimentary sequence.
- The basin-margin deposits are distinctive: along the active strike-slip margin, debris-flow-dominated fans occur of limited extent whereas along the inactive southern margin larger streamflow-dominated fans are present.
- Basin extension (towards the west) is opposite to that of the general flow of paleocurrent (towards the east). This feature is compatible with findings from the Dead Sea Rift and Ridge Basin (Manspeizer 1985; Nilsen & McLaughlin 1985, respectively).

An important assumption in the tectonic model is the higher subsidence rates in the east compared to the western part of the basin. This is not only supported by the distribution of the basin infill but also by the c. 10 m vertical step in the actual topography along the line of the Guarapas river (Fig. 4A). This step partly explains the position of the Guarapas river: If this river was not influenced by the actual surface topography one would expect a far more direct course towards the NW outlet. The question if and to what extent the eastern part really shows higher subsidence rates can be solved by dating the sediments at several depths in the eastern as well as in the western part. As the available datings are restricted to the upper c. 20 m of sediments they yield no information for this problem.

Besides the effects on subsidence caused by movements along the fault patterns, the proposed tectonic model for the Pitalito Basin includes loading as effect on subsidence as well. McClean & Jerzykiewicz (1978) pointed out that loading may have a substantial influence on local subsidence as well as on subsidence of the surrounding areas ('distal loading effect'). If distal loading played a role in the Pitalito Basin, the enormous sediment accumulation and subsequent local loading effects in the east may have induced the initial downwarp of the west. This leads to sediment accumulation beyond the orignal basin margins and subsequent progressive subsidence towards the west. In other words, distal loading may have had its influence on the enlargement of the Pitalito Basin in the opposite direction of the general paleocurrent. Similar basin extension and depocentre migration is found in several other strike-slip basins (Reading 1980; Nilsen & McLaughlin 1985). Although in all these cases tectonics is undoubtfully the major controlling factor in the migration of the depocentre and its direction, loading may have

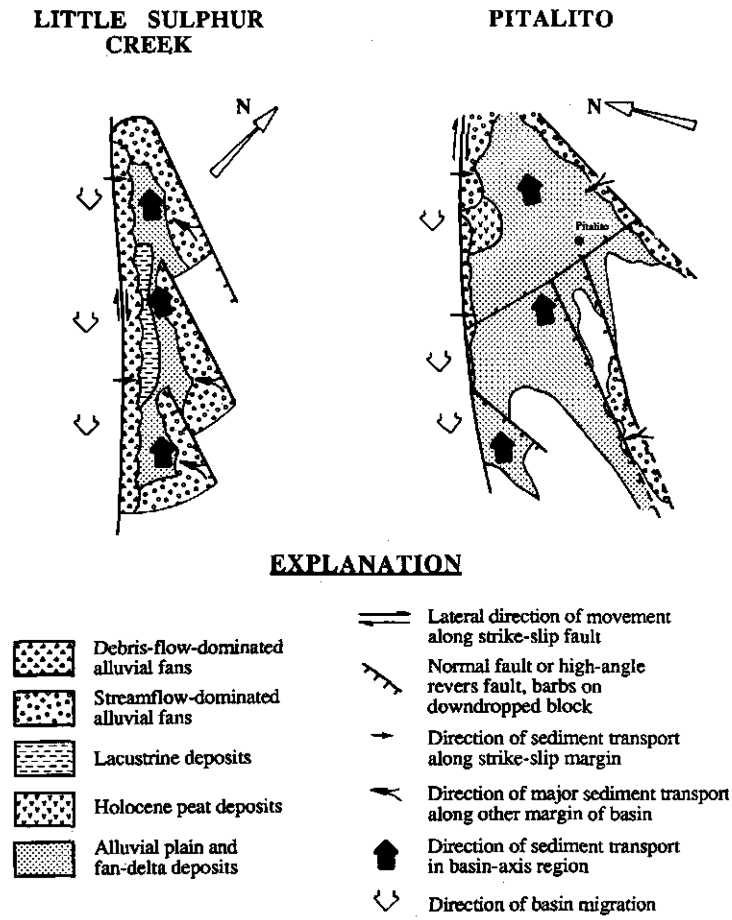


Fig. 19.
Tectonic and sedimentary comparison of the Little Sulphur Creek basins in California (redrawn from Nilsen & McLaughlin 1985) and the Pitalito Basin. The faults are drawn as if not concealed. For easy comparison both basins are drawn roughly at the same size and are reoriented.

played a role as well.

In the proposed tectonic model two phases of westward extension can be distinguished which are delineated by the fault zones 3 and 4, respectively. Fault zone 4 is the youngest zone and tectonic activity along this 'third generation' of branching faults probably lead to the 90°-turn of the drainage pattern and a lowering of the local base level which mark the transition from phase 3 to phase 4.

Reading (1980) distinguished three phases in the formation of a strike-slip basin: (1) a phase of transtension, (2) a phase of basin infilling and (3) a phase of transpression. The transition from phase 2 to phase 3 shows a change in the basin fill characterized by a basinwide coarsening-upward sequence due to a decreasing subsidence rate. The last stage of infilling of the Pitalito Basin (phase 3) might reflect this transitional period.

The average sedimentation rate for the surface sediments, calculated on the basis of ^{14}C datings in peats, is 0.25-0.30 m/1000 years over the last 60,000 years (section 5.2.2.3). Generally, average sedimentation rates in strike-slip basins are higher than the calculated ones for the Pitalito Basin (Schwab 1976; Miall 1978). When this calculated figure is accepted for the entire infill of 1200 m, this would imply that subsidence of the Pitalito Basin started around 4.5 Ma ago. However, a faster/slower deposition rate for the bulk of the sediments (that are beyond ^{14}C reach), cannot be excluded. This rate would increase for instance when phases of non-deposition are taken into account and not included in calculations of average values. Such phases, which are described by Pitman III & Andrews (1985), take place especially when stretching begins. Furthermore, compaction of the organic-rich sediments on which the calculations for the Pitalito Basin are based can also result in lower accumulation rates.

A final remark concerns the negligible role of sediment input of the Guarapas river in former times; throughout the entire registered period of basin infill there is no evidence of an important sedimentary influx coming from the narrow Guarapas Valley. One would expect a relatively coarse-grained, fan-shaped body directly SE of Pitalito town where the Guarapas river leaves its narrow valley. Such a body has not been detected by the geoelectrical survey. It seems that the water flow velocity of the Guarapas river is not able to transport coarse material beyond its narrow valley into the Pitalito Basin. The present situation is quite similar: coarse clastics in the actual bed of the Guarapas river are restricted to the area of the shallow valley and along the northern basin margin (Appendix I). When crossing the basin plain only fine clastics are found.

CHAPTER 4

SEDIMENTARY CHARACTERISTICS OF THE LAST-STAGE BASIN INFILL

The actual near-surface and surface sediments constitute the final stage of sedimentary infill of the basin. Geoelectrical data indicate that these sediments are characterized by a transgressive overlap of coarse material from west to east (Chapter 3). This coarsening-upward megasequence (tens of meters thick) is interpreted as an eastward prograding ancient fluvial system which also has its expression in the morphology of the plain (Appendix I). The sedimentary sequence of the uppermost 40 meters of these sediments have been studied in numerous clay pits, borings and outcrops along the riverbanks of the incised Guarapas and Guachicos river and will be discussed in this chapter. The exposed sediments are referred to as sections followed by a number (*e.g.* Section 20A) whereas the borings are abbreviated as PIT followed by a number (*e.g.* PIT 11). The position of the described lithological sequences are shown in Appendix IV.

4.1 Areal facies distribution

As was stated in Chapter 3 the Pitalito Basin can be subdivided into a shallow western part and a deep eastern part. The transition between both parts lies along the line of the Guarapas river. It appeared that this twofold division could also be used for the surface morphology and drainage pattern (Appendix I): west of the Guarapas river a pockmarked relief with small seasonally ponded depressions and east of the Guarapas river interconnected ridges that enclose badly drained depressions (pages 22 and 23).

Additionally, based on their lithological content and characteristics, the same division in a western and eastern part could be used for the surface and near-surface sediments.

4.1.1 WESTERN PART

The surface sediments in the western part, *viz.* west of the Guarapas river, are roughly made up by two types of sediments (Figs. 20 to 23): 1/ coarse-grained sediments (pebbles, gravel and sand) and 2/ fine clastics which encase the first-mentioned type of sediments and form the bulk of the total lithofacies in the western part of the basin with about 60%.

Coarse-grained sediments

Description—The coarse sediments range from fine sand to clast supported cobbles (max. diameter of c. 10 cm). Although small isolated lens-shaped bodies are present, they commonly have a sheet-like appearance with a maximum thickness of about 3 m and

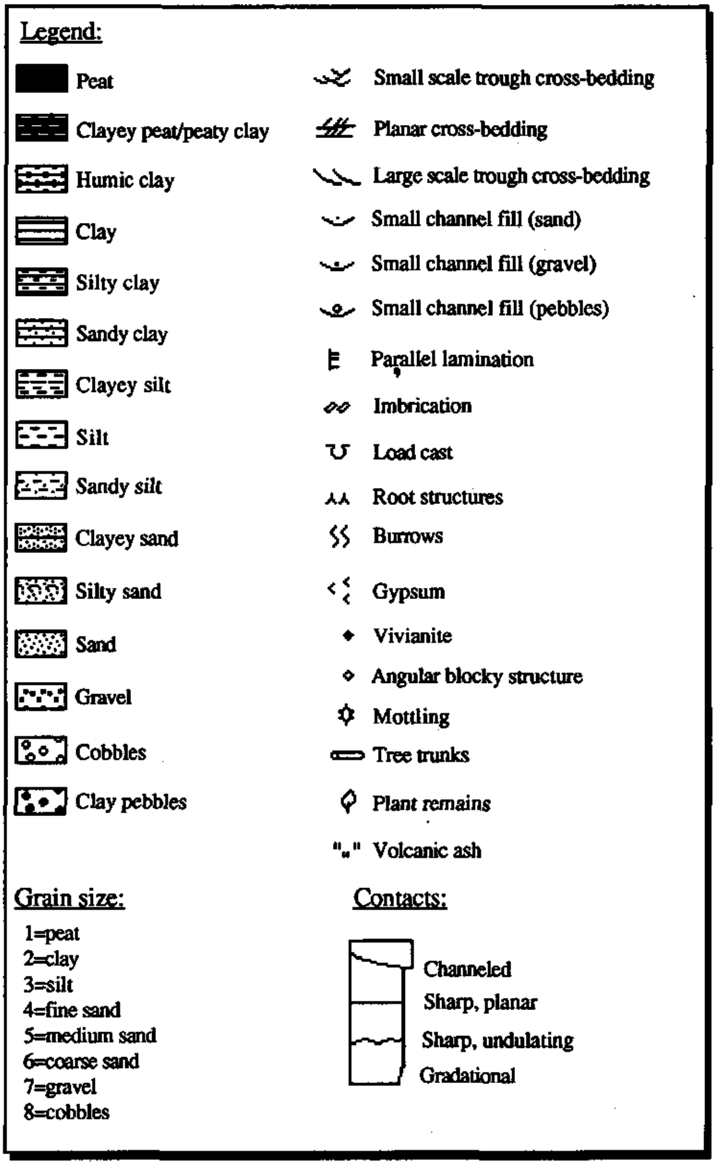
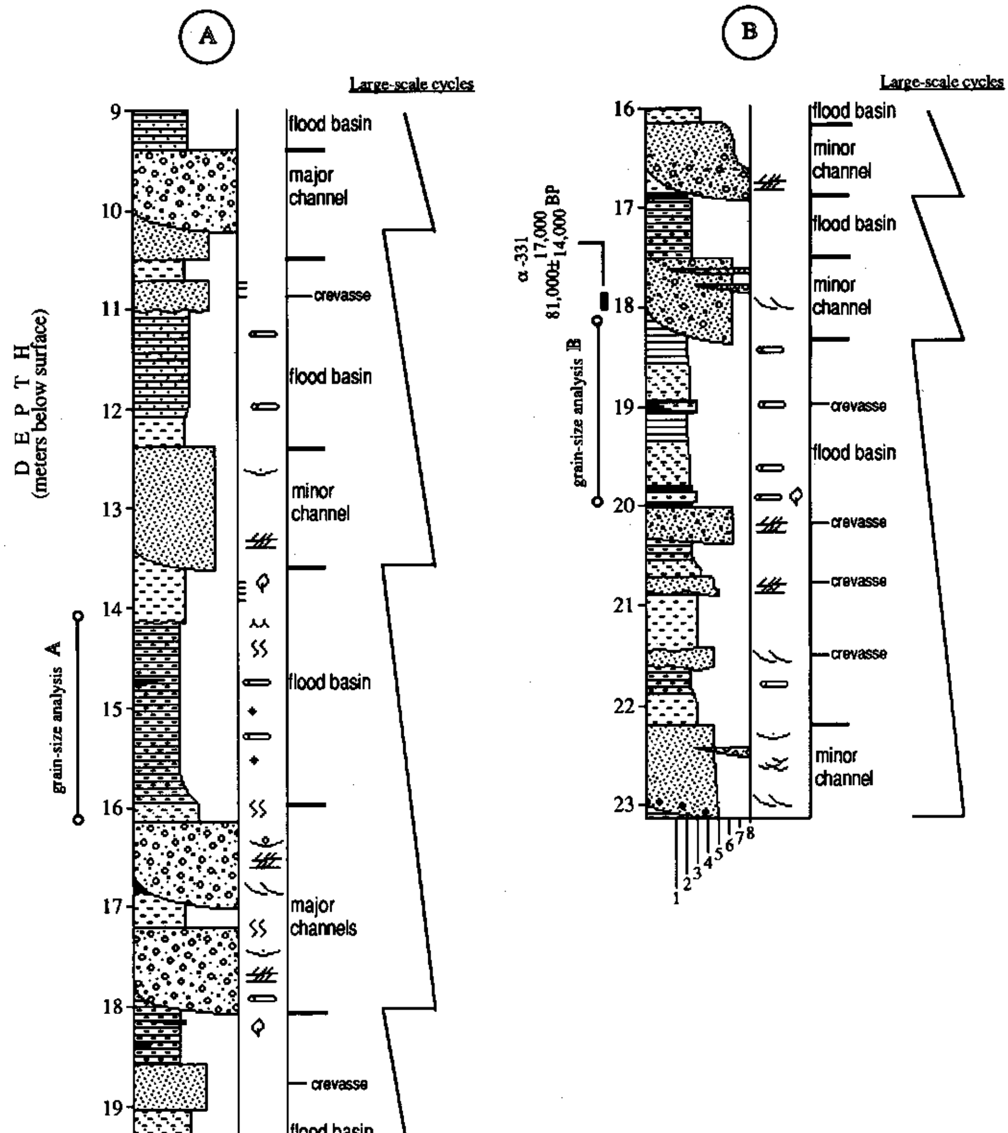
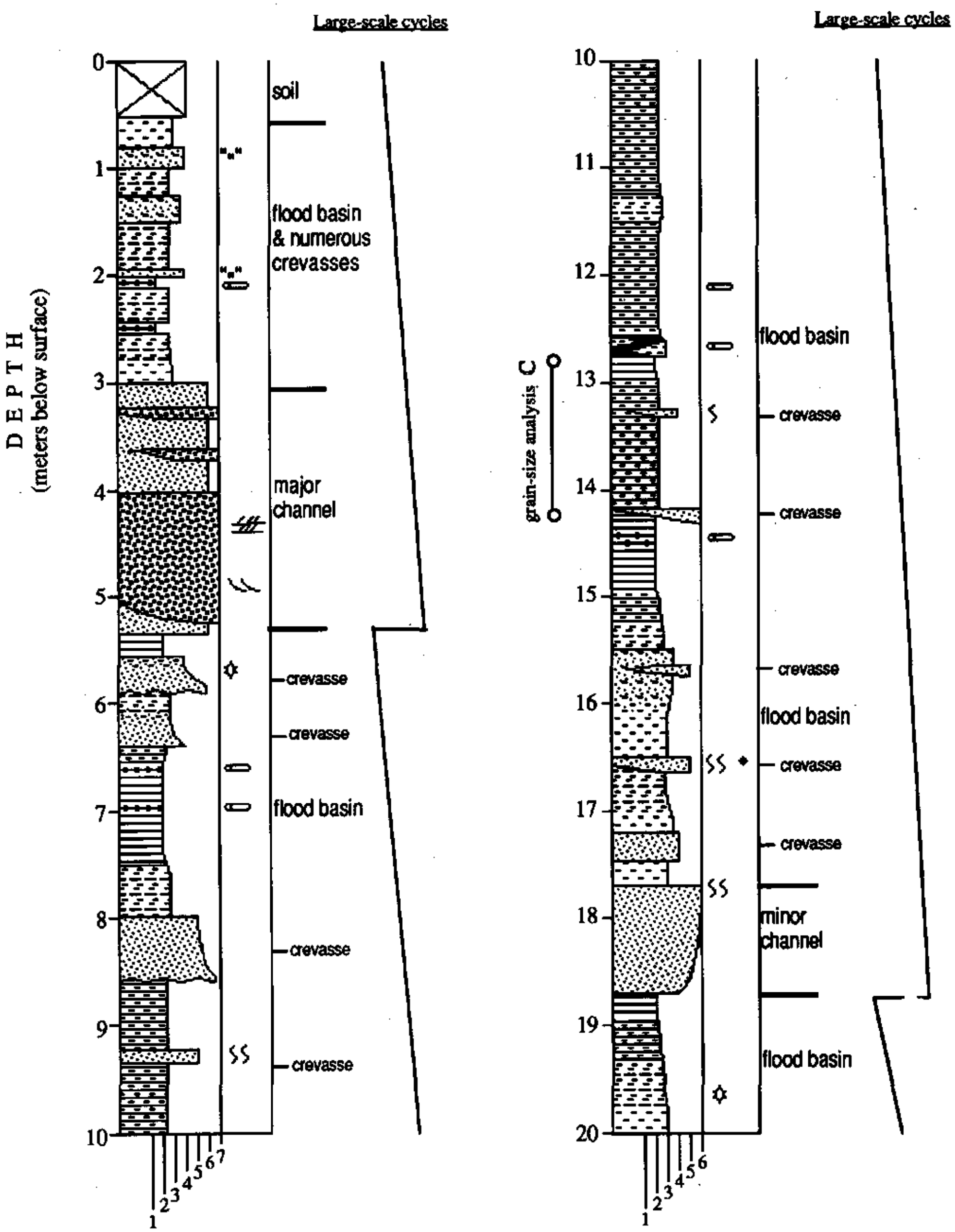


Fig. 20.
Columnar sections representative for the last-stage sedimentary infill in the western part of the Pitalito Basin. At the right side of each column the respective sedimentary environment and the fining-upward large cycles are shown (see also text). At the left side the stratigraphical position is indicated of the sediments which were analysed on their grain size (see Fig. 24). See Appendix IV for the location of the sections.

Section 20


A
B
Large-scale cycles
Large-scale cycles
D E P T H
(meters below surface)
grain-size analysis A
grain-size analysis B
α -331
81,000± 17,000
14,000 BP
9
10
11
12
13
14
15
16
17
18
19
20
21
16
17
18
19
20
21
22
23
1 2 3 4 5 6 7 8
1 2 3 4 5 6 7 8
flood basin
major channel
crevasse
flood basin
minor channel
flood basin
major channels
crevasse
flood basin
minor channel
flood basin
minor channel
flood basin
minor channel
crevasse
flood basin
crevasse
crevasse
crevasse
minor channel

Section 97


Large-scale cycles
DEPTH (meters below surface)
0
1
2
3
4
5
6
7
8
9
10
soil
flood basin & numerous crevasses
major channel
crevasse
crevasse
flood basin
crevasse
crevasse
1 2 3 4 5 6 7
Large-scale cycles
grain-size analysis C
10
11
12
13
14
15
16
17
18
19
20
flood basin
crevasse
crevasse
crevasse
flood basin
crevasse
crevasse
minor channel
flood basin
1 2 3 4 5 6

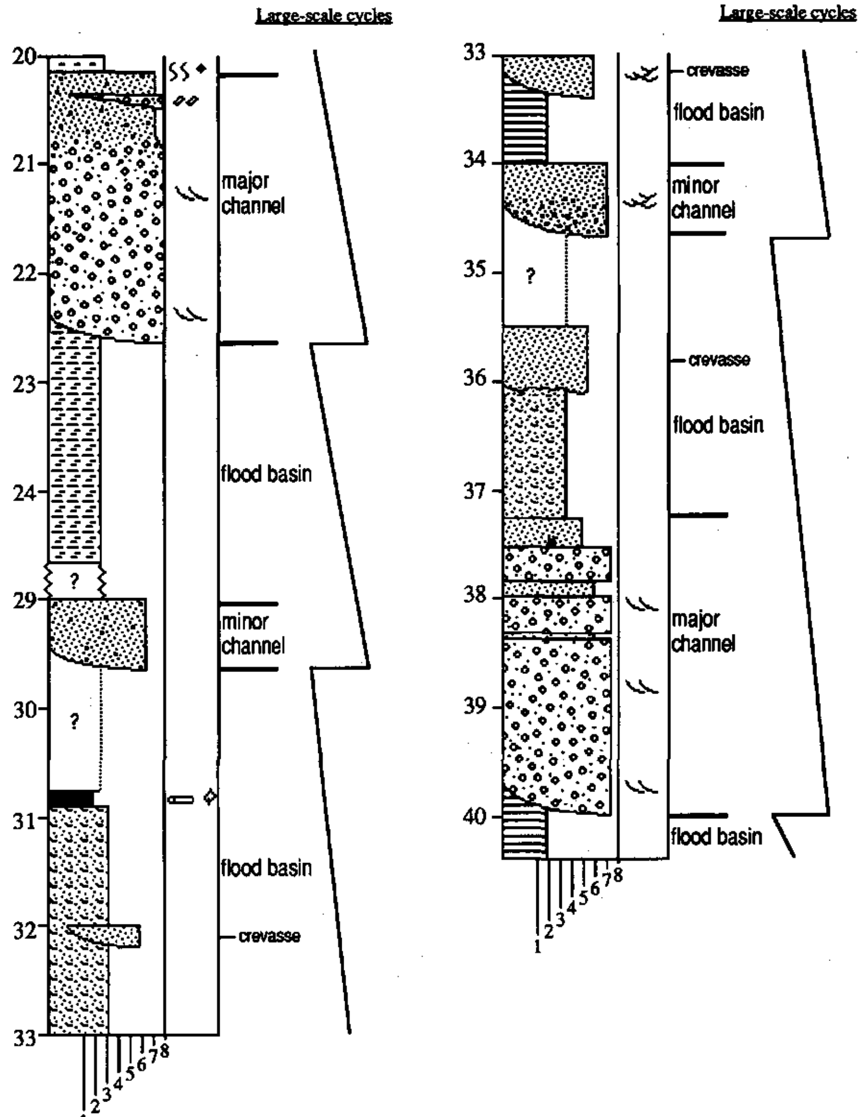


Fig. 21.
Columnar sections representative for the last-stage sedimentary infill in the western part of the Pitalito Basin. At the right side of each column the respective sedimentary environment and the fining-upward large cycles are shown (see also text). For the detailed grain-size analysis indicated at the left side at the column one is referred to Fig. 24. For the location of the sections one is referred to Appendix IV and for the legend to Fig. 20.

a width ranging between 20-30 m. Their bases are rather planar and do not display clear evidence of deep incision in the underlying finer deposits. Based on the lithology two types of coarse-grained bodies can be distinguished: (1) with pebbles as the predominant grain size and (2) with the sand fraction as prevailing grain size.

(1) Gravelly deposits The sediments show an internal structure of mutually cross-cutting channels and usually are massive to faintly bedded. Sometimes waterlogged wooden trunks are incorporated within the massive deposits. Large scale pebble-strewn foresets up to 0.5 m high are present. Locally, the channel bars are incised by small channels with a fine clastic infill. Clast imbrication is found. Occasionally, the top of the coarse-grained bodies shows features of animal burrowing. Examples of these deposits are shown in Section 20A between 16 m and 18 m depth (Fig. 20) and Section 97 at 21 m and 39 m depth (Fig. 21).

(2)Sand deposits These deposits range from very fine to coarse sand. They have either a sheet-like appearance (*e.g.* Section 97 at 8 m and 36 m depth; Fig. 21) or form small concave lenticular bodies between c. 50 cm and 2 m thick (*e.g.* Section 20B at 16.50 and 18 m depth; Section 97 at 29 m and 34 m depth). Small and large-scale trough and planar cross-stratification are the most common sedimentary structures. The cross-stratification may show variation within the foresets where concentrations of sand and silty sand are interbedded conform with the cross strata (Photo 11). In transverse view the strata appear as concentric concave upward traces, either parallel or tangent to the trough boundary. The height of the sets vary between 10-60 cm. Measurements on cross-bedded structures reveals a general flow direction from southwest to northeast. Penecontemporaneous deformation structures like load structures and ball-and-pillow structures are frequently present at the base. Commonly, the sediments contain animal burrowing features (Photo 11). The coarse sediments have a widely scattered character in a lateral and vertical sense. Its spatial distribution is expressed in two transects: one transect with a SW/NE orientation (Fig. 22) and one with a SE/NW orientation (Fig. 23).

Interpretation—The coarse-grained sediments are interpreted as channel deposits; the gravels have been deposited by a major channel (second order channel) and the sandy deposits by a channel of minor magnitude (first order channel). The absence of a 'vertical-stacked' pattern of the sandbodies (Fielding 1984b), their sheet-like appearance, the lack of evidence for incision of the fluvial sequences and the distribution of the channel facies embedded in fine clastics suggest relative short-lived shallow channels that wandered over wide flood basins present in the western part of the Pitalito Basin. The sharp bases of the channel facies indicate instantaneous introduction of coarse clastics in the clay flats. The sediments were usually deposited upon water-saturated clays of the flood basin which is evident from the presence of load casts. The internal geometry of the major channel deposits can be described as a sequence with complex depositional-erosional structures caused by lateral shifting channels in the active belt. Internal scouring and filling of the channels occurred in response to flow divergence across the convex upper surface of aggrading bars. Similar features are described by Turner (1986) for the more proximal reaches of a braided river system. The presence of heterogeneous cross-stratification is indicative for periodic changes in current velocities.

Based on the presence of coarse material, *viz.* sand and gravel, the western sedimentary infill can be divided into two units (Appendix I):

Unit A/ Gravel and sand with intercalations of inorganic clay (*e.g.* Section 136; Fig. 22).

Unit B/ Inorganic clay with intercalations of gravel and sand (*e.g.* Sections 75, 88 and 90; Fig. 23).

Fine-grained sediments

Description—The clay-rich sediments are usually of greater cumulative thickness than the coarse-grained sediments and form the major feature in the western area. They are made up by massive banks up to 4 m thick which show a vertically-stacked pattern (*e.g.* Section 20B between 22 m and 18 m depth; Fig. 20 & Photo 8). Such a pattern is only occasionally observed in the coarse-grained deposits. Except for some thin laminae containing plant detritus (*e.g.* Section 20B at 20, 19 and 17 m depth), the absence of organic-rich strata in the entire western region is conspicuous. Sometimes a thin (c. 10 cm thick), sharply bounded, humic-rich, dark-coloured layer is present (Section 20B, at 19 m and 20 m). These layers consist of horizontal-oriented leaves, stems, and branches which support each other and are traceable over tens of meters (Photo 8). The matrix consists of small silty lenses. Laterally these layers grade into finer clastics. Locally, flat-lying wooden branches can be observed floating in the matrix. Vivianite is present as small nodules and seems to concentrate in rooted and bioturbated horizons. Mudcracks are sparse. Stringers and sheets (max. c. 50 cm thick) of very fine to coarse sand are incorporated in the massive fine-grained deposits (*e.g.* Section 20B, between 18 and 22 m; Fig. 20). The most common sedimentary structures in these coarser sediments are planar and trough-stratification though they can be completely absent as well, especially in the thin sand sheets of c. 10 cm thick. Laterally, they wedge out in sandy clay deposits. The sandsheets show load-casting and are often bioturbated. Their base is sharp. Deposits of massive to finely parallel-laminated silt and silt-enriched material may appear as isolated layers in the flood basin sediments but usually show a prevalence in underlying the coarser facies (*e.g.* Section 20A at 18 m, 14 m and 10.60 m depth; Fig. 20). The silty layers do not exceed 50 cm in thickness and reveal an undulating sharp base. Occasionally, horizontal-oriented plant remains are present in the silty deposits. Locally, massive gravelly/silty sediments with a maximum thickness of ≈1 m are interbedded in the fine clastics (*e.g.* Section 75, at 4.25 m and 9.50 m; Fig. 22). Besides the massive and uniform character of the fine sediments (Fig. 24; histograms 2 & 4), detailed grain-size analysis reveals also the presence of small-scale, fining-upward sequences (Fig. 24; histograms 1 & 3). At the base of the cycle a silty lamina is present which contains plant detritus. This lamina is succeeded by clayey silt which grade into silty clay towards the top.

A 20 cm thick volcanic ash layer is present over almost the entire western part and was used as marker horizon for this part of the basin. Its position below the surface ranged between c. 1 m (*e.g.* Section 78; Fig. 25) and 3.5 m depth (Section 114, no. G; Fig. 29). The ash layer is made up by two light-coloured strata with silt as the dominating fraction and a third interposed dark-coloured stratum in which sand prevails (Photo 9 and Fig.

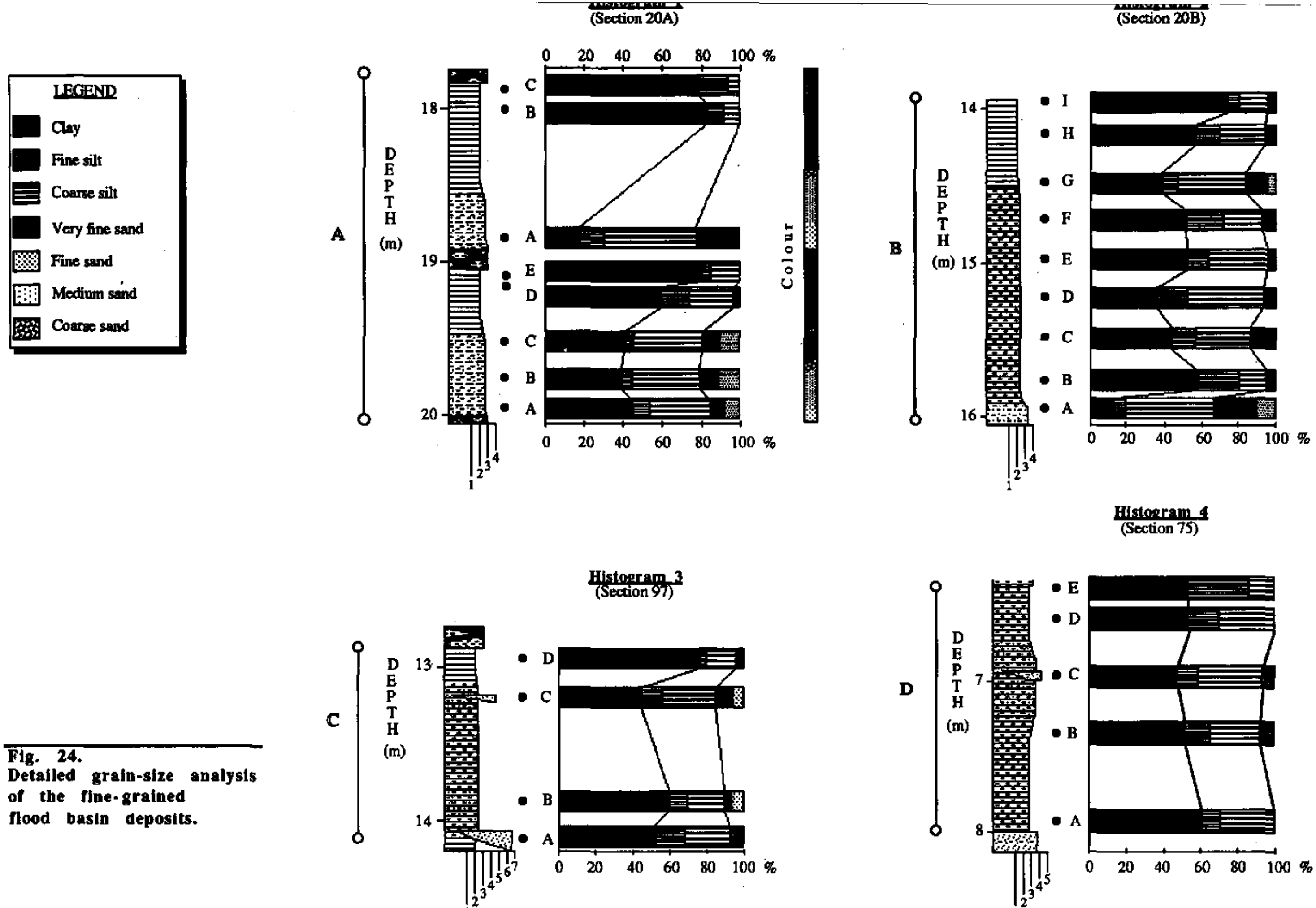


Fig. 24.
Detailed grain-size analysis of the fine-grained flood basin deposits.

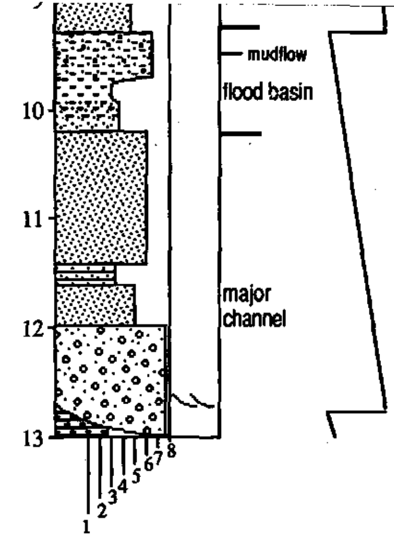
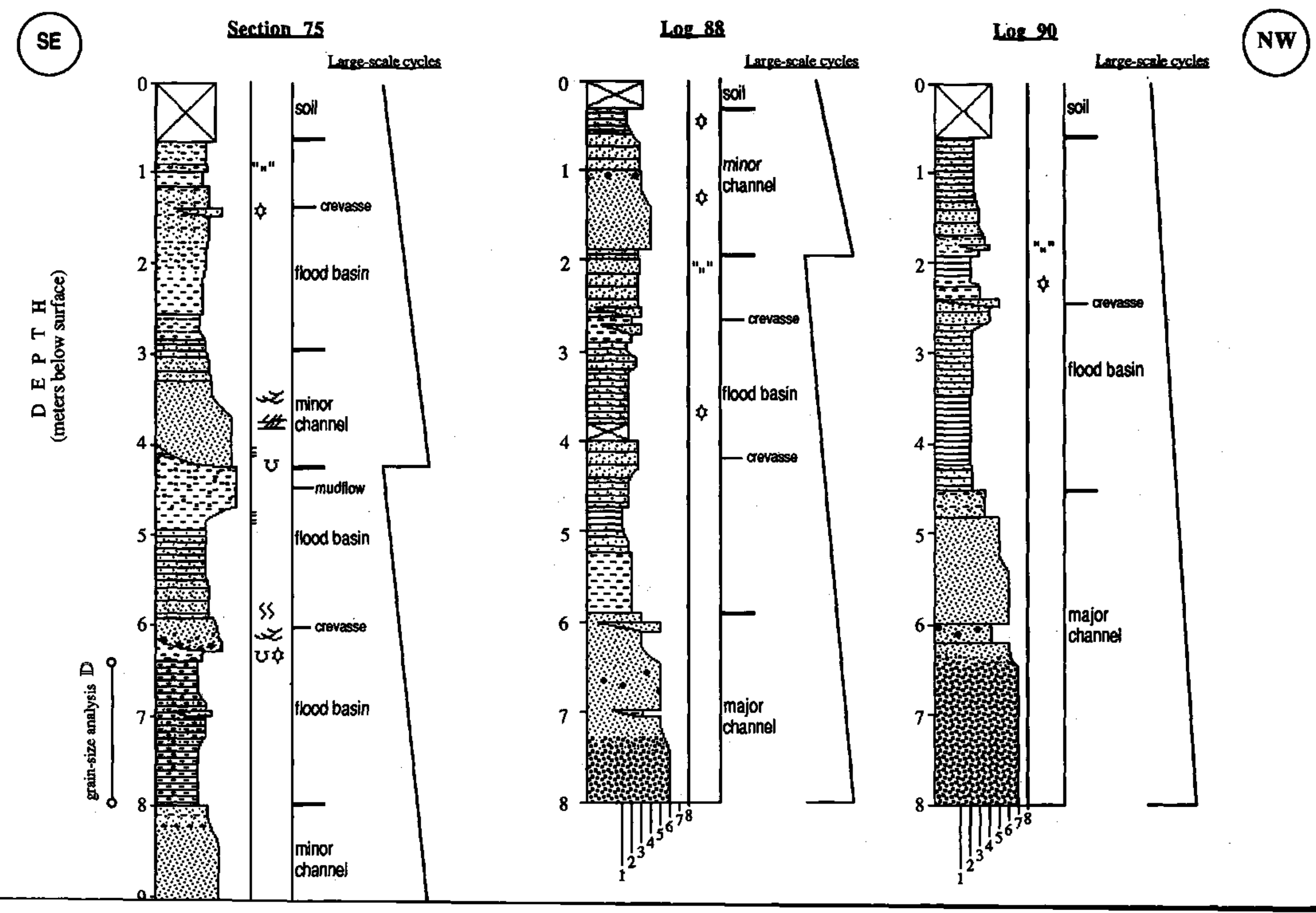


Fig. 23.
SE/NW-oriented cross-section over the western part of the basin illustrating the variability in the sediments. All the sections are representatives of the lithostratigraphical unit B (Appendix I). For the location of the cross-section one is referred to Appendix IV. See Fig. 20 for the legend.



SE
NW
Section 75
Log 88
Log 90
Large-scale cycles
DEPTH
(meters below surface)
grain-size analysis
soil
crevasse
flood basin
minor channel
mudflow
major channel

25). In comparison with the two light-coloured layers the dark layer is enriched with biotite. The bedding planes of the respective laminae of the ash layer is sharp and undulating. Loading and bioturbation structures are found.

Interpretation—The clay-rich sediments are interpreted as flood basin deposits. The overall high proportion of fine-grained overbank sediments indicates that flood basins formed a significant part of the western alluvial plain. The high relative proportion of overbank deposits compared with the channel facies suggests that most of the time a particular site was occupied by a clayey flood basin which was occasionally replaced by an active channel belt. Most of the time the extensive flood basins received clayey material from suspension with an increasing proportion of silt when an active channel was closer to the site. On the whole the flood basins had a submerged character and were inhabited by small and shallow lakes which is reflected by the abundancy of load casts, the sparse presence of mudcracks, the absence of paleosoils and the lack of evidence of a prolonged vegetation cover. The absence of laminated clay and the presence of rooted horizons indicate that lacustrine conditions were never persistent and that plant growth sometimes was possible in the flood basins which implies that the water-table was generally low. The incorporation of stringers and thin sheets of sand in the clayey sediments indicates that water with a certain flow regime was present in the flood basin itself. These sediments are probably deposited by the proximal part of a crevasse splay at short distance from the feeder channel. Clastic material (coarse sand to silt) is transported through crevasse feeder channels into the wetlands and deposited as sheets of sands and silty sediments. Crevasse splay deposits can form a simple sand unit which represents a single episode of deposition. When a crevasse system progrades into the flood basin they may show a composite structure: the coarse-grained channel sediments overlie and sometimes dissects its own silty distal splay sediments resulting in a coarsening-upward sequence which is succeeded by a fining-upward sequence indicating the abandonment of the system. The presence of load casts at the base of the sand sheets suggests sedimentation in water-saturated clayey flood basins. At some locations bioturbated horizons were found on top of the sandy crevasse deposits which demonstrate there was sufficient time for the fauna to establish. The organic-rich laminae which form the base of some fining-upward sequences in the clayey flood basin sediments are interpreted as distal crevasse splay deposits and form the continuation of the crevasse channels. The parallel orientation of the plant detritus indicates subaqueous deposition under quiet water conditions (Coleman *et al.* 1964). Whether the plant remains originate from more ancient -deeper situated- lignitic layers, or reflects a situation in which a vegetation coverage farther upstream was eroded and swept away, remains uncertain. The fining-upward sequences from the silty and organic-rich distal crevasse splay deposits (*e.g.* histogram 1; Fig. 24) are believed to represent the dying out or regression of a crevasse system so that at the end only sediment from suspension was received. Many authors have described in detail the lithofacies of crevasse splay deposits, among others: Elliott (1974), Flores (1981), Gersib & McCabe (1981), Fielding (1984a, 1984b, 1986), Guion (1984), Rust *et al.* (1984). The intercalated silty layers can be interpreted either as overbank deposits (Gersib & McCabe 1981) or represent distal crevasse splay sediments (*e.g.* Guion 1984). Where these sediments prevail with sandy deposits the latter possibility is accepted. The presence of massive gravelly silts must be attributed to an unconfined single flood event, *i.e.*, a density current.

The ash layer is considered as airfall tephra: tephra that falls from the air onto land. The alternation of fine-grained and coarse-grained strata may be explained by a change in composition of material being erupted, or an irregular pulsating voluminous eruption of high intensity lasting several hours or days (Fisher & Schmincke 1984). The sharp and undulating bedding plane follows the small-scale topographic irregularities of the underlying flood basin deposits. According the ^{14}C datings of Section 78 (Fig. 25) this volcanic layer must have been deposited after c. 20,000 years B.P. This dating agrees with the fact that this volcanic ash layer is also found in the eastern part of the basin where it is intercalated in the compact inorganic top layer. The latter have been deposited between 19,000 and 7,000 years B.P. (see PIT 11; Fig. 27 and PIT 2; Fig. 28) and served as a marker horizon for the eastern part of the basin. The most likely source area of the material is the volcanic chain in the Central Cordillera, NW of the Pitalito Basin (Fig. 1). This chain constitutes three volcanoes: Puracé, Pan de Azucar and Paleterá of which the first-mentioned is still active.

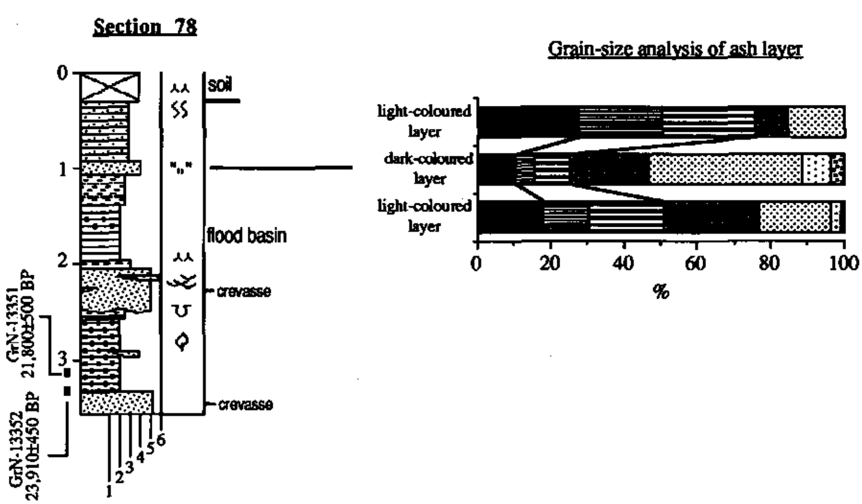


Fig. 25.
A section illustrating the stratigraphical position of the volcanic ash layer (at 1 m depth), its grain-size distribution and the ^{14}C datings at 330 cm and 310 cm depth, respectively. For the location of the cross-section one is referred to Appendix IV. For the legend of the lithological column and for the histograms one is referred to Figs. 20 and 24, respectively.

Large-scale cycles

Description—Besides the aforementioned small-scale sequences within the fine facies also a larger scale cyclicity can be distinguished. These large-scale, fining-upward sequences (several meters thick) are made up by alternating coarse and fine facies and are shown in Figs. 20 to 23. The base is formed by coarse-grained material (gravelly and sandy bodies). The top of these relatively coarse sediments may be abrupt, but usually grade into a topstratum of sandy silt which on its turn grade into the clayey sediments. Regularly, silty deposits with a massive to laminated character (*e.g.* Section 20A, at 18 m, 14 m and 10.60 m depth; Fig. 20) underly the base of a cycle.

Interpretation—The presence of large-scale cycles can be explained in terms of autocyclic (terminology after Beerbower 1964) mechanisms inherent to a fluvial regime. The base of each cycle is made up by channel deposits of the first or second order (minor or major channel, respectively). The prevalence of silty material situated immediately under these channel deposits may indicate that the lateral shifting of the channels was initiated by crevassing and 'announce' the arrival of a channel at that particular site. The succeeding channel facies implies that distributary channels may be the successors of the distal finer-grained crevasse splay deposits. It indicates the progradation and enlargement of the crevasse system into the flood basin which may eventually lead to the development of a channel of minor or even major magnitude. Such processes are commonly known in crevasse systems (*e.g.* Flores & Hanley 1984; Guion 1984). The subsequent fining indicates waning flow and abandonment of the channel due to for instance the lateral shifting of the active channel belt to another location. After abandonment the site received sediments from overbank floods indicating that flood basin conditions were established again until another incursion occurs. Fielding (1984b) suggests that differential compaction leads to transformation of the flood basin into a low-lying depression that was overrun and fed by numerous small channels. According to him the final result will be lateral-stacked channel deposits.

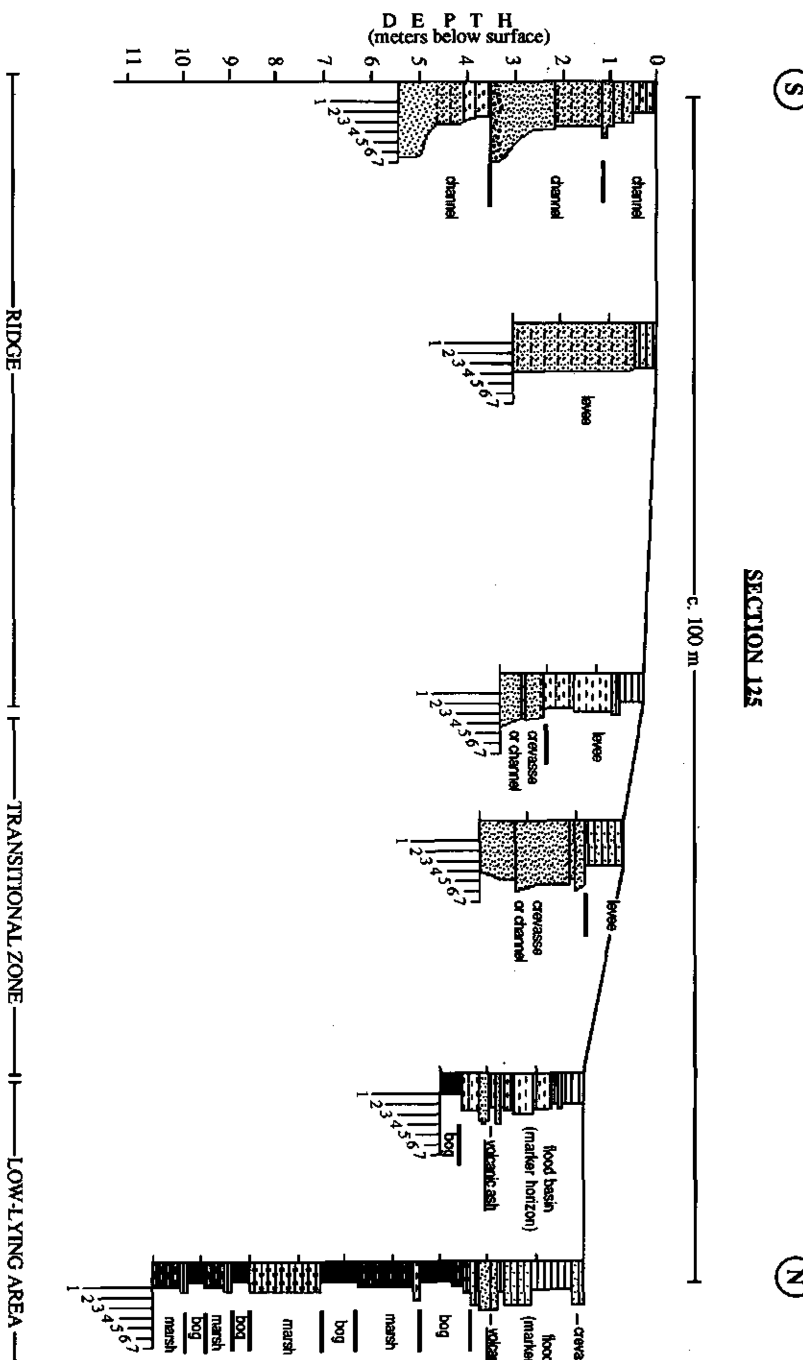
4.1.2 EASTERN PART

The area east of the Guarapas river roughly consist of two different morphological features: ridges and poorly-drained depressions (section 2.2.2).

Ridges

Description—The ridges are mainly built up by coarse-grained deposits which laterally grade into silty material (Photo 7 and Fig. 26). The coarsest sediments are found in the centre of the ridges and vary from gravel (mean diameter of c. 6 mm) to fine sand. Trough and planar cross-stratification has been recognized in these sandy deposits. Measurements reveal a northeastern paleocurrent. On top of these sediments often a thin layer of fine clastic material is found (sandy/silty clay). By means of a geoelectrical survey it appears that coarse material is present at least until 10 m depth and that its lateral distribution perpendicular from the side of a ridge amounts no more than some meters. The finer-grained sediments at the side of the ridge constitute laminated fine sand and silt which are generally devoid of organic material. Further away from the channel, towards the poorly-drained depressions, an increasing amount of clay is incorporated. More laterally, this inorganic material interfingers with the general organic-rich and clayey sediments of the depressions (Fig. 26).

Interpretation—The coarse sediments in the ridges are interpreted as channel deposits. According to the distribution of the sediments the highests flow regimes were


S
N
SECTION 125
c. 100 m
D E P T H
(meters below surface)
0
1
2
3
4
5
6
7
8
9
10
11
channel
channel
channel
levee
levee
crevasse or channel
levee
crevasse or channel
flood basin (marker horizon)
volcanic ash
bog
crevasse
flood basin (marker horizon)
volcanic ash
bog
marsh
bog
marsh
bog
marsh
bog
marsh
RIDGE
TRANSITIONAL ZONE
LOW-LYING AREA

accomplished in the middle part of the ridge. This part therefore, is considered as the channel of the fluvial system (Fig. 26). The finer-grained toplayer of the channel facies is deposited during waning flow. Laterally these channel facies grade into silty sediments. These finer-grained sediments are interpreted as levee deposits. In contrast to several other areas in which the presence of rootlets, bioturbation and pedogenesis on the levees has been reported (Coleman *et al.* 1964; Smith 1983; Kraus 1986), these phenomena could not be distinguished in the Pitalito Basin. The restricted lateral extension of coarse material is illustrative for the fixed character of the channel. The sandy material at the side of the channel (Fig. 26) might be interpreted either as channel deposits or as crevasse deposits. Discrimination between channel deposits and the proximal part of a crevasse splay is difficult: the proximal part of a crevasse is often channelized and resembles much the material of a small-scale lateral migrating main channel. Collinson (1978) and Guion (1984) already depicted this problem of differentation which is enhanced by the fact that in the Pitalito Basin the original channel structure is obscured by compaction and that the data are based on boreholes. Crevassing however, did play a role in the eastern alluvial system as these deposits are very well recognizable in the flood basins as layers of sand encased by fine-grained often organic-rich sediments. The character of the channels in the form of prominent ridges in the landscape can partly be ascribed to compaction of the peat and clay in the flood basins. The depositional processes however, might have played an important role as well; the overbank deposits form natural levees so that channels can built themselves up above the adjacent flood basins. Therefore, in this case, the term relief enhancement is preferred to the term relief inversion as proposed by Havinga (1986). Fig. 26 illustrates the relation between ridges and depressions and their sediments in the eastern part of the Pitalito Basin.

Low-lying areas

Description—The sediments in the depressions consist of alternations of peat and clay which are occasionally intercalated with sandy deposits. The clayey sediments vary from massive sandy clay without organic matter to laminated humic and peaty clay (Fig. 27). Commonly the sandy clay is related with sandy deposits (*e.g.* PIT 11 at 10.30 m and PIT 77 at 5 m; Fig. 27). Plant remains are incorporated in these thin sheets which range from clayey to coarse sand. Usually their base and top are sharp. Some laminae entirely consist of horizontally oriented wood fragments and leaves. Large trunks were encountered and in some horizons rootlets are present. In some borings north of Pitalito an intercalated 5 to 10 cm thick diatomaceous layer was found (Appendix IV) at a depth varying between 1 and 3 m. Its content was analyzed and several *Melosira*-species like *M. distans*, *M. italica* and *M. granulata* were found in this layer (van Dam, pers. comm.).

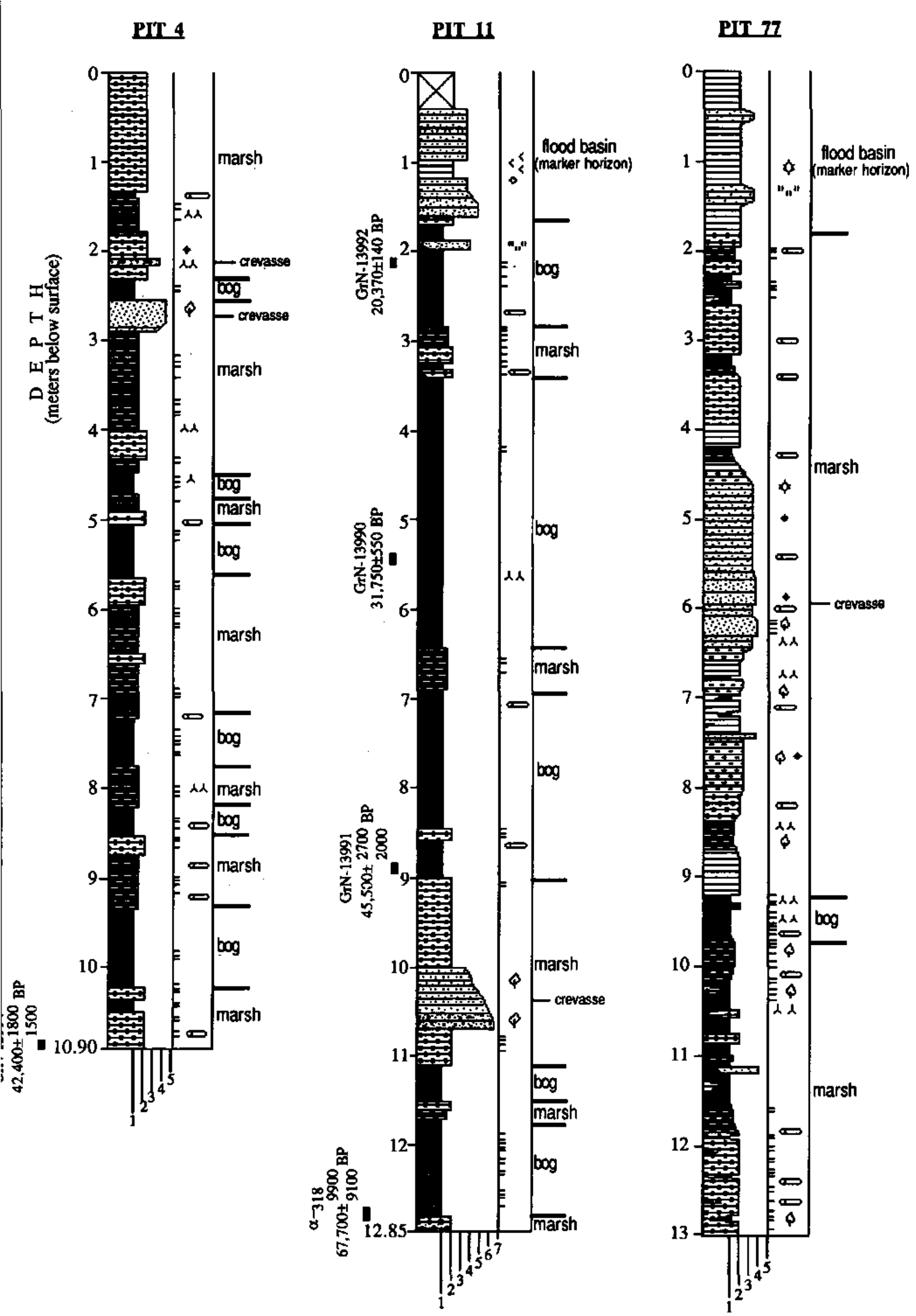
A very conspicuous feature is the c. 2 m thick, compact, sandy/clayey top layer which is present at many sites in the eastern part of the basin (*e.g.* PIT 11 & PIT 77; Fig. 27). The light, green/grey-coloured, sandy clay is completely devoid of organic material and at

Fig. 26.
Cross-section and environmental interpretation of a ridge (at the left) and the adjacent lower-lying flood basin (at the right) which are characteristic for the eastern part of the Pitalito Basin. For the location of the section one is referred to Appendix IV. See Fig. 20 for legend.

some locations contains gypsum (PIT 11; Fig. 27). The transition from the deeper organic-rich layers to these sterile clays is abrupt. This layer served as a marker horizon in the eastern area. It appears that in the northeastern part of the Pitalito Basin (La Coneca) this same horizon is not found at the top like elsewhere, but at a depth varying between 3 and 5 m (*e.g.* PIT 2 & PIT 109; Fig. 28). The succeeding sediments constitute peat (PIT 2) with an increasing proportion of inorganic clay proximate to the northern mountain front (PIT 109).

Interpretation—The fine clastics in the poorly-drained depressions are considered as flood basin deposits. The low-lying areas in the eastern half of the Pitalito Basin represent former flood basins which were alternately occupied by peat bogs, marshes and lakes (marshes do not drain seasonally and occasionally receive floodwater charged with suspended load whereas peat bogs usually are out of reach of floodwater; terminology after Smith 1983). The bogs, areas out of reach of floodwaters, are characterized by peat development. The presence of large trunks implies that trees and larger plants could inhabit the bogs. In some horizons rootlets are present which indicate stable conditions and possibly somewhat drier local circumstances in the bogs. The flood basins in the Pitalito Basin were relatively small and always within the reach of the active channels so that floodwater and sandy crevasse splay deposits could reach them. Therefore, partially dependent on the distance between the site and the active channel, the peat bogs generally were short-lived and frequently replaced by marshes (compare for instance PIT 77 with PIT 11 which are situated near a fossil channel and at some distance from it, respectively). The laminated, organic-rich to peaty clays are interpreted as sediments originating from floodwater which frequently reaches the lower-situated flood basins. According the terminology of Smith (1983) these sediments are representative for the marsh facies. Smith (1983) describes the sedimentary facies of interdistributary lakes as clay and silty clay which mostly lack variability and distinguish them from a marshy environment by the absence of organic material. The fact that the flood basins in the eastern part of the Pitalito Basin are almost entirely built up of organic-rich materials implicates that the lake facies was rarely present in the eastern half of the Pitalito Basin. However, the content of the diatomaceous layer do not exclude the presence of stagnant water bodies: several *Melosira*-species prefer a mesotrophic, neutral to weakly acid lacustrine environment (van Dam, pers. comm.). Therefore, it is suggested that marshes and bogs form the main sedimentary environment of the interdistributary areas and that small, shallow, not persistent, stagnant water bodies frequently received organic-rich sediments from suspended load. The sandy intercalations are interpreted as crevasse splay deposits. The grain size of these sediments largely depends on the fact whether we are dealing with proximal or more distal locations in the splay lobe. The sharp base of the sand sheets indicate a sudden incursion at that particular site whereas the sharp top is indicative for a fast abandonment of the respective crevasse system. The top layer of

Fig. 27.
Several lithostratigraphic sequences which are representative for the eastern flood basin facies. They all belong to unit D (Appendix I). At the right side of each column the respective sedimentary environment is shown. For the location of the borings one is referred to Appendix IV. See Fig. 20 for legend.


PIT 4
PIT 11
PIT 77
D E P T H
(meters below surface)
marsh
crevasse
bog
flood basin
(marker horizon)
GrN-13992
20,370±140 BP
GrN-13990
31,750±550 BP
GrN-13991
45,500± 2700 BP
2000
α-318
67,700± 9900 BP
9100
42,400± 1800 BP
1500
10.90
12.85

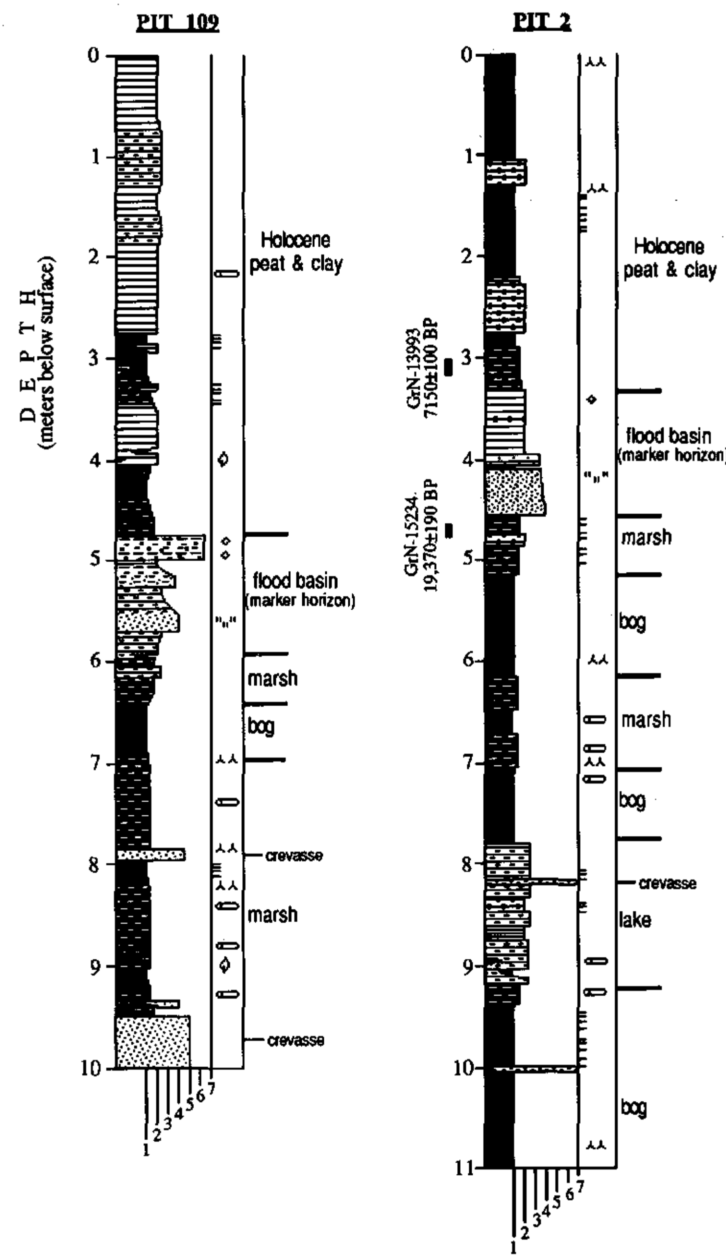


Fig. 28.
Two lithostratigraphical sequences which are representative for the facies in the area of La Coneca. This area is the only place where sediments of Holocene age are found. Both borings belong to unit F (Appendix I). At the right side of each column the respective sedimentary environment is shown. For the location of the borings one is referred to Appendix IV. See Fig. 20 for legend.

compact sandy clay can be found in the depressions over the entire eastern area. The peat just below the base of this layer yields a ^{14}C age between 20,000 and 19,000 years B.P. (PIT 11; Fig. 27 and PIT 2; Fig. 28). From that period on the flood basins were choked with a clastic influx impeding peat growth. The presence of gypsum shows that the interchannel flood basins frequently fell dry. After the deposition of this inorganic layer no more sediment was deposited on top of it with the exception of the La Coneca area (Appendix I). In this area peat and fine-grained material are found on top of the marker horizon. The increasing proportion of clastic material towards the mountain front suggests that these sediments originate from the northern mountain range. A constantly high water table, a necessity for peat development, is favoured by the impermeability of the compact marker horizon. A ^{14}C age of 7150 years B.P. (PIT 2; Fig. 28) indicates the time that peat development started at that site.

Based on the distribution of the sandy ridges and organic-rich depressions two units can be distinguished in the eastern part of the basin (Appendix I):

Unit D/ Inorganic clay (marker horizon) on peat and peaty clay (former flood basins; *e.g.* PIT 11, PIT 107 & PIT 4; Fig. 27)

Unit E/ Sand (former channels) on peat and peaty clay

A third unit could be distinguished with the help of litho-and chronostratigraphy:

Unit F/ Holocene peat on unit D-deposits (*e.g.* PIT 2 & PIT 109; Fig. 28).

The description of unit E needs some explanation: The total thickness of the sandy ridges could only be estimated as being more than 10 m thick by geoelectrical methods as the sediments did not allow a deep perforation. The thickness of organic-rich material is more than 10 m as can be seen from the diverse borings and may even reach far beyond this figure when the geoelectrical soundings are taken into account (Chapter 3; Fig. 16). It is therefore believed that channel deposits are situated on top of organic-rich sediments. However, it may not be excluded that the channel deposits reach deeper than about 10 m.

4.1.3 TRANSITIONAL ZONE

Immediately west and east of the Guarapas river a transitional zone is present. This zone reveals similar sedimentological characteristics as the western area: the absence of organic-rich material, and comparable geomorphological features as the eastern area: E-W oriented ridges which enclose depressions (Appendix I).

Description—On the whole, the sediments in the ridges consist of medium sand to gravel (Fig. 29) and can be traced across the deeply incised valley from west to east where they become more prominent. The lower areas are made up of inorganic fine clastics: clayey and silty material with intercalations of about 1 m thick sandy material. The transition from the inorganic sediments in the west to the organic-rich sediments in the eastern part of the basin is located just east of the line of the Guarapas river (Appendix I) and is reached within a horizontal distance of about 400 m (PIT 107 & PIT 106, respectively; Fig. 30).

Interpretation—The ridges are interpreted as channel deposits and the finer material in the depressions as flood basin deposits. The channels of the fluvial system in this zone are much alike the eastern ones: they are relatively stable and built themselves above the flood basin. The flood basins however, are more alike the western ones: they frequently received sediments during overbank floods impeding plant growth. Based on the E/W-orientation of the ridges and on some measurements on cross-stratification it appears that on the whole, flow direction was from southwest to northeast. The fact that the ridges can be traced over the deep valley of the Guarapas river indicates that this phase of last-stage basin infill preceeded the erosional processes of this river. UTD datings and geoelectrical data suggest that the position of the transitional zone continued for tens of thousands of years without any significant change. This and the relatively short horizontal distance in which one sediment type changes into the other, imply that the transition can be considered to be stable in time and is relatively abrupt.

Based on the lithological content of this transitional area two units can be discerned (Appendix I):
Unit B/ See page 71
Unit C/ Sand on B-deposits (former channels)
The same remarks given for unit E are valid for unit C.

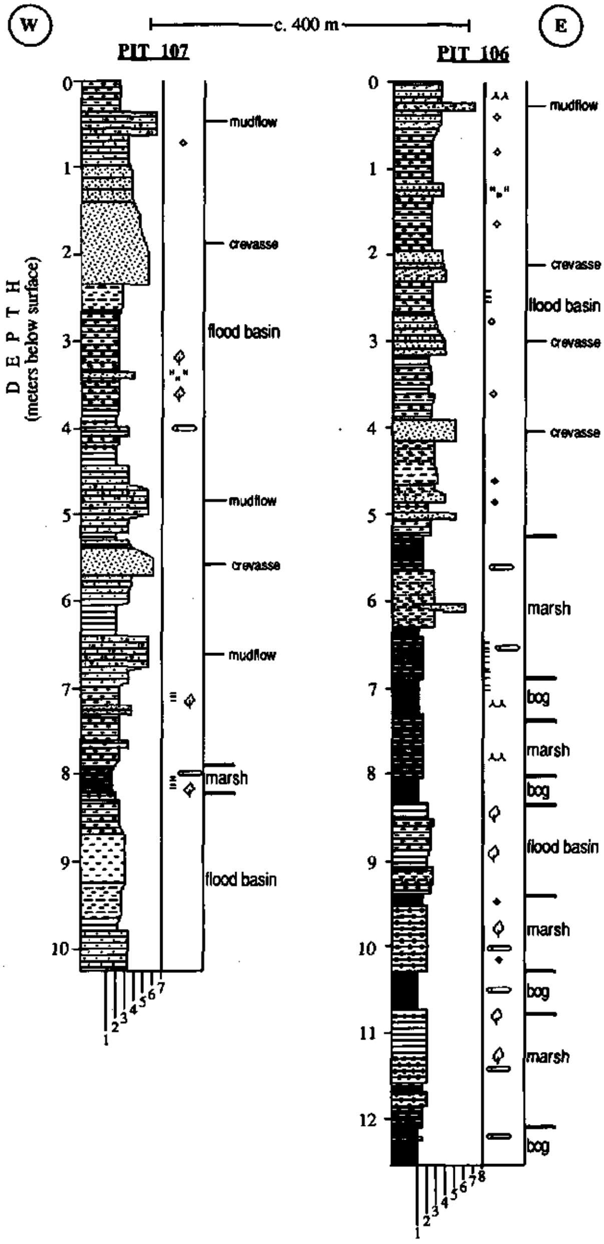


Fig. 30.
Two lithostratigraphical sequences illustrating the relative abrupt transition from the inorganic sediments in the west (PIT 107=Unit B) to the organic-rich flood basin deposits in the east (PIT 106=Unit C). Over a short distance of about 400 m there is a markable increase in the organic content. For the location of the borings one is referred to Appendix IV. See Fig. 20 for legend.

4.2 CONCLUSIONS AND DISCUSSION

The last stage of sedimentary infill of the Pitalito Basin is reflected by an eastward basin-wide prograding fluvial system which preceeded the actual erosional state. Sediment transport was from west to east. The architecture of the western, proximate part of the fluvial system differs considerably from that of the eastern, distal part. In this discussion it is tried to find out what caused the differentation between the western proximate part of the fluvial system and the eastern distal part.

Sedimentary environment of the western area (Fig. 31A)
During the last stage of sedimentary infill of the Pitalito Basin the western half was dominated by regularly ponded, wide flood basins which were occasionally invaded by channels. The fluvial system consisted of northeast-heading channels which shifted over the area possibly through crevassing. The large-scale cycles reflect the lateral migration and short-term presence of an active channel system at a particular site. This process was favoured by the apparent absence of prominent, vegetated natural levees. Small-scale cycles represent minor fluctuations in the flood basin itself. The total absence of organic accumulation in the flood basins probably can be ascribed to the high rate of detrital influx. The sedimentary aspects of the western half is much like the Breconian cyclothem described by Allen (1964). He suggested that this cyclothem records the development of a flood basin sequence in which a possible braided shallow river system ultimately was replaced by backswamp conditions. The river system in this part of the basin may represent an intermediate between a braided and anastomosing river model as proposed by Casshyap & Tewari (1984).

A conspicuous feature of the western part of the fluvial system is the high relative proportion of overbank versus channel deposits (60% to 30-40%, respectively). Although several authors describe present-day rivers where the average amount of overbank material appears to be small, there is an increasing believe that this might not always have been the case in the past (*e.g.* Friend 1978; McClean & Jerzykiewicz 1978). The absence of narrow valley walls have favoured the thick accumulations of overbank sediments in the Pitalito Basin.

Sedimentary environment of the eastern area (Fig. 31B)
The morphological and sedimentological features in the east are consistent with a fossil anastomosing river pattern as described by several authors (Flores & Hanley 1984; Schumm 1968; Smith & Smith 1980; Smith & Putman 1980; Rust 1981; Rust *et al.* 1984). They suggest that anastomosing channels result from prograding crevasse channels into the flood plains which eventually leads to stabilization. Smith (1986) gives the following definition of anastomosing rivers:

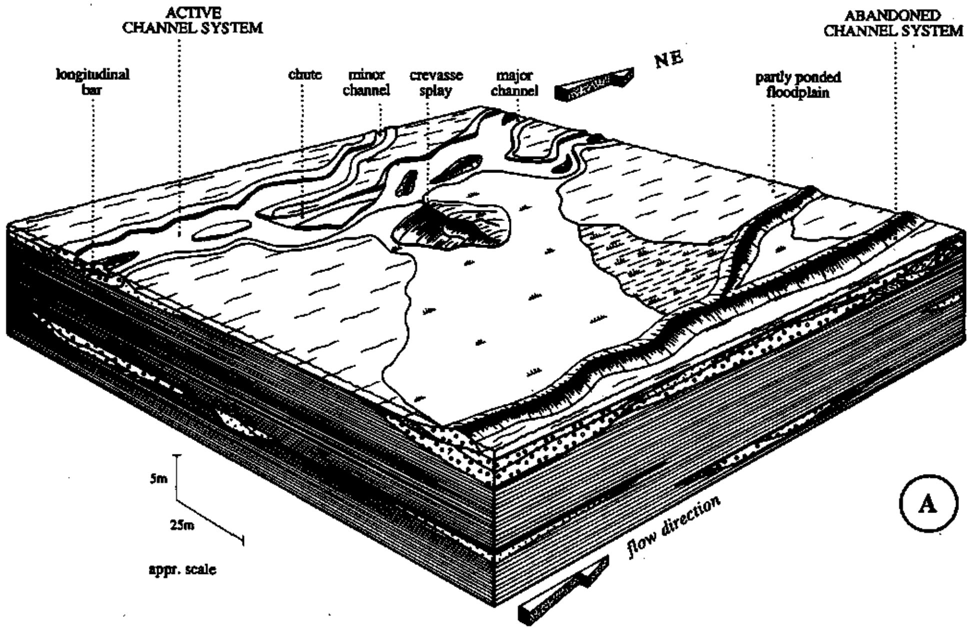
> *"Anastomosing rivers consist of low-energy, multiple, interconnected, laterally stable, deep sand-bed channels, confined by prominent levees."*

He distinguished six depositional environments: channel, levee, crevasse-splay, lake, marsh and peat bog. The ridges can be interpreted as old, nowadays inactive, sand-bed channels and the depressions as interchannel flood basins.

Characteristic for this river type is the stability of the channels which can partly be ascribed to the cohesive character of the flood basin sediments and the very low gradient in the eastern area which is practically reduced to zero: Fig. 4A shows that the eastern part of the Pitalito Basin has an almost flat topography which is in contrast with the slightly eastward-dipping western part. This also explains the low-energy of the eastern fluvial system and the development of natural levees which were less frequently overflowed during bankfull stage. Therefore, the flood basins did not receive the amount of clastic detritus as compared with the western ones and peat could accumulate. According to Smith and Smith (1980) and Smith (1983) the presence of levees flanking the channels are characteristic for anastomosing rivers. Thick and laterally extensive autochtonous peats may form in backswamps between well-developed levees. These low-situated sites remain poorly-drained and are sustained by a groundwater table that rose near the surface creating a stable reducing environment (Flores 1981). It must be stated however, that Rust (1981) and Rust & Legun (1983) describe discontinous poorly developed levees in an arid-zone anastomosing river type.

The characteristic anastomosing pattern of the eastern part of the fluvial system can be found all over the world in very different climatic environments: semi-arid regions in Australia like the Lake Eyre Basin (Wells & Callen 1986), near the Rocky Mountains in Canada (Smith 1983), the humid and warm tropics at the northern part of Colombia (Smith 1986) and in former times during a temperate and humid climate in the Rhine/Meuse delta in the Netherlands (Van der Woude 1981). This variance suggests that other factors than climate are controlling this type of river pattern. Smith & Putnam (1980) and Smith (1983) suggest that a rising base level due to subsidence or cross-valley alluvial fans might induce anastomosis.

It appears that the differences in the alluvial architecture in the Pitalito Basin is largely determined by channel behaviour: instable channels in the west and stable channels in the east. Horne *et al.* (1978) and Fielding (1984b) stated that the presence of rapid, structurally controlled subsidence may lead to a vertical stacking pattern of the sandbodies whereas differentational compaction leads to diagonal stacking. This may explain the complete different alluvial architecture of the western part (rapid lateral shifting of the channels leading to diagonal stacking due to relatively low subsidence rates) compared to that of the eastern part (stable channel types that leads to vertical stacking due to relatively high subsidence rates). From this it may be concluded that differentation in alluvial architecture indeed was determined by tectonic and thus allocyclic control (Beerbower 1964). This conclusion is supported by the fact that the transition between the two different fluvial systems is situated along the line of the N/S-oriented fault (no. 5; Fig. 12). This tectonically determined feature separates the basin in a shallow western and a deep eastern part (Chapter 3). A further extended conclusion might be that subsidence rates in the east are higher than in the west. If so, this would



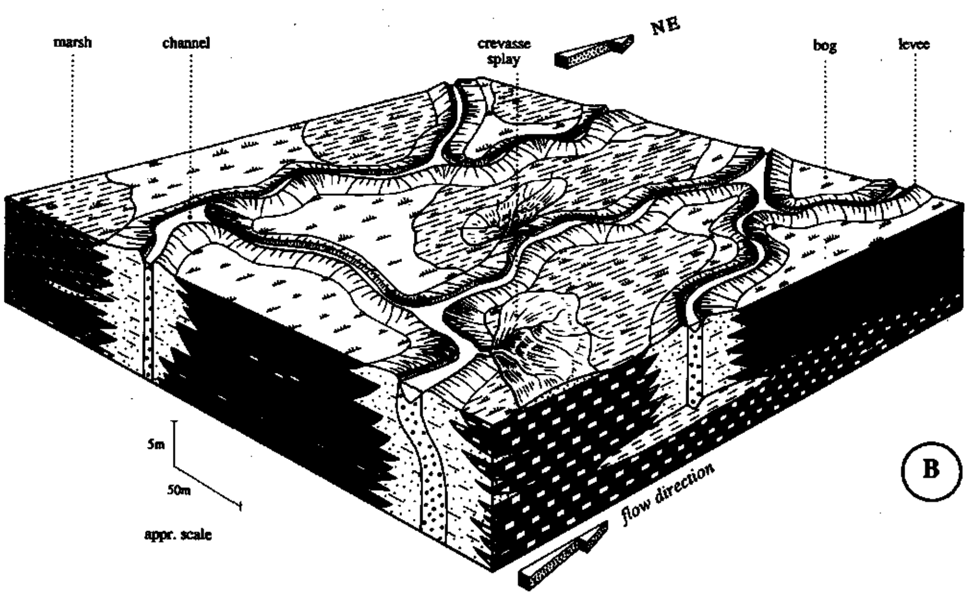


Fig. 31.
Two block diagrams. A/ Schematic paleogeographical model of the western proximal part of the fluvial system; B/ Schematic paleogeographical model of the eastern distal part of the fluvial system. See Fig. 20 for legend.

have its consequences for the interpretation of the evolution of the basin especially for phase 1 and 2 (section 3.3). This however is beyond control as datings from relatively greater depths are absent. The available datings from the upper 20 meters of sediments in the west (*e.g.* Section 20B at 18 m depth; Fig. 20) are in the same order of age as in the east (*e.g.* PIT 4 & PIT 11; Fig. 27) and do not support the theory of a slowly subsiding western block and a faster subsiding eastern block. Therefore, the proposed sedimentary evolution as given in section 3.3 is maintained.

One of the more conspicuous aspects in an anastomosing river environment is the relatively fast aggradation rates. Smith (1983) calculated rates of accumulation in anastomosing systems which roughly average between 15-60 cm/100 years. With the help of several radio-carbon datings sedimenation rates were calculated over the last 60,000 years for the Pitalito Basin (section 5.2.2.3). They appeared to be far less (≈ 2.5-3.0 cm/100 years) than the ones calculated by Smith (1983). This difference may be partly explained by compaction of the peats. According to Rust *et al.* (1984) and Smith (1986) the relatively low sedimentation rates may favour the development of thick organic deposits as the fluvial dynamics are more stable compared to the areas where tectonics differ in time.

Around 20,000 years B.P. the flood basins received considerable amounts of clastic material (marker horizon). This feature suggests a change in the flow regime of the alluvial system leading to frequent overbank flooding of the natural levees: channels are unable to cope with the abnormal and probably infrequent discharge after 20,000 years B.P. and sheet flows cover the lower-lying areas. The flood basins fell frequently dry and the accumulation of organic material came to an end caused by this high detrital influx. These phenomena are typical for the semi-arid ephemeral streams where a sparse vegetation cover and peak floods lead to sheet-wash erosion and an irregular supply of sediment-rich water into the distributary system. All this points to a change in the environmental conditions characterized by a relative low annual precipitation around the Pitalito Basin. Several authors indeed report a climatic change at high elevations (around 2500 m altitude) around 20,000 years B.P. in the Colombian Andes (*e.g.* Van Geel & van der Hammen 1973; Salomons 1986; Kuhry 1988). Around this time the climate became extremely dry and cold at these altitudes (Fuquene Stadial).

The Late-Glacial/Holocene transition is a notable period for the Pitalito Basin; somewhere between 17,000 and 7,000 years B.P. sedimentation came to an end and the course of an infilling fluvial system with a NE paleocurrent changed into the modern NW-heading system characterized by erosional processes (section 2.2.1). The fact that the 90°-turn of the drainage system and the start of erosional processes are contemporaneous with a very dry and probably cold period (Chapter 5) suggests that both events might have been induced by climatic changes instead of tectonic control. In the case of the Pitalito Basin we may think of an increasing erosion rate due to a sparse and open vegetation cover such as nowadays present in the Upper Magdalena Valley below c. 1100 m altitude. The bare soil around the basin may have been eroded more easily and sediment supply to the river system increased considerably as we have seen from the inorganic top layer. Accelerated headward erosion in the NW part of the basin may have diverted the paleocourse of the Guachicos river from NE to NW and the base level changed from c. 1300 m altitude (Pitalito Basin) to c. 1200 m (Magdalena river). The lowering of the base level with c. 100 m favoured erosional processes and the rivers incised in their own deposits. A similar change from accumulation to incision is also registered along the amazonian Caquetá river (Absy *et al.*, unpubl.); the low river terrace was eroded somewhere between 35,000 years ago and 12,500 years B.P. Between 12,500 and 10,000 years B.P. the river rose and filled in the river valley. Showers & Bevis (1988) record an increase of Amazon freshwater discharge around 13,500 years B.P. and the flooding of the Amazon shelf at 11,500-9,000 years B.P. According Showers & Bevis (1988) an increase of pluvial activity is the most plausible hypothesis for the initiation of freshwater discharge. Apparently the Amazon Basin was influenced by climatic changes around 12,000 years B.P. As the basin is not very susceptible to tectonic activity, the changes in the Caquetá river regime around this time may be ascribed to climatic control. We may conclude that both, climatic as well as tectonic controls might be responsible for the drastic changes in the Pitalito Basin around the Late-Glacial/Holocene transition. Possibly both controls worked in concert.

From that time on the only site where sediments are deposited is in the La Coneca area along the dextral strike-slip fault. The distribution of these Holocene sediments suggests that the basin floor is tilted towards the northern margin (see also Fig. 19). This is supported by the present geomorphology and the distribution of Holocene sediments: small debris-flow-dominated, tilted alluvial fans prevail along the northern margin whereas from the opposite basin margin larger streamflow-dominated fans were deposited (Appendix I). These streamflow-dominated fans are made up by better-sorted and finer-grained detritus and have gentler slopes. Comparable morphological and sedimentological phenomena are also found in other strike-slip basins like the Hornelen Basin in Norway and the Ridge Basin and Little Sulphur Creek Basins both in California (*e.g.* Nilsen & McLaughlin 1985). Hooke (1972), Steel (1976) and Nilsen & McLaughlin (1985) describe these features as being characteristic for strike-slip basins and explain them by tilting of the basin floor towards the strike-slip fault. Besides a northern gradient, tilting also may have a gradient towards the west. In that case, the apparently flat topography of the eastern part of the basin may have instead a slightly west/northwestern-directed dip which could not be detected by the digitized areal photographs. This might explain the sharp-angle turns from NE to NW of the intermittent streams in this part of the basin and the actual position of the Guarapas river; the river would be present in the knickpoint of an eastward-dipping western part and a west/northwestern-dipping eastern part. The paleocourse of the Guachicos river would be forced towards the northern basin margin and finally found its way out through the NW outlet.

CHAPTER 5

LATE QUATERNARY VEGETATIONAL AND CLIMATIC HISTORY OF THE PITALITO BASIN

The importance of the low altitudinal zones (1000-2300 m altitude) for the agricultural sector has been mentioned in Chapter 1. A paleoenvironmental study in these zones would contribute to a better understanding of the effects of human activity on natural processes. Untill now the majority of paleoenvironmental studies in the Andean regions of Colombia are concentrated on higher elevated zones, *viz.* above 2500 m altitude (*e.g.* Van der Hammen & González (1960a, 1960b, 1963), Van Geel & van der Hammen (1973), Schreve-Brinkman (1978), Hooghiemstra (1984), Melief (1985), Salomons (1986) and Kuhry (1988). The main reason is the availability of good lake and peat deposits at these altitudes. These studies reveal important information about the altitudinal shifting of the forest belts in time, due to changes in temperature and precipitation. The palynological studies of the tropical lowlands (lower than 500 m altitude) mostly reveal information about the precipitation regime only (*e.g.* Wijmstra & van der Hammen 1966; Van der Hammen 1972, 1974; Absy 1979). Therefore, palynological studies from lower montane regions are of importance to link the information from high altitudinal belts with data from the tropical lowlands. Some interesting papers dealing with the altitudinal zone below c. 2500 m are by Van der Hammen (1974), Monsalve (1985) and Cleef *et al.* (in prep.).

The Pitalito Basin is one of the sites where the abundance of organic-rich sediments, the presence of an impermeable clayey top layer and the high water table provides well-preserved pollen. This makes it possible to form a link between the palynological records from the higher altitudinal zones and the record from lower elevations. With the help of all these records a reconstruction of the paleoclimatical and paleovegetational conditions is made. These conditions, combined with sedimentological aspects, may give information to what extent climatic variation played a role on changes in the alluvial architecture.

5.1 Variety in modern pollen rain and vegetation with altitude

In this section attention is briefly paid to the relation between altitude and the variation in actual pollen rain and vegetation cover. Only the most important zonal taxa present in the pollen diagrams (Appendix V) are considered.

Former palynological studies like those of Van Geel & van der Hammen (1973), Schreve-Brinkman (1978), Hooghiemstra (1984) were concentrated along lat. 5°N where the wide Magdalena Valley (c. 50 km wide) separates the Eastern and the Central Cordillera at an altitude of about 300 m. At that latitude the zonal vegetation belts of higher elevation on both slopes of the cordilleras are separated by dry to xerophytic woodland which is present in the valley (see Fig. 9). Coming from the north and passing through the Upper Magdalena Valley this vegetation type has its most southern extent around the town of Altamira (Fig. 1) up to an altitude of 1000 m. This means that such a xerophytic vegetation is lacking in the narrow Magdalena Valley floor at the latitude of the Pitalito Basin. The absence of a broad valley floor makes it possible that vegetational elements in and around the Pitalito Basin not only originate from the Eastern Cordillera but from the Central Cordillera as well. Therefore, existing data on recent pollen rain and actual vegetation from both cordilleras are taken in consideration.

For the Eastern Cordillera relevés of the stands of vegetation were made in 1967 and 1968 by T. van der Hammen, R. Jaramillo & M.T. Murillo in the Eastern Cordillera at lat. 5°N. At the same sites moss samples were taken to study the modern pollen rain. These data were partly elaborated by Grabandt (1980). She plotted the results of the vegetational and pollen analysis against the altitude (=relevés). For the Central Cordillera the data are from the relevés (c. lat. 5°N) made during the ECOANDES-expedition in 1979. They were partly elaborated by Melief (1985). At lat. 5°N the Central Cordillera reaches heights up to 5000 m whereas the summits in the Eastern Cordillera, except for the Sumapaz, do not exceed 4000 m altitude. In the Eastern Cordillera therefore, the upper limit of the vertical range of some taxa might be restricted merely due to mountain topography. A distinction was made between relevés originating from east and west-facing slopes. For the Eastern Cordillera 75 relevés were available for the west-facing slopes and 84 for the east-facing ones; for the Central Cordillera these figures were 25 and 24, respectively.

Fig. 32 illustrates the relation between altitude, the face of the slope, the modern pollen deposition (in percentages of the pollen sum) and the estimations of the plant cover[1)] of the most important zonal taxa. Along the Y-axis the altitude is shown in classes each with an interval of 250 m. As far as possible, the vegetation cover and the actual pollen deposition was calculated for each relevé. Subsequently, the mean of the respective percentages of those relevés belonging to the same altitudinal class (Y-axis) were calculated and plotted along the X-axis. The occurrence of a pollen type that is not represented in the corresponding vegetation is ascribed to 'background effect' (González *et al.* 1965). This effect is caused by a relatively high pollen production and good dispersal so that pollen is not only deposited on the spot but is transported by air to places where the tree or plant does not grow. A clear example of this effect is illustrated in the Central Cordillera by *Alnus* (Fig. 32A): pollen grains of this tree are found along the entire altitudinal range whereas trees are absent in the vegetation relevés.

In each histogram the corresponding t/p-value (Grabandt 1980) or v/p-value[2)] (Melief 1985) is given for each element. This value describes the relation between the modern

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1) *Grabandt (1980) estimated the plant cover in percentages of the total number of individuals whereas Melief (1985) expressed the plant cover in percentages of contour cover. Therefore, the differences in the vegetation percentages for the Eastern and the Central Cordillera can partly be explained by this different approach.*

2) *The v/p-values of Melief (l.c.) describe the relation between pollen and vegetation in the same way as do the p/t-values of Grabandt (l.c.). The values however, are not comparable due to a different definition for the vegetation cover (see note 2).*

pollen rain of an element and the representation of this taxon in the present-day vegetation. When t/p (v/p)=1 this means that the taxon in question is equally represented in both pollen rain and in the stand of vegetation. When t/p (v/p) < 1 this means that the taxon in question is over-represented in the pollen rain and vice versa. Generally, the lower the values the higher is the pollen productivity.

-*Alnus* (Fig. 32A) Stands of this tree are found in the Eastern Cordillera between 2500 m and 3100 m altitude. Locally, this tree occurs up to an altitude of 3500 m (Van der Hammen 1974). Low percentages of *Alnus* pollen are found over a wide altitudinal range and reflects the well-known high pollen productivity of this tree. Hooghiemstra (1984) ascribes the first 20% of *Alnus* pollen to background effects. In our case this would mean that no *Alnus* forests were present *in situ* as pollen percentages remain below 20%. However, from Fig. 32A it appears that alder trees were present. From the Central Cordillera no stands of *Alnus* are recorded as the relevés were made of zonal vegetation types. *Alnus* however, may be abundant on wet slopes. The presence of *Alnus* in the pollen rain of the zonal vegetation is therefore a 'background effect'.

-*Hieronima* (Fig. 32B) In the Eastern Cordillera stands with *Hieronima* reach a maximum between 2500 m and 2800 m altitude at the east-facing slopes. Higher pollen percentages however have been registered at far lower levels. For the west-facing slopes a maximum is discernable around 1300 m altitude for both vegetation and pollen deposition. In the Central Cordillera *Hieronima* and its pollen only appear along the humid west-facing slopes between 2000 m and 1500 m altitude.

-*Quercus* (Fig. 32C) Along the west-facing slopes of the Eastern Cordillera a maximum of both pollen rain and vegetation cover is reached between 2100 m and 3000 m altitude. With the exception of the outer eastern flank where *Quercus* trees are absent, the maxima of pollen rain and vegetation cover along the inner east-facing slopes are situated from 2500 m to 3200 m altitude. Relatively high pollen percentages have been registered up to 4000 m altitude along the western side. In the Central Cordillera a clear distinction between west- and east-facing slopes is present. Apparently, *Quercus* prefers the somewhat drier east-facing slopes to the more humid western slopes. The high maximum is reached between 2500 m and 3000 m altitude.

-*Weinmannia* (Fig. 32D) In the Eastern Cordillera maximum percentages are reached for both the modern pollen rain as well as vegetation cover around 3000 m altitude. *Weinmannia* trees occupy a wide altitudinal range (2300-3500 m). The relatively high values in vegetation and pollen spectra at an elevation of 1300 m are ascribed to local favourable edaphic conditions (Grabandt 1985). The increasing pollen percentages above 3750 m altitude must be ascribed to the so-called 'superpáramo effect' (Grabandt 1985). This effect is caused by a very low local pollen production due to a sparse vegetation cover: a situation which is often found in the superpáramo. In that case pollen which is blown upslope from lower altitudes may become dominant in the pollen rain. In the Central Cordillera the percentages of modern pollen rain and vegetation cover are higher at the west-facing flanks than at the eastern ones. On the western slopes maxima are reached between 2875 m and 3875 m altitude for modern pollen rain of *Weinmannia*

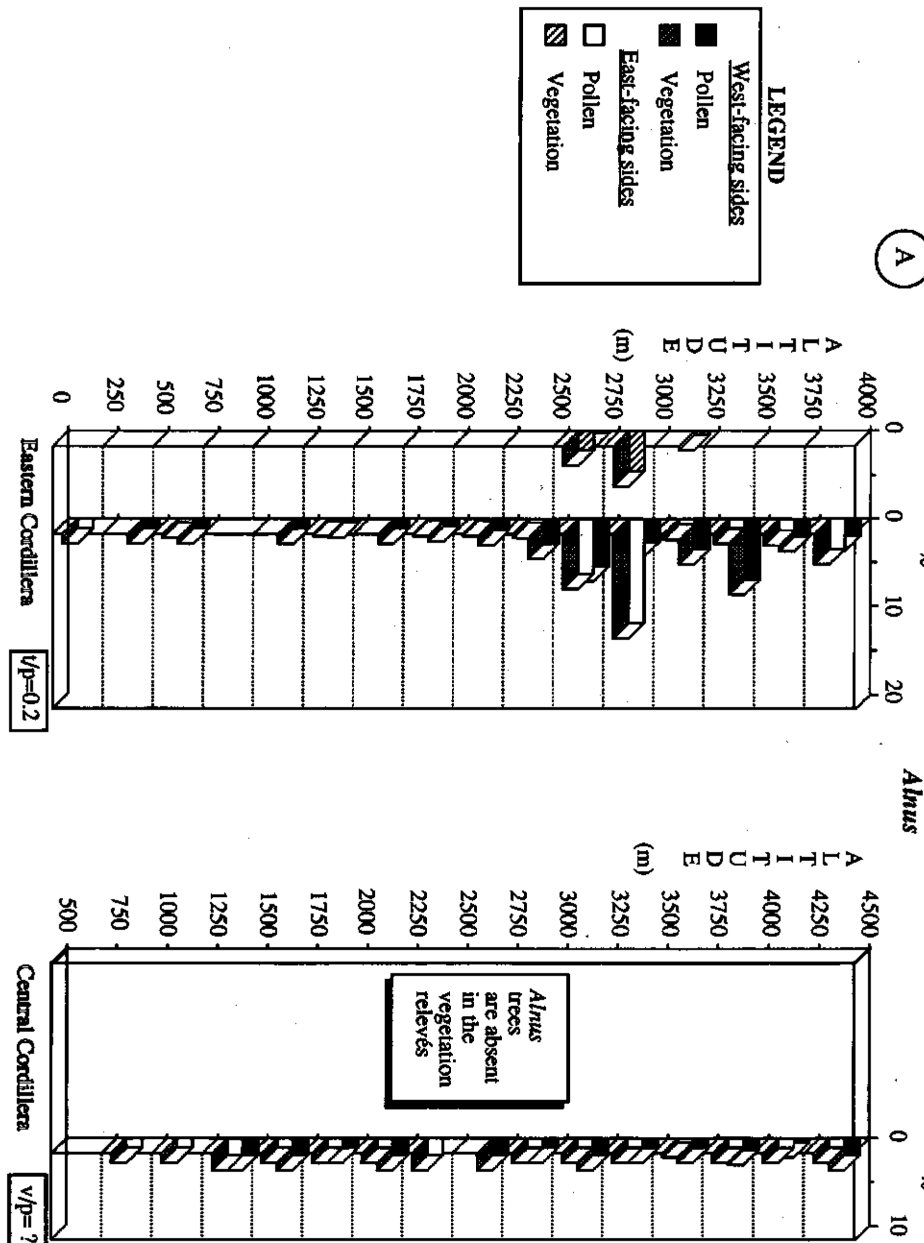
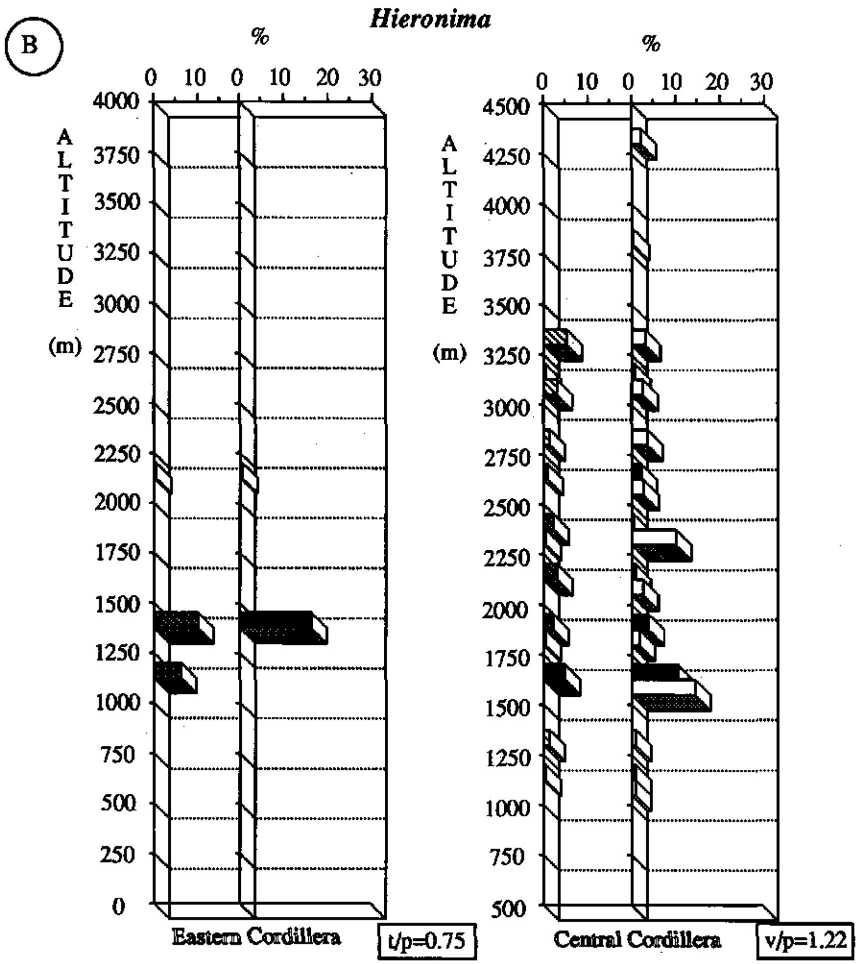
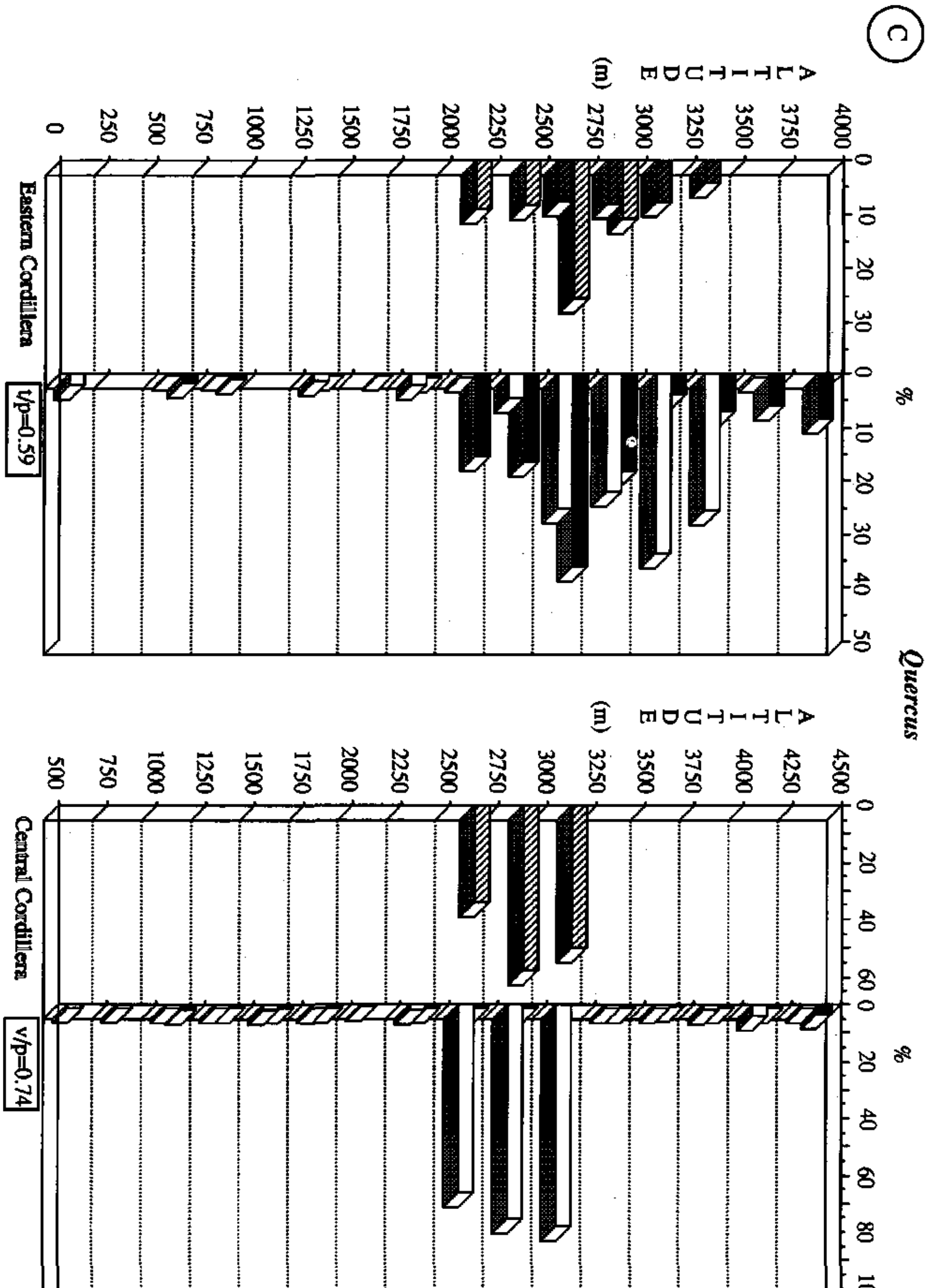
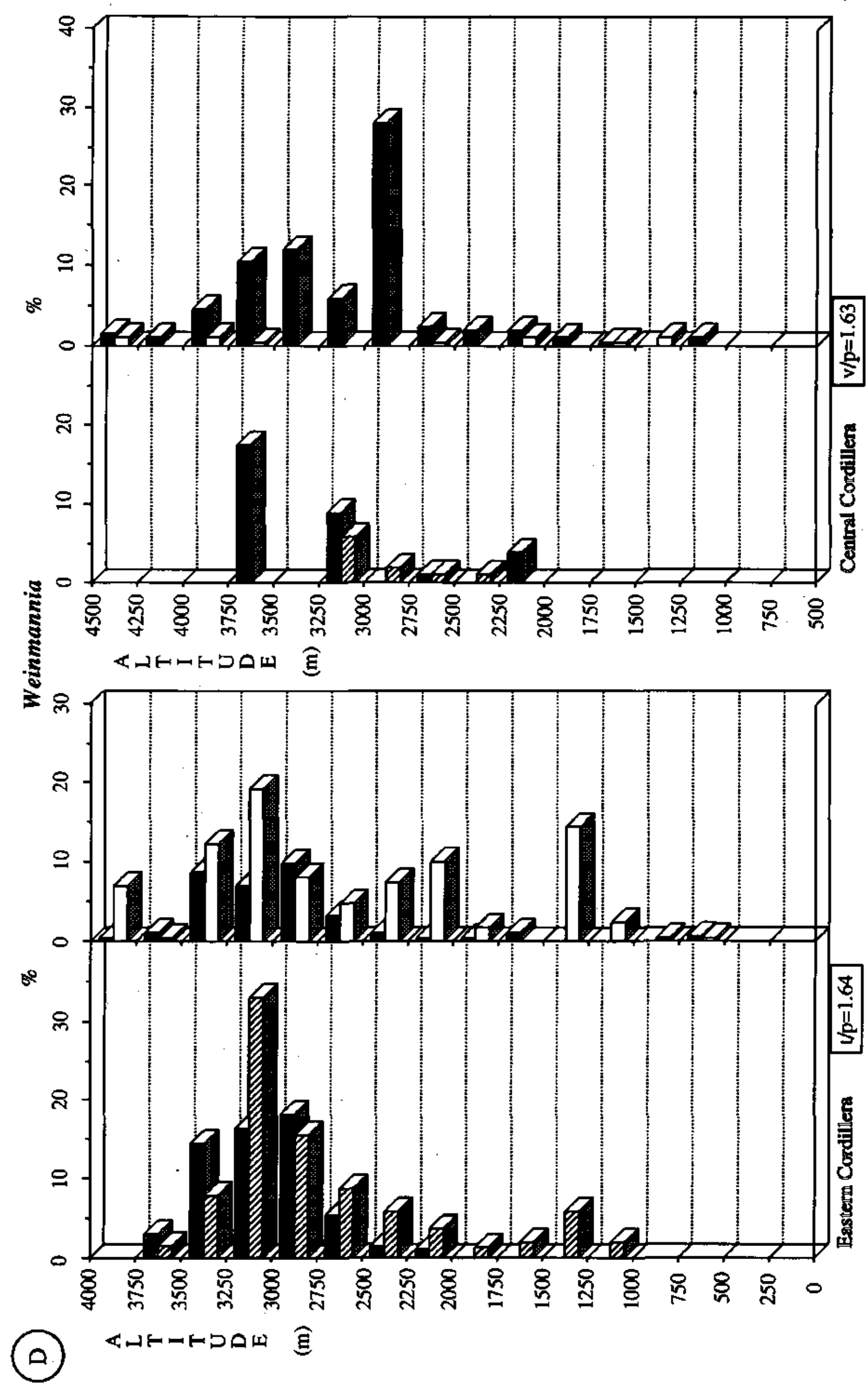


Fig. 32 A to D (see also next pages).
The actual tree cover in the vegetation and pollen deposition (in percentages of the total vegetation cover and pollen sum of each relevé, respectively) of some important taxa. A distinction was made between the western and eastern flank of the Central and Eastern Cordillera (data according Grabandt 1980 and Melief 1985).


B
Hieronima
%
0 10 0 10 20 30
ALTITUDE (m)
4000
3750
3500
3250
3000
2750
2500
2250
2000
1750
1500
1250
1000
750
500
250
0
Eastern Cordillera
t/p=0.75
%
0 10 0 10 20 30
4500
4250
4000
3750
3500
3250
3000
2750
2500
2250
2000
1750
1500
1250
1000
750
500
Central Cordillera
v/p=1.22


C
Quercus
ALTITUDE (m)
%
Eastern Cordillera
v/p=0.59
Central Cordillera
v/p=0.74


D
Weinmannia
%
ALTITUDE (m)
Central Cordillera
v/p=1.63
Eastern Cordillera
t/p=1.64

as well as for vegetation cover. At the drier east-facing slopes a low maximum in the vegetation is reached around 3000 m altitude. The low vegetation cover of *Weinmannia* and the low pollen percentages (< 5%) along the entire eastern flank might be related to human influence on the vegetation which is more intensive at the eastern side than at the western side of the Central Cordillera (Van der Hammen, oral information).

-*Hedyosmum* (Fig. 32E) *Hedyosmum* pollen is present over a wide vertical range (c. 1250- 3500 m altitude) in the actual pollen rain of the Eastern Cordillera. There is a distinct optimum of *Hedyosmum* pollen around 3000 m altitude at the west-facing slopes. This optimum is absent along the east-facing slopes. The presence in the vegetation of *Hedyosmum* is altitudinally somewhat more restricted (c. 1750-3000 m altitude) and the percentages are relatively low. In the Central Cordillera high values for *Hedyosmum* pollen as well as for the vegetation cover are reached at an altitude of about 3500 m at the east-facing slopes and around 2200 m altitude at the west-facing slopes.

-*Myrtaceae* (Fig. 32F) Representatives of the Myrtaceae are more abundant on the west-facing slopes of the Eastern Cordillera than on the east-facing slopes. Notwithstanding the maxima in pollen rain and vegetation percentages between 1000 m and 2000 m altitude a wide-scattered altitudinal distribution of Myrtaceae pollen has been registered. This wide vertical scatter is explained by the fact that the Myrtaceae family includes several genera each with its own specific vertical distribution. Compared with the west-facing slopes, in the Central Cordillera the eastern flanks show somewhat higher percentages of Myrtaceae in the vegetation cover as well as in the pollen contribution. There is an optimum for Myrtaceae pollen as well as for the vegetation cover around c. 2250 m altitude. Relatively high values for pollen as well as for the vegetation cover are reached between 1200 m and 1700 m altitude along the western side of the Central Cordillera.

-*Clusia* (Fig. 32G) This genus occurs at all altitudes. In the Eastern Cordillera high values of recent pollen rain are reached between c. 2500 m and 3500 m altitude along both flanks. In the Central Cordillera the distribution of *Clusia* strongly suggests a preference for the drier east-facing slopes. With increasing altitude the contribution of *Clusia* in the vegetation declines and disappears above an altitude of c. 3000 m without reaching a distinct maximum. The high peak in pollen percentages around 3000 m altitude along the west-facing slope is not provided with a description of the local vegetation (Melief 1985). In view of the high v/p-value it is most probable that *Clusia* grew *in situ* at that particular site which is confirmed by Cleef (pers. comm.).

-*Ilex* (Fig. 32H) In the Eastern Cordillera *Ilex* occurs above 2500 m altitude along both slopes. Pollen as well as scrubs of this genus reach high values around an altitude of c. 3600 m. In the Central Cordillera scrubs of *Ilex* have been registered only on the west-facing slopes from c. 2500 m to c. 3200 m altitude. The pollen percentages show a wide-scattered occurrence but tend to increase between approximately 1700 m and 2100 m altitude. These relatively high pollen percentages indicate that *Ilex* was growing somewhere in the surrounding of the relevé.

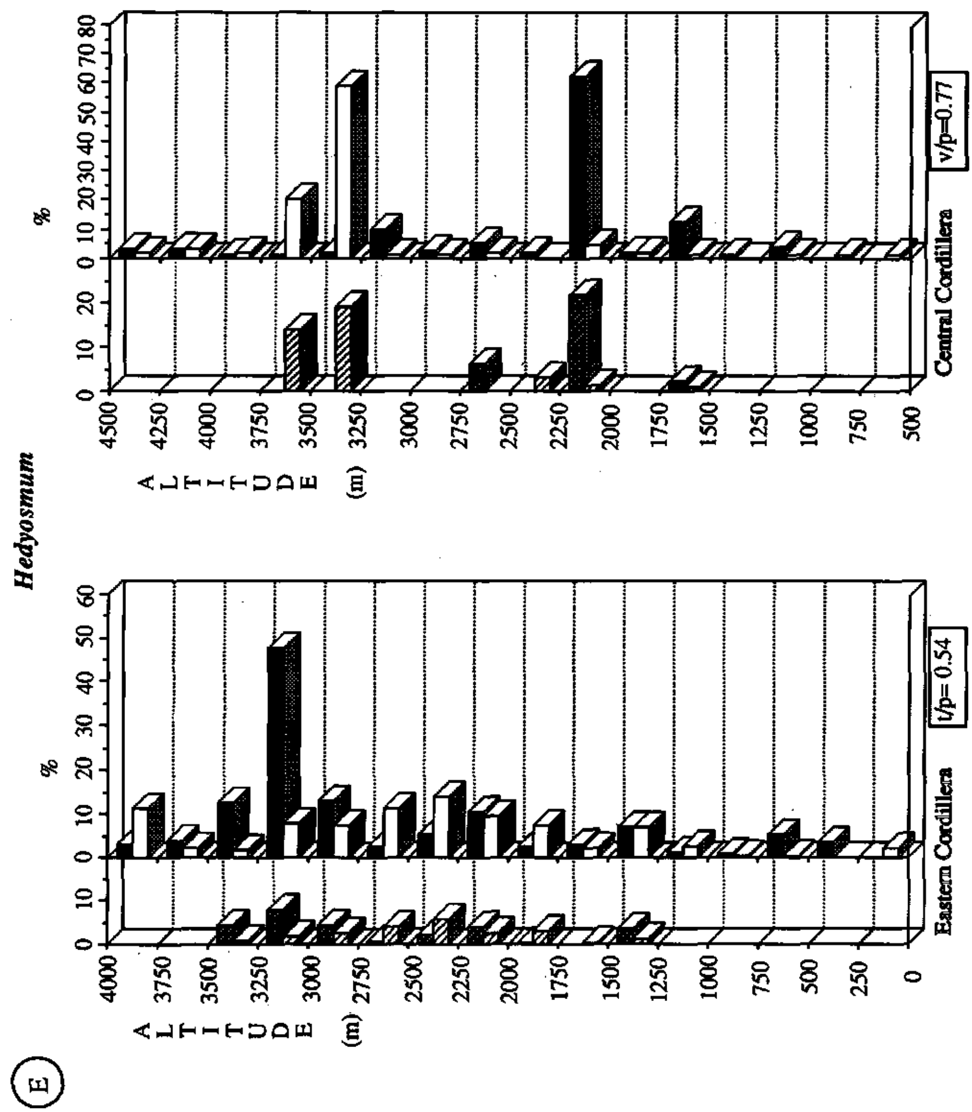
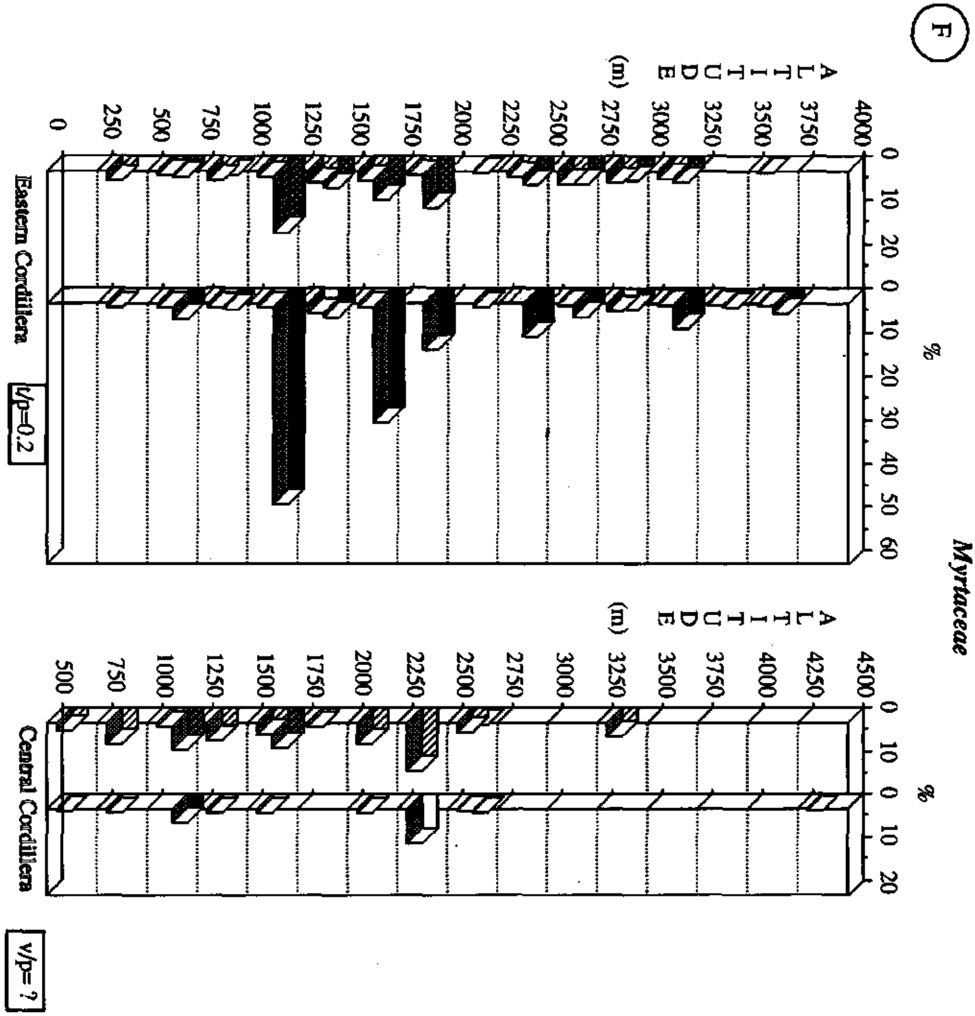
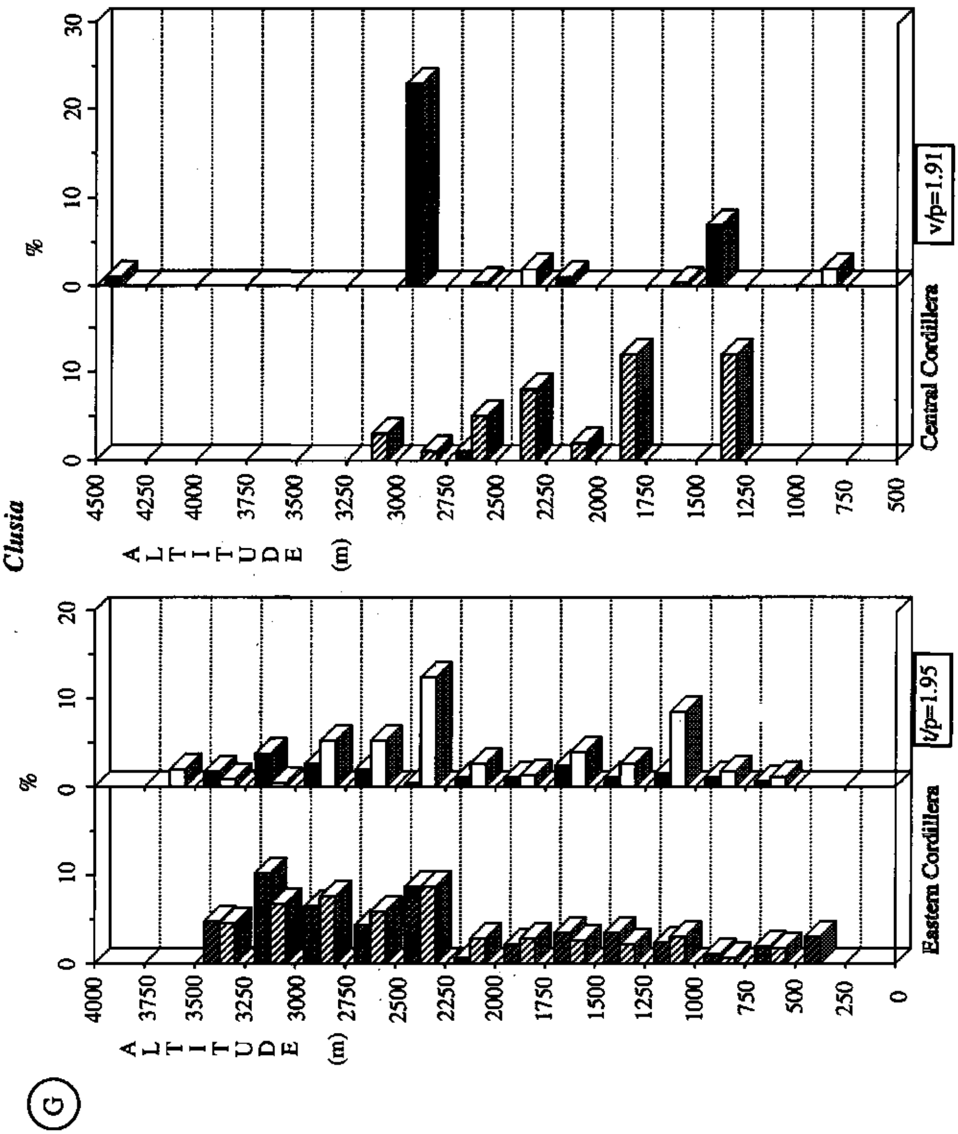
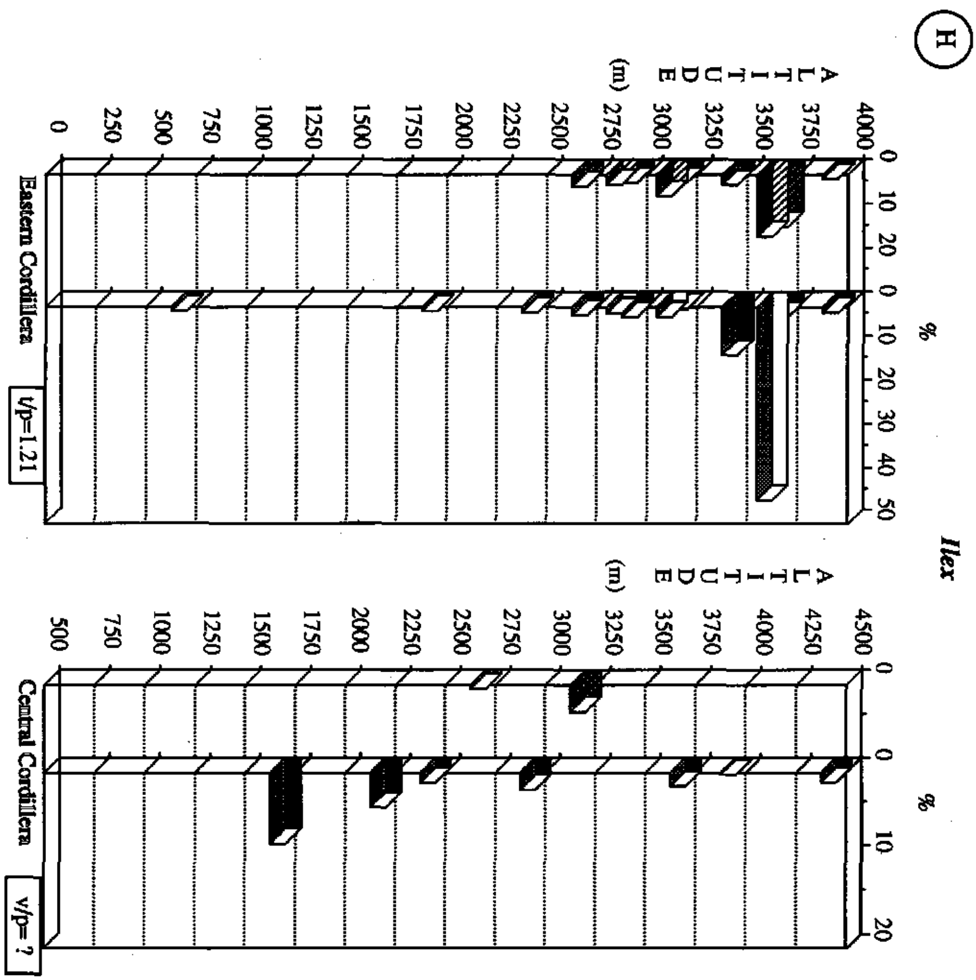


Fig 32 E to H (see also next pages)


F
Myrtaceae
%
ALTITUDE (m)
Eastern Cordillera
t/p=0.2
Central Cordillera
v/p= ?



Clusia
%
v/p=1.91
Central Cordillera
ALTITUDE (m)
4500 4250 4000 3750 3500 3250 3000 2750 2500 2250 2000 1750 1500 1250 1000 750 500
%
t/p=1.95
Eastern Cordillera
ALTITUDE (m)
4000 3750 3500 3250 3000 2750 2500 2250 2000 1750 1500 1250 1000 750 500 250 0
G


H
Ilex
%
ALTITUDE (m)
4000
3750
3500
3250
3000
2750
2500
2250
2000
1750
1500
1250
1000
750
500
250
0
Eastern Cordillera
t/p=1.21
4500
4250
Central Cordillera
v/p= ?

5.2 Pollen analysis

5.2.1 MATERIAL AND METHODS

Several borings were carried out by the author and special attention was paid to the organic-rich sediments in the low-lying areas in the eastern part of the basin. The used drilling equipment is a modification of the 'Streif-sampler': It consists of a movable piston in an open-ended steel cylinder in which an exchangeable plastic tube (length c. 1m, ø7 cm) is placed. At the required depth the piston is hold rigid and the cylinder with the inner plastic tube is pushed further down into the sediment. With the piston at the top of the core the sample is withdrawn. The plastic tube in which the sampled material is collected is replaced by an empty one and the drilling procedure continues. A depth of 13 m could be reached with the help of extension rods. Through measurements in the field it was determined whether the sediments were compacted during the drilling procedure or not. The tubes were transported to the Hugo de Vries-Laboratory, Department of Palynology and Paleo/Actuo-ecology, University of Amsterdam, The Netherlands.

In the laboratory the plastic tubes were sawed open after which the content was carefully described. Borehole 11 and 2 (named PIT 11 and PIT 2 respectively; see Appendix IV for location) were found to be the most appropriate cores for a palynological study: PIT 11 was chosen as one of the most representative boreholes for the eastern flood basins and for its length which totals almost 13 m. Core PIT 2 is important as it represents the Holocene period that is lacking in PIT 11. Subsequently, with an interval of 5 cm the sampled material of PIT 11 was cut in slices each 1 cm thick. However, only the samples at every 15th cm were used in this study[3)]: *e.g.* 600 cm, 615 cm, 630 cm, 645 cm etc. Of these selected samples a known volume was taken for the extraction of the pollen. The sampled sediment of PIT 2 was cut each centimeter and every 10th cm was used for the palynological study. All interjacent samples of both cores which were not used were sealed and stored in plastic bags and are available for further study. For the preparation of the pollen slides the following method was used: First of all the sample volume was determined by means of a pycnometer. The peaty samples were boiled during 10 minutes in a 10% KOH-solution or when very clayey, in a 10% Na-pyrophosphate solution. This last treatment was proposed by Bates *et al.* (1978) in order to remove clay particles. *Lycopodium* tablets each containing a known amount of pollen grains were dissolved in a solution of HCl (5%) and rinsed to remove the sodium carbonate from the tablets after which they were added to the seaved samples. By adding a known number of exotic *Lycopodium* spores to a known sample volume the number of pollen grains per cm^3 can be calculated[4)] for each sample (Stockmarr 1971; Birks & Gordon 1985). Thereafter, the samples were acetolysed (Erdtmann 1952) and

---

3) *The sampling method of the core did not always allow to take a 15 cm-interval as the lowermost 5 centimeters of each plastic tube had to be collected as one sample. These 5 cm-long core samples were thoroughly mixed in the laboratory and considered as representative for the middle part of these core intervals, i.e. the sample from 1280 to 1285 cm depth was mixed and graphed at depth 1282.5 cm in the pollen diagram.*

4) *Only the vegetation elements which are included in the pollen sum (see Table 4) are considered in the calculations as the pollen supply of the local elements are too highly subjected to local changes.*

gravitionally separated by means of a bromoform-alcohol mixture (s.g. of 2.0). Finally, the samples were mounted in glycerin jelly, a method described by Faegri & Iversen (1975). The analysis was generally continued until a minimum of at least 100 arboreal pollen (excl. *Alnus*) had been counted.

5.2.2 CORE PIT 11

Core PIT 11 is situated in the area of El Higueron (see Appendix I), south of Pitalito near the southern basin margin. Although the site itself is well drained due to human interference, the surrounding area is poorly drained. The site is actually used as grass land with local stands of shrubs. For the legend and interpretation of the stratigraphical column of core PIT 11 one is referred to Fig. 27.

5.2.2.1 STRATIGRAPHY

0-40	cm	Clay; brown-grey (dry). Compact with rootlets; gradual to:
40-98	cm	Sandy clay; light brown-grey (dry). Dense character; at the base small lenses made up of hornblende-enriched fine sand (105-150 µm); abrupt transition to:
98-120	cm	Clay; green-grey (dry). Angular blocky structure; locally spaces enriched with gypsum; abrupt transition to:
120-172	cm	Sandy clay to clayey sand; green-grey (moist). Dense character; towards the base an increase of the sand-size component; abrupt transition to:
172-180	cm	Humic clay; light-brown (moist). Abrupt transition to:
180-285	cm	Peat; dark brown to brown (moist). Where stratified the macro-remains show a horizontal orientation. At some intervals the tube was completely filled up with wood (*e.g.* at 270 cm). From 190 cm to 200 cm a small band of hornblende-enriched fine sand (105-150 µm). From 232 cm downward abrupt intercalations of laminae (1 to 5 mm thick) made up of humic clay; abrupt transition to:
285-306	cm	Clayey peat; brown (moist). Horizontal stratification; wood at 298 cm; distinct transition to:
306-325	cm	Humic clay; brown-grey (moist). Intercalated horizontal, organic-rich lamina; gradual transition to:
325-640	cm	Peat; dark brown to brown (moist). Where stratified the macro-remains show a horizontal orientation. Wood from 342 cm to 359 cm; intercalated laminae of humic clay from 390 cm to 490 cm. Compound very weak from 560 to 580 cm; many roots from 570 cm to 575 cm; gradual transition to:
640-696	cm	Peaty clay; brown (moist). Horizontal stratification; gradual transition to:
696-844	cm	Peat; dark brown to brown (moist). Horizontal stratification; wood from 695 cm to 702 cm. From 832 cm downward an increase of clayey laminae; gradual transition to:
844-857,5	cm	Humic clay; brown (moist). Intercalations of horizontal-oriented, organic-rich laminae (c. 0,5 cm thick); distinct transition to:
857,5-897	cm	Peat; dark brown to brown (moist). Many wood fragments intercalated; distinct transition to:
897-1007	cm	Humic clay; brown (moist). Slightly silty at some intervals. Intercalations of organic-rich laminae (c. 1.5 mm thick); gradual

		transition to:
1007-1068	cm	Sandy clay to sand; light brown-grey (moist). This interval shows a fining-upward sequence (top; 105-150 µm, base; 1400-2000 µm). At the base the sand is clast-supported. At 250 cm a 3 cm-thick band of laminated, humic clay with horizontally oriented leaves; abrupt transition to:
1068-1111	cm	Humic clay; brown (moist). Many intercalations of organic-rich laminae (<1 mm thick); distinct transition to:
1111-1150	cm	Peat; dark brown to brown (moist). Amorphous; abrupt transition to:
1150-1170	cm	Humic clay; dark brown to brown (moist). Intercalated macro-remains; gradual transition to:
1170-1274	cm	Peat; dark brown to brown (moist). Horizontal stratification; from 1215 cm dispersed fine sand; at 1230 cm dispersed angular gravels (ø 1 cm); abrupt transition to:
1274-1285	cm	Humic clay; brown (moist). Dispersed fine sand.

5.2.2.2 SOME CHEMICAL PROPERTIES

The absence of pollen at 750 cm depth and from 615 cm to 550 cm depth (Appendix V) is interesting. During these phases a soil possibly developed and pollen might be absent due to corrosion. Soil development may be registered by changes in C-, N- and C/N-values. Plant production depends on the availability of N, which is largely determined by the process of mineralization. During mineralization the organic N present in plant residues, microbial mass and humus is transferred to inorganic N (NH_4^+, NO_3^- and NO_2^-) by microbial activity. N in excess of the microbial demand is released to the inorganic pool which means a rise in the N-concentration. When oxygen becomes depleted the process of denitrification takes place. During this process nitrate is transferred to gaseous N and will escape from the system leading to a loss of nitrogen. According to Martin & Holding (1978) and Hemond (1983) loss of nitrogen due to denitrification is negligible in bog ecosystems. Stanford (1975) however stated that loss by denitrification may reach values up to 15%. Loss by erosion and run off varies considerably from zero to c. 30% (Martin & Holding, 1978). During the process of decomposition there will be a loss of organic C due to the formation of CO_2. With the assumption that the loss of gaseous nitrogen is negligible, the foregoing would lead to the conclusion that on the whole the development of a soil would lead to decreasing C/N-ratios. However, the interpretation derived from the C-, N- and C/N-curves is tentative as the C- and N-content depend on several variables like: temperature, pH, type of vegetation, humidity, topography (*e.g.* Jenny 1941; Hendrickson 1985). To give an example: the concentration of N varies in different plant taxa (Rodin & Bazilevich 1967), in different micro-habitats (Dowding 1981; Middeldorp 1984) and even may vary inside one individual. Consequently, a variation in the floristic composition would also lead to changing C/N-values.

C- and N-contents of 75 samples were established with a sample interval varying between 10 cm and 20 cm. The C-content is measured after digestion of organic carbon by wet combustion (Begheijn 1976). To obtain the N-content, organic matter was destructed conform the Kjeldahl procedure. The different carbon and nitrogen quantities are expressed as weight percentages per sample (Fig. 33). On the whole the C- and N-curve show a negative relation with the minerogene content of the sediments: low values

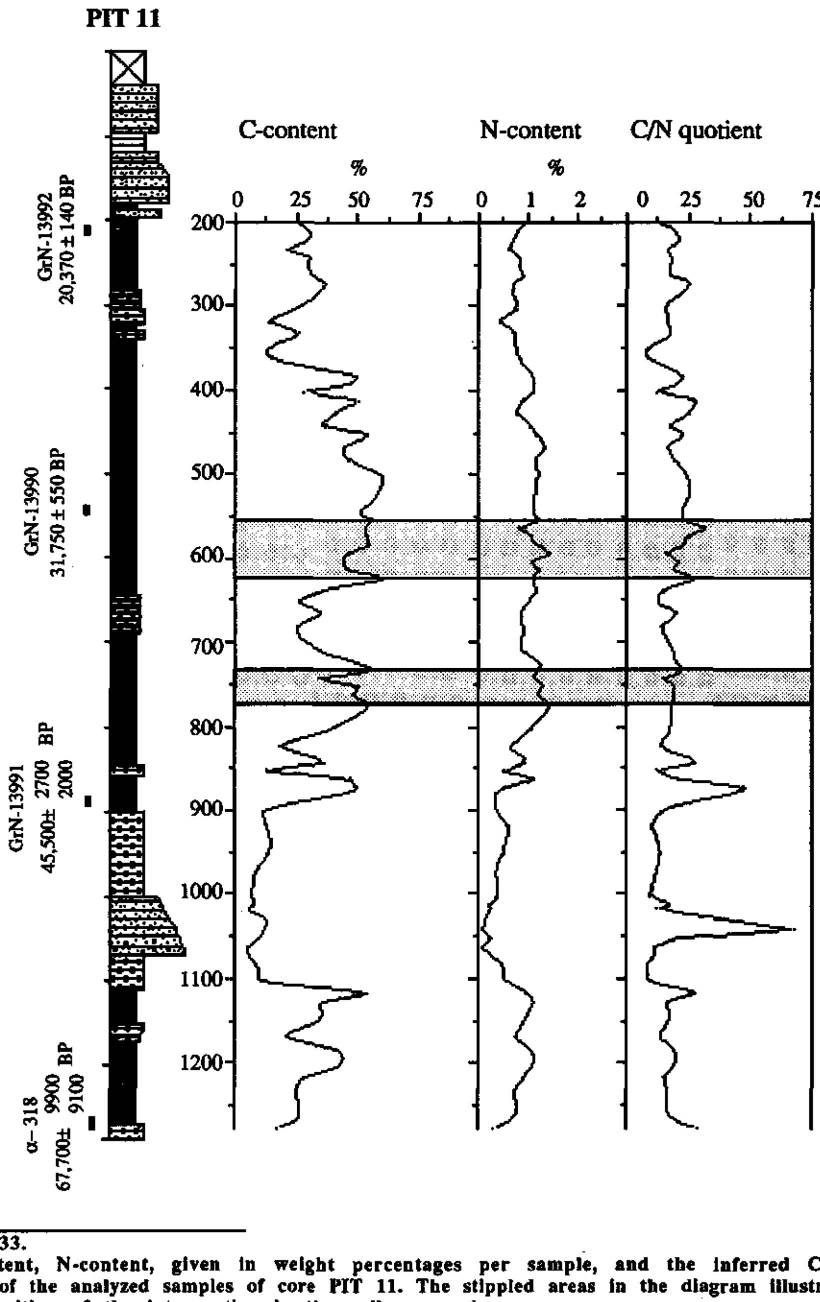


Fig. 33.
C-content, N-content, given in weight percentages per sample, and the inferred C/N-ratio of the analyzed samples of core PIT 11. The stippled areas in the diagram illustrate the position of the interruption in the pollen record.

in clastic-rich material and high values for the organic-rich levels (*e.g.* between c. 1100 and 900 cm, around c. 850 cm, between c. 700 and 650 cm). The C-content around the levels where pollen is lacking (Fig. 33) do not differ significantly from the C-content at other organic-rich levels where pollen is abundantly present. The somewhat decreasing C values at the 750 cm and around the 600 cm level might be consistent with the theory of soil development around these levels but might be ascribed to a somewhat higher minerogenic content of the sediments as well. Moreover, if soil development is involved one might expect stronger fluctuations in the C and N-content due to relatively fast and high decomposition rates in the tropics. It is unlikely therefore that the absence of pollen at the 750 cm level and in the 615-555 cm interval can be ascribed to soil development.

5.2.2.3 DATING

Absolute dating From the peaty parts of PIT 11 a total of five samples was selected and dated at the Centre for Isotopic Research in Groningen. This resulted in four radiocarbon ages and one age based on the U/Th disequilibrium (UTD) dating method (Van der Wijk *et al.* 1986; Van der Wijk 1987). The results are depicted in Table 3. The dating GrN-13781 is probably too young as the type of sediment did not allow a complete treatment of the sample (Mook, pers. commun.). The UTD dating has a large standard error of about 9000 years.

Table 3.
The obtained absolute datings of core PIT 11.

Laboratory numbers	Type of dating	Depth below surface (cm)	Age (years B.P.)
GrN-13992	C-14	206-210	20,370± 140
GrN-13990	C-14	541-545	31,750± 550
GrN-13991	C-14	891-895	$45{,}500\pm^{2700}_{2000}$
α-318	U/Th	1271-1275	$67{,}700\pm^{9900}_{9100}$
GrN-13781	C-14	1280-1285	$41{,}000\pm^{1400}_{1200}$

Regional and local pollen influx Average annual pollen influx numbers of pollen sum elements (=ΣAP) were calculated for some time intervals between 45,500 and 20,370 years B.P. Additionally, pollen influx calculations were also performed for the non-arboreal pollen elements (Σ=NAP) and for *Alnus*. The results are given in Table 4.

The mean annual pollen influx of the pollen sum elements between 45,500 and 20,370 years B.P. is fairly constant which means that over successive long time intervals the average pollen supply was stable. Pollen influx figures of the local elements show a much higher variability, as may be expected; *e.g.* in the period between 45,500 and 31,750 years B.P. the annual pollen deposition of local elements is twice as much as in the succeeding period between 31,750 and 20,370 years B.P.

Table 4.
Pollen influx data (pollen.cm^{-2}.year^{-1}) of core PIT 11.

Time-interval (years B.P.)	Pollen sum elements (=ΣAP-*Alnus*)	Local elements (=ΣNAP)	*Alnus*
45,500/20,370	571	207	184
31,750/20,370	602	130	190
45,500/31,750	544	271	180
37,000/34,500 (base of Zone 11W)	605	376	381
39,000/37,000 (Subzone 11V2)	342	196	27

Relative dating The pollen concentration for each analyzed sample is shown in Fig. 34A. To calculate the total amount of pollen accumulated over the entire sedimentary column, the pollen concentration per sample have been multiplied by a factor corresponding with the sample interval. This factor was chosen on the assumption that the pollen concentration calculated for each sample is representative for the intermediate interval. The results of this operation are presented cumulatively in Figure 34B, starting with a number of zero pollen at the base of the core. If a constant pollen influx is assumed, *viz.* the amount of pollen accumulated per surface unit per time unit (pollen.cm^{-2}.year^{-1}), the changes in pollen concentration are a function of the sedimentation rate. The combination of the obtained absolute datings and the cumulative pollen concentration in

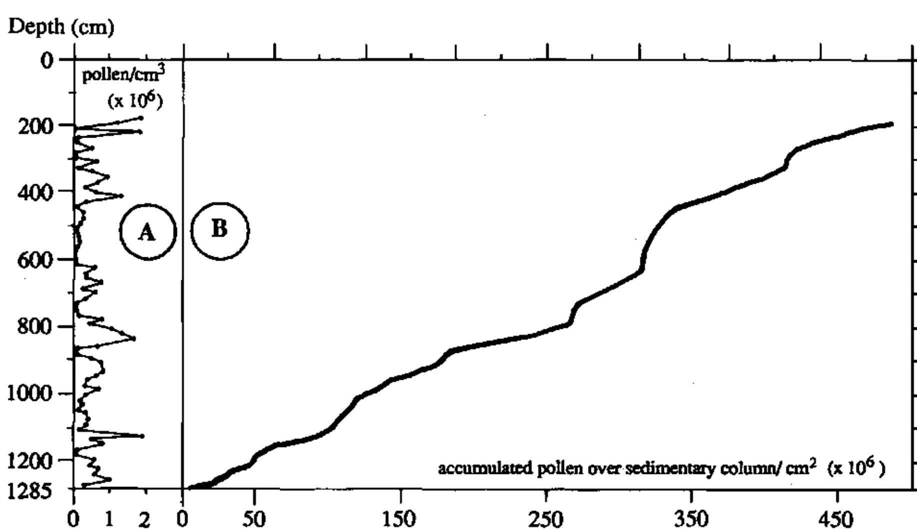


Fig. 34.
(A) Pollen concentration curve and (B) cumulative pollen concentration of pollen sum elements of core PIT 11.

the sediment (Fig. 34B) makes an indirect estimated time calculation possible (Middeldorp 1982, 1984). The excessive pollen production of *Alnus* in combination with its possible occurrence in the local vegetation stand may form a disturbing factor in the calculations. Therefore, the same calculations were performed with and without *Alnus*.

The datings GrN-13781 and α-318 were considered too uncertain, and hence both were excluded from the calculations. The latter has been used merely as a reference to verify those datings which were obtained by extrapolation of the performed calculations. The remaining three ^{14}C datings in combination with the cumulative pollen concentration including *Alnus*, yield the following simple linear regression line (Fig. 35A) given by the equation:

$$Y_i = 676{,}338{,}512 - 11{,}000\, X_i \qquad (r^2 = 0.998) \qquad (1)$$

where the dating for a specific depth i is labelled X and the cumulative pollen concentration at that specific depth is labelled Y. By means of equation (1) an indirect time calculation can be carried out for each level in the section (Middeldorp 1984). As for each level the cumulative pollen concentration is the known parameter we have to perform the calculations with a known Y-value to obtain a dating X for that specific level (see Fig. 35A). These calculations have been performed and the data points are plotted in a scatterdiagram (Fig. 35B). Subsequently, a simple linear regression equation was calculated trough the scatter of points which illustrates the relationship between depth in the section (labelled I) and the calculated age (labelled Y):

$$Y_i = 12{,}000 - 37.21\, I \qquad (r^2 = 0.985) \qquad (2)$$

With the exclusion of *Alnus* from the calculations the results do not change significantly and yield the following equations:

$$Y_i = 479{,}225{,}318 - 8110\, X_i \qquad (r^2 = 0.999) \qquad (1b)$$

and

$$Y_i = 12{,}340 - 35{,}79\, I \qquad (r^2 = 0.999) \qquad (2b)$$

Table 5.
The performed age calculations for some depth levels with and without *Alnus*, respectively.

Level (cm)	Age (incl. *Alnus*) (years B.P.)	Age (excl. *Alnus*) (years B.P.)
180 (top)	17,214	17,029
240	22,384	21,689
405	28,523	27,744
700	37,573	37,744
1110	52,722	50,710
1285 (base)	61,456	59,071

To demonstrate the role of *Alnus* in relative dating techniques, for some levels time calculations were carried out with the help of equations 1 and 1b, respectively. Table 5 displays the relatively minor disturbing role of *Alnus* in the performed calculations; the calculated ages only differ between 200 and 2500 years. Therefore, *Alnus* pollen is included in the time calculations as it also might be an element of regional importance. The base of core PIT 11 has a calculated age of c. 61,500 years B.P. This lies within the error of the UTD-dating (Table 3).

It should be stressed that the above-mentioned approach of provisional dating has a number of underlying assumptions:

-A constant pollen influx over the entire sediment pile;

-No loss of pollen by deterioration;

-No major erosional hiatuses in sedimentation.

The first condition is only met when pollen supply is constant in time. This supply may be influenced by changes in the floristic composition of the forest vegetation, shifting of the vegetation belts and local factors such as sediment focussing processes (Davis 1973; Davis *et al.* 1984). The relative good fit of the linear regression model (1) (r^2= 0.998; Fig. 35A), suggests a constant average pollen influx over at least the interval from 200 cm to 900 cm which is the interval of the used ^{14}C-datings. This is confirmed by the calculated pollen influx numbers over this interval which only show a small variation (Table 4). Several authors stress the importance of low pH-values (< 5.5) for a good pollen preservation (Dimbleby 1957; Berglund 1979). The peat in the Pitalito Basin reveals pH values of 5 (Espinal and Bernal 1966) which apparently is cogent for the second condition. The fact however, that at some intervals pollen is lacking (what might be caused by oxidation), makes this assumption less reliable. In view of the third condition it must be emphasized that the absolute datings which were used in the indirect time calculations are located between 200 cm and 900 cm depth. In this interval peat is the main constituent. From 900 cm to 1111 cm, however, there is a change in lithology which is partly interpreted as crevasse splay deposits. Guion (1984) states that the proximal part of a crevasse may show erosional features in contrast to the more distal part. It is unknown whether the sharp base at 1068 cm embodies an erosional hiatus or not. Most probably we are dealing with distal crevasse deposits which generally do show a sharp but not erosive base.

Equation (2) can be applied to calculate an approximate age for the deeper-positioned sediments at each depth in the basin. By using this equation it is assumed that the sedimentation rates for these sediments are comparable with those calculated for the surficial sediments. This condition however, is beyond control but the geoelectrical survey records a continuous deposition of peat and peat-like material at greater depth in the eastern part of the basin. The proposed lacustrine paleoenvironment (section 3.3; phase 2) in which these sediments were deposited was probably not very susceptible to erosion. When calculations are performed with equation (2) it appears that the base of the sediments at 1200 m depth has an age of c. 4.5 Ma. Equation (2) yields an approximate average sedimentation rate of 0.25-0.30 m/ 1000 years over approximately the last 60,000 years (=calculated age of the base of the core).

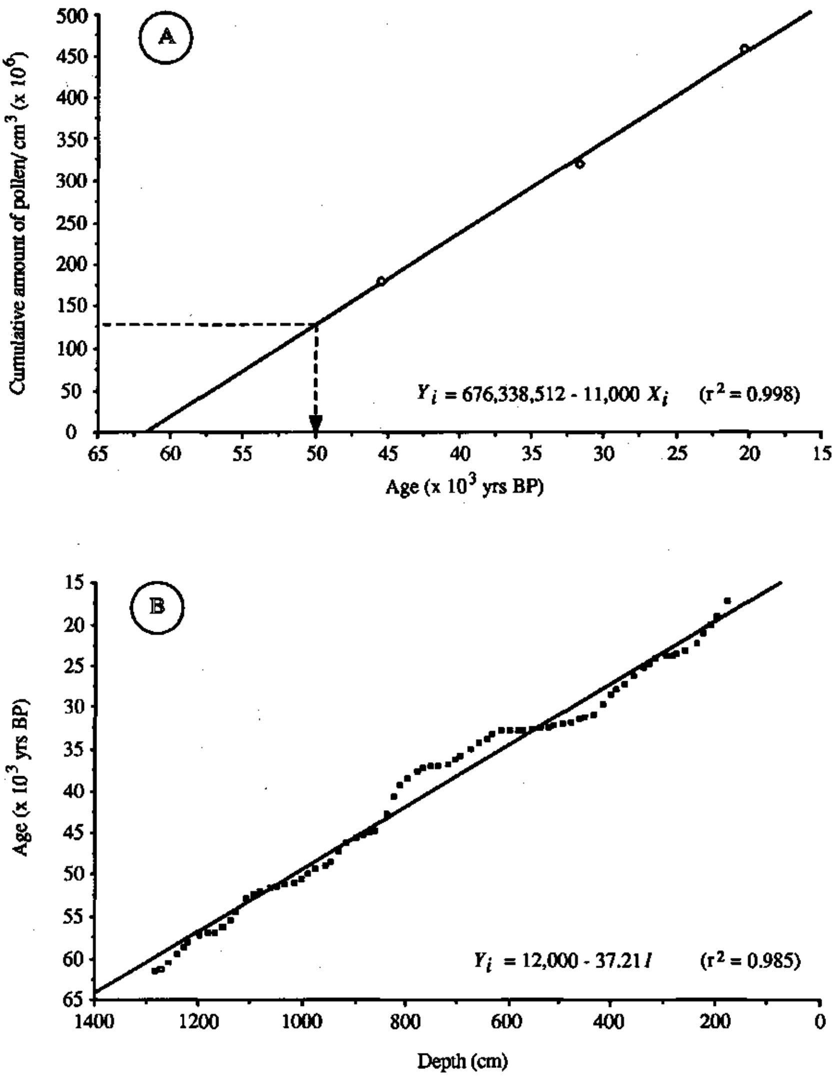


Fig. 35.
A/ The relationship between the obtained ^{14}C datings (X-axis) and the cumulative pollen concentration (Y-axis) of core PIT 11 expressed by a linear regression model.
B/ The relationship between depth (X-axis) and the calculated ages (Y-axis) of core PIT 11 expressed by a linear regression model. The Y-values are obtained by using equation (1).

5.2.2.4 PALYNOLOGICAL ZONATION

The results of the palynological analysis of core PIT 11 are expressed in Appendix V. From left to right the following items can be distinguished:

- the stratigraphical column with depth and the respective ^{14}C datings;
- the local pollen zonation;
- general diagram I (excl. *Alnus*);
- general diagram II (incl. *Alnus*);
- a diagram which illustrates the percentage of arboreal pollen (=ΣAP-*Alnus*) versus the elements of the open type of vegetation (herbaceous elements=ΣNAP);
- separate percentage curves of taxa.

The general diagram I includes those pollen taxa which are of regional importance and characteristic for a zonal type of vegetation. Except for *Alnus* it is assumed that the arboreal elements did not play an important role in the local vegetation types. For that reason, these elements were included in the pollen sum as they give information about regional changes without the influence of local changes. Most of these arboreal elements are not confined to a single vegetation belt but extend over a broader altitudinal zone. Nevertheless, they are mostly restricted in terms of quantitative composition of the plant cover as well as the pollen rain to only one or two of such belts. Taking this into account, four groups of arboreal pollen taxa have been distinguished which represent as clearly as possible the present zonal vegetation. The elements of these groups are given in Table 6.

Table 6.
Taxa, in alphabetical order, used in the pollen sum.

Group I Tropical/Lower sub-Andean elements	Group II Sub-Andean/Andean elements	Group III Andean elements	Group IV High-Andean elements
Acalypha	*Alchornea*	*Alnus**	*Acaena-Polylepis*
Bauhinia	*Hieronima*	*Clusia*	*Clethra*
Cecropia	Myrtaceae	*Cordia*	*Monnina*
Euphorbiaceae	*Phyllanthus*	*Hedyosmum*	Symplocos
Ficus-type	*Quercus*	*Ilex*	
Leguminosae		Meliaceae	
Malpighiaceae		*Miconia*	
Menispermaceae		*Myrica*	
Norantea-type		*Oreopanax*	
Palmae		*Podocarpus*	
Papillionidae		*Rapanea*	
Proteaceae		Solanaceae	
Trema		*Vallea*	
Urticacae/Moraceae		*Viburnum*	
		Weinmannia	

**Alnus* is excluded from the pollen sum in general diagram I but included in general diagram II

Many taxa mentioned in Table 6 have not only their significance as forming part of the regional vegetation types but may play an important role in the local vegetation types as well. *Alnus* for example, is indicative of the Andean forest zone and nowadays is hardly present below c. 2500 m altitude. On the other hand this tree might also be present in ill-drained areas so that the possibility must be considered that *Alnus* formed local stands in the Pitalito Basin. This is confirmed by the findings of tree trunks of *Alnus* in the borehole. Another important reason to remove *Alnus* from the pollen sum is its excessive pollen production. The enormous amount of *Alnus* pollen obscures other, more general tendencies at such levels (Janssen 1959; Hooghiemstra 1984).

The zonation is based on distinct changes in the ratios between the four different groups shown in in the general diagram. Where useful, subzones have been recognized. This subdivision is based on minor changes in the curves of the individual taxa which are included in the pollen sum. The pollenzones are indicated by capital letters and the subzones are indicated by numbers. For the individual percentage curves, pollen sum I, *viz.* excluding *Alnus*, was used.

DESCRIPTION

Zone 11R (1282.5-1205 cm)

Zone 11R is characterized by high percentages of Andean forest elements (c. 60%). The tropical/lower sub-Andean and sub-Andean/Andean group both attain c. 25%. The high-Andean elements show very low percentages (<5%). The arboreal pollen percentages are about 25%.

Alnus completely dominates this zone and reaches c. 80% if included in the pollen sum (general diagram II) and 400% when excluded (see individual percentage curve). Menispermaceae, Palmae type 1 and Urticales are the main contributors of Group I whereas *Quercus* represents Group II with c. 25%. The Andean forest elements are represented by *Clusia, Hedyosmum, Rapanea* and *Viburnum*. Other Andean forest elements as well as the high-Andean elements attain low values (<5%). The Melastomataceae (c. 100%), Compositae, Cyperaceae and Gramineae are well presented. Noteworthy is the presence of *Lysipomia* (c. 6%) and *Plantago* (c. 8%). Except for the monolete spores (mean c. 50%) the presence of other fern spores is low.

Zone 11S (1205-1150 cm)

The tropical/lower sub-Andean group decreases to c. 5% in favour of the Andean forest group which increases to c. 80%. The sub-Andean/Andean group remains unchanged. The arboreal pollen percentages fluctuate sharply with a minimum of c. 15% at the 1170 cm level.

Palmae type 1, Menispermaceae and Urticales decline in favour of *Myrica, Hieronima, Weinmannia* and *Hedyosmum*. *Rapanea* reaches a maximum of 20%. *Quercus* declines to c. 3% at the transition to the succeeding zone 11T. The group of high-Andean elements still shows low values (c. 3%) with *Monnina* as the main contributor. *Alnus* declines to c. 150%. The Melastomataceae drastically decline to c. 45%. The curve of Ericaceae reaches a maximum at the 1185 cm level. The Cyperaceae attain high percentages (mean value c. 200%). *Elaphoglossum* attains 10% and the fungal (asco-)spores inrease.

Zone 11T (1150-1105 cm)

This zone is characterized by a further increase to c. 90% of the Andean forest-group and a decline of Group II to c. 5%. The tropical/lower sub-Andean elements do not change appreciably. The high-Andean elements almost disappear. The arboreal pollen percentages reach high values (c. 50%) around the 1140 cm level.

Clusia and *Weinmannia* increase at the cost of *Quercus* and *Hieronima*. Other important contributors of the Andean forest group are: *Hedyosmum*, *Myrica* and *Rapanea*. The scores of other pollen sum elements attain values below 5%. *Alnus* declines further (c. 60%). Several local elements like Compositae, Cyperaceae, Gramineae and monolete fern spores show an initial decline but reestablish at the end of the zone.

Zone 11U (1105-885 cm)

Zone 11U is characterized by a rapid increase to c. 60% of the sub-Andean/Andean elements followed by a maximum of c. 55% of the tropical/lower sub-Andean group. The increase of these low montane elements coincides with a decline to c. 35% of the Andean forest elements. The high-Andean elements show low scores (generally c. 5%). The curve of the arboreal pollen percentages starts with high values (c. 55%) but decreases to a mean of c. 35% towards the top.

Subzone 11U1 (1105-955 cm)

At the start of the subzone there is a remarkable maximum of *Hieronima* up to c. 70%, followed by a maximum of *Quercus* (c. 50%) and *Alchornea* (c. 5%) while *Hieronima* falls almost back to c. 5%. cf. Menispermaceae and Palmae type 1 rise progressively in the upper half of the subzone. With the exception of *Hedyosmum* the percentages of all the other Andean elements decrease, some to almost zero (*e.g. Rapanea* and *Myrica*). *Alnus* reaches very high percentages (700%) in the upper half of the subzone. Other taxa worth mentioning are: *Antidaphne* , *Lysipomia*, *Eriosorus cf. flexuosus* (isolated peak of c. 100%) and Cyatheaceae. Several herbaceous elements attain maxima around the 970 cm level (Caryophyllaceae, Compositae, Cyperaceae and Gramineae). The same holds for some aquatic elements like *Ludwigia* and *Sagittaria*. The curve of the fungal spore Type 563 (Kuhry 1988) shows a peak of c. 100%.

Subzone 11U2 (955-885 cm)

The tropical/sub-Andean group sums 55% due to rising percentages of Palmae type 1 and cf. Menispermaceae. *Quercus* falls back to c. 25%. *Hedyosmum* and Solanaceae are the main contributors of the Andean forest elements. All other percentages of the Andean elements remain low (<5%). The high-Andean elements sum c. 10% due to the presence of *Clethra*. *Alnus* still attains very high values and obscures the increase of Group I-elements when included in the pollen sum (general diagram II). Polygalaceae, *Ludwigia* and *Sagittaria* are present between other local elements. The curve of the monolete fern spores rises to c. 180%.

Zone 11V (885-770 cm)

This zone is characterized by a decrease of the elements of Group I and II to c. 8% and 12%, respectively. The Andean forest elements rise and sum about 75%. The high-Andean elements progressively decrease to almost zero. The arboreal pollen percentages attain comparable values compared to the previous zone.

Subzone 11V1 (885-790 cm)

Especially the decline of the Palmae type 1 curve causes the decrease of Group I. *Quercus* shows a further decrease to c. 10%. All Andean forest elements increase; first *Myrica*, immediately followed by *Clusia, Weinmannia, Hedyosmum* and *Rapanea*. *Clusia* rapidly decreases to c. 7%. *Alnus* decreases to c. 150%. The curves of *Antidaphne*, Compositae and Gramineae decline whereas Cyperaceae increase. *Lysipomia* is present. *Ludwigia* and *Sagittaria* reduce practically to zero. The curve of the monolete fern spores shows a slight decrease.

Subzone 11V2 (790-770 cm)

Weinmannia increases to a maximum of c. 75% mainly at the expense of *Hedyosmum*. Several other Andean forest elements decrease to almost zero. *Alnus* (25%) and Compositae also decrease. Begoniaceae increase sharply to 25% at the end of the subzone together with the fern spore Type 668 and the fungal spore Type 672.

Zone 11W (735-510 cm)

After the interruption of the pollen record at the 750 cm level the curves of the arboreal elements show strong fluctuations. Notwithstanding these fluctuations an increase is discernable in the percentages of the lower montane forest elements (Group I and II) which totals approximately 40% to 50%. The percentages of the Andean forest elements decrease to a mean of c. 50%. The high-Andean elements attain c. 5%. The arboreal pollen percentages decline to c. 25%.

Subzone 11W1 (735-615 cm)

Quercus rises to a mean of c. 20%. The appearance of Myrtaceae in the upper half of the subzone corresponds with an increase of Urticales. With the exception of the 705 cm level *Weinmannia* attains c. 35%. The absence of *Weinmannia* at the 705 cm level coincides with an isolated peak of 50% of Palmae type 1 and *Alnus* (>1000%). Several curves of the Andean forest elements like *Clusia, Podocarpus* and *Viburnum* show strong fluctuations. *Rapanea* drops to zero. On the whole the contribution of sedge pollen and herbaceous elements to the pollen rain shows an increase: *e.g.* Begoniaceae, Cyperaceae and Gramineae. Other representatives are Compositae, Cruciferae, Ericaceae, *Hydrocotyle, Plantago, Puya, Typha* and pollen Type 659 and Type 662. Worth mentioning are the high maxima of some fern spores at the end of the subzone and the presence of *Anthoceras/Hymenophyllum* and *Gaeumannomyces*. Remarkable is the high peak of the trilete spore Type 668 a the 735 cm level (c. 220%) and the peak of the microfossil Type 676 at the 660 cm level (c. 120%).

Subzone 11W2 (555-510 cm)

After a 60 cm thick layer (615-555 cm) almost devoid of pollen the general diagram I resembles that of the previous subzone. The total sum of Andean forest elements remains the same but there is a remarkable rise of *Weinmannia* to c. 65%. *Rapanea* is present again. From now on *Alnus* attains relatively low percentages. At the transition to zone 11X Begoniaceae reach an absolute maximum with 220%. Compositae, Gramineae, Melastomataceae and monolete fern spores decline drastically. There are maxima of *Paepalanthus, Typha* and *Cheilanthes*. (Asco-)spore types are numerous but irregularly represented.

Zone 11X (510-335 cm)

Characteristic for the zone is the renewed domination of Andean forest elements (c. 75% to 90%). Elements of the lower vegetation zones attain relatively low percentages: both groups approximately 5%-10%. The high-Andean elements also show low scores (<5%). The curve of the arboreal pollen percentages show a progressive increase from 50% at the base of the zone to 70% at the top of the zone.

Main contributors of Group I are Urticales, Leguminosae and Palmae type 1 (all less than 5%). *Quercus* represents the sub-Andean/Andean group with c. 7%. Almost all Andean forest elements show rising values from which the most important are *Clusia Hedyosmum* and *Viburnum*. *Weinmannia* remains with high values (minimum: c. 25%; maximum c. 50%). *Myrica* as well as *Rapanea* progressively decline to almost zero. Begoniaceae, Cyperaceae and Gramineae show a general decline towards the top. Loranthaceae are present. *Typha* and *Paepalanthus* disappear. A number of fern spores and (asco-)spore types disappear in the upper part of the zone.

Zone 11Y (335-205 cm)

The increase of Group I and II (to a mean of c. 18% and c. 25%, respectively) and the reduction of importance of the Andean forest elements (c. 45%) characterize the zone. The high-Andean group attains low percentages (<5%). In general the percentages of the arboreal pollen total remain high (c. 65%).

Subzone 11Y1 (335-285 cm)

The increase of the tropical/lower sub-Andean group is explained by the appearance of Palmae types and Papilionideae. Myrtaceae (c. 20%), *Hieronima* (c. 12%) and *Quercus* (c. 8%) represent the sub-Andean/Andean group. *Weinmannia* declines considerably to c. 5%. *Ilex* increases to a remarkable maximum of c. 25% followed by increasing values of *Clusia* (c. 50%) towards the end of the subzone. Ericaceae, Loranthaceae, *Mühlenbeckia* (no separate percentage curve) and *Elaphoglossum* are present. (Asco-) spore Type 674 reaches a maximum at the transition to subzone 11Y2.

Subzone 11Y2 (285-257.5 cm)

Especially Papillionidae contribute to a further increase of Group I. Myrtaceae and *Quercus* both decline. *Hedyosmum* and *Weinmannia* show a minor increase while *Ilex* declines to c. 5%. *Clusia* remains high. Compositae, Cyperaceae, Ericaceae and Gramineae show a minor increase.

Subzone 11Y3 (225-205 cm)

The interruption in the pollen record is due to the presence of alder wood which completely filled the sampler. Elements of Group I do not exceed 5%. *Quercus* slightly increases to c.15%. The sharp increase of *Hedyosmym* to c. 40% coincides with a drop in the curve of *Clusia*. *Weinmannia* is present with c. 20%. *Alnus* reaches a maximum of approximately 200%. The contribution of the herbaceous elements and the fungal remains is low. Important elements are Compositae, Cyperaceae, Gramineae and Melastomataceae.

Zone 11Z (205-180 cm)

Characteristic is the increase of Andean forest elements to c. 90%. Group I and II decrease to low values (c. 4%). The percentages of the high-Andean elements remain low (less than 5%). The arboreal pollen total increases to c. 70%.

The massive increase of *Myrica* from an initial 10% to c. 60% is characteristic for this zone. It coincides with a sharp decline of *Hedyosmum* and a slight rise of *Clusia*, *Viburnum* and *Rapanea*. All lower montane elements (elements of Group I and II) attain values lower than 5%. *Alnus* decreases to c. 30%. All local elements decrease or disappear.

The upper 180 cm are characterized by high amounts of monolete psilate fern spores and some badly preserved *Myrica* pollen. Other pollen grains, spores or fungal remains are absent.

INTERPRETATION

The base of zone 11R has a calculated age of c. 61,500 years B.P. which lies within the error of the UTD dating (Table 3). The top of the zone has an estimated age of c. 57,500 years B.P. The pollen rain during zone 11R is completely dominated by *Alnus* pollen (general diagram II). It is very likely that alder trees grew on the spot. The findings of alder wood fragments (trunks and branches) in core PIT 11 indeed prove that *Alnus* formed local stands in the ill-drained alluvial plains thereby surpressing the contribution of regional elements in the pollen rain. With the exclusion of *Alnus* from the pollen sum (general diagram I) it appears that *Quercus* and *Hedyosmum* dominated in the regional forest together with *Clusia* and *Rapanea*. An association of *Quercus humboldtii* and *Hedyosmum huilense* is actually present between 1800 and 2600 m altitude near the Merenberg area in the Central Cordillera (Rangel & Lozano 1986; Rangel & Lozano, in prep.). *Rapanea guianensis*, *Rapanea ferruginea* and *Clusia multiflora* are common elements in these forests. The presence of Menispermaceae and Palmae type 4 suggests we are dealing with the lower ranges of this forest type. It must be stressed that nowadays *Clusia* sp. is also found as local element in the La Coneca area. Besides numerous local stands in the alluvial plain itself, *Alnus* possibly grew also on the lowermost part of the slopes of the surrounding hills. Other important local elements were Cyperaceae, Compositae and Polygalaceae. Nowadays *Baccharis nitida* (Compositae) is a dominant species in the swamps of the La Coneca area in the northeastern part of the basin whereas cyperaceous elements like *Cyperus* cf. *haspan*, *Fuirena* sp. and *Eleocharis mutata* occupy more oligotrophic environments (Van Straaten 1989). *Plantago* and *Lysipomia* are elements commonly present around and above the actual forest line (alt. c. 3400 m). Both are considered as local, azonal elements related to the specific edaphic conditions in the alluvial plain: low pH values and ill-drained areas. A comparable combination of Andean forest elements with an azonal páramo-like vegetation is nowadays present at an altitude of c. 2300 m in the El Calendaria bog in the inactive volcanic crater of Merenberg (Rangel & Lozano 1986; Rangel & Lozano, in prep.).

In zone 11S (≈57,500 to 55,500 years B.P.) the variety of taxa increases and several Andean forest elements such as *Weinmannia*, *Myrica*, *Hedyosmum*, *Viburnum* and the fern *Elaphoglossum* appear. In comparison with the previous period it manifests a shift towards an Andean forest type that is found at higher elevations. This is sustained by the disappearance of Palmae type 1 and the presence of *Rapanea*, an element that is abundant in the upper Andean forest belt (cf. *Rapanea dependens*). A mixed *Quercus* forest with *Hedyosmum*, *Weinmannia*, *Myrica* and *Clusia* was present around the basin. *Quercus* forest with *Weinmannia* trees and *Clusia multiflora* are present in the Eastern Cordillera between 2400 and 2800 m altitude (based on unpublished relevés of Van der Hammen, Jaramillo & Murillo). Rangel & Lozano (1986) describe an association of *Weinmannia* and *Myrica* between 2700 and 3100 m altitude and an association of *Quercus* with *Clusia* between 2200 and 2700 m altitude. A combination of these arborescent taxa suggests the presence of a forest type around the Pitalito Basin which is actually present somewhere between 2400 m and 2800 m altitude. The decreasing percentages of *Alnus* indicate a change in the azonal stands of the swamp forest: alder trees grew somewhat further away from the location. The site was covered by a mire vegetation with Begoniaceae (possibly *Begonia patula*), Compositae, Cyperaceae and Gramineae in conjunction with Ericaceae. The absence of aquatic elements like *Botryococcus* or *Isoetes* indicates that open water bodies were generally lacking or dessicated rapidly.

In zone 11T (≈55,500 years B.P. to ≈52,500 years B.P.) the zonal forest is characterized by an increase of trees of *Hedyosmum* and *Weinmannia* combined with *Clusia* and *Rapanea*. *Weinmannia* is actually most abundant between c. 2600 m and 3000 m altitude but may also appear at lower altitude. *Hedyosmum* is common above c. 2000 m altitude but may appear at lower elevations on humid slopes. Except for the decreasing values of Ericaceae the local vegetation is comparable with the previous zone 11S.

Zone 11U ranges from 52,500 years B.P. to 45,000 years B.P. During this period the mixed Andean forest withdrew from the hills and is replaced by a *Quercetum* with co-dominance of *Hieronima*, *Alchornea* and *Hedyosmum*. Rangel and Lozano (in prep.) describe a *Hedyosmo-Quercetum humboldtii* association for the Central Cordillera between approximately 2200 m and 2500 m altitude. *Hieronima* is more common below 2000 m altitude (Fig. 32B) and may be present in the tropical forest around 1000 m altitude. The presence of some elements of Group I like Menispermaceae and Palmae suggests we are dealing with a forest type nowadays present around 2000 m altitude or somewhat lower. *Antidaphne* (Loranthaceae) is an epiphyte which requires much light (Salomons 1986) and is most abundant in the tropical/sub-Andean forest belt. *Eriosorus cf. flexuosus* is a fern registered in *Quercus* forests (Kappelle 1987). It must be stressed that the increase of these elements coincides with a lithological change. During this period the site received sediment and the bog turned into a marshy wetland. Due to crevassing, water flows from the channel into the wetland which may result in the supply of pollen by river water into the wetland. Van der Woude (1981) stated that pollen taxa that occur exclusively in a specific clay layer may be attributed more positively to river supply than taxa that attain comparable or higher values in the peaty beds. From the fact that high percentages of *Hieronima* occur only in the humic clayey layer (from 1105 cm to 1068 cm) it may be concluded that the pollen grains indeed might have been transported by water. The same holds for the increase of *Quercus* which coincides with the base of the crevasse splay at 1068 cm. It is obvious however, that these 'warmer' elements must have grown upstream of the drainage system or in the near surroundings of the basin plain. It seems that the Pitalito Basin was situated somewhere in the upper sub-Andean forest belt and that a *Quercus* forest actually found between 2200 and 2500 m altitude was present at the hills surrounding the basin.

Subzone 11U2 (≈48,000 to ≈45,000 years B.P.) reflects a similar situation as zone 11R: trees of alder grew at the site and obscure the contribution of regional elements in the pollen rain (general diagram II). With the exclusion of *Alnus* from the pollen sum (general diagram I) it appears that elements of the lower tropical/sub-Andean zone replaced the sub-Andean/Andean forest elements. *Quercus* was still present in the surrounding area of the basin but elements of the tropical vegetation belt like Palmae and Menispermaceae increase progressively at the expense of typical Andean forest elements like *Weinmannia*, *Hedyosmum* and Melastomataceae. It indicates a further warming of the climate enabling the approach of tropical/sub-Andean elements. The local vegetation is strongly influenced by the crevasse. The base of the zone coincides with the deposition of sandy material originating from the crevasse. The coarse clastic material is illustrative for an increasing flow regime during which pollen is not deposited. This is also indicated by the low pollen concentration at this level (Fig. 34A). After the deposition of coarse material the crevasse receded and the site was provided with finer clastic material (clay). At that period the site received suspended load during stages of flood. The increase of some aquatic elements like *Ludwigia* and *Sagittaria* suggests the presence of a shallow standing water body. The shore zone of this small pond was occupied by Gramineae, Compositae and Cyperaceae.

In zone 11V (≈45,000 to ≈37,000 years B.P.) the marshy conditions came to an end and made place for a bog environment (see page 78 for definition). Peat growth indicates a more stable condition compared to the previous period and floodwater did not reach the site. From these local changes it may be concluded that pollen was mainly supplied to the bog by air. The type of vegetation in the region is comparable to the one which was present during zone 11T: A typical mixed Andean forest, rich in taxa, returned at the hills. Many taxa like *Clusia*, *Weinmannia*, *Myrica*, *Hedyosmum* and *Rapanea* reappear. *Alnus* withdrew from the place again. Cyperaceae are the main constituent of the local bog vegetation together with ferns. The fungal remains show a slightly increase. *Hedyosmum* is completely replaced by *Weinmannia* in subzone 11V2 (≈39,000 to ≈37,000 years B.P.). It is very likely that the floristic composition of the Andean forest changed in favour of *Weinmannia* and that the hills were covered by a *Weinmannietum* with *Quercus*, *Clusia*, *Rapanea* and *Myrica* as minor contributors. The increase of *Weinmannia* might be ascribed to an increasing humidity as *Weinmannietum* forests prefer the more humid sides of the Central Cordillera (Rangel & Franco 1985). From the low pollen influx of *Alnus* during this subzone (Table 4), it may be concluded that trees of alder grew at greater distance from the site and that the areal extension of swamp forests was reduced.

During zone 11W (≈37,000 to ≈32,000 years B.P.) *Alnus* initially formed a local stand near the site. The high pollen production of *Alnus* (Table 4) obliterates the importance of regional elements (general diagram II). With the exclusion of *Alnus* from the pollen sum (general diagram I) it appears that the contribution of *Quercus* increases, followed by Urticales and Myrtaceae. The high isolated peak of Palmae type 1 coincides with a sharp decline of all other regional elements and a maximum of alder. Possibly an individual of this Palmae type was present near the site and made part of the local swamp forest. A mixed Andean forest surrounded the Pitalito Basin with *Quercus* and *Weinmannia* as dominant taxa. The increase of some sub-Andean and tropical elements suggests that the basin plain was situated around the lower Andean/sub-Andean forest belt (c. 2000 m and 2500 m altitude).

The increase of several local elements in the upper half of subzone 11W1 (*e.g.* Begoniaceae, Caryophyllaceae, Gramineae and some fern spores) is ascribed to the disappearance of alder trees from the site. The groundlayer received more sunlight and herbaceous elements were able to settle around the site. From this time on *Alnus* shows relatively low and less fluctuating percentages which indicate that swamp forests were remote from the site or may not have been present at all. Remarkable is the high pollen influx (Table 4) of several local elements as Begoniaceae, Cyperaceae, Polygalaceae and also some fern and fungal spores at the base of zone 11W. This feature also occurs around the 600 cm level. At both levels pollen of regional importance are absent. A possible explanation for the absence of arboreal pollen around these levels is corrosion of the pollen due to soil development (Havinga 1962; Salomons 1986). Differences in susceptibility to corrosion (*e.g.* Havinga 1963, 1971) may cause the absence of pollen taxa vulnerable to destruction. This automatically results in the overrepresentation of other, less susceptible pollen taxa in the percentage diagrams. It is very unlikely, however, that especially the pollen taxa of regional importance show a higher susceptibility to corrosion. Although changes in C- and N-percentages do not exclude the possibility of soil development (Fig. 33), the supposed increasing humidity during this period seems to be in contradiction with this explanation. Another possibility is a high local pollen production obscuring the contribution of the regional elements to the pollen rain. The high pollen influx of herbaceous elements (=ΣNAP; Table 4) at the base of zone 11W indeed confirms this possibility. The presence of *Gaeumannomyces* is also found by Hooghiemstra (1984) and positively correlated with local *Carex* species (Pals *et al.* 1980). *Plantago* and *Paepalanthus*, elements especially common above the treeline, are found here at the altitude of a mixed Andean forest. Again, we consider them as elements of a local azonal páramo-like vegetation due to the edaphic conditions.

The absence of pollen between 615 cm and 555 cm depth is a phenomenon that needs special attention. There is no change in the sedimentological record: the environment remains a bog and peat was formed. A plausible explanation is the loss of pollen due to corrosion. This would mean a drop of the water table so that oxygen could penetrate temporarily into the soil profile causing increased microbial activity. This is sustained by the intensely rooted horizon around 570 cm depth and by the increase of several fungal remains and not contradicted by the C- and N-content (Fig. 33). The presence, although in bad condition, of *Quercus* pollen in this hiatus is puzzling as this pollen is very susceptible to deterioration in the temperate climatic zones (Havinga 1963, 1984). If the results of Havinga (*l.c.*) are also valid for the tropics, *Quercus* pollen would be one of the first to disappear. An explanation of its presence would be that trees of *Quercus* were abundantly present around the Pitalito Basin producing a high amount of pollen from which just a few survived. The preference of *Quercus* for a relative drier climate (Cleef *et al.* 1983; Rangel & Franco 1985) suggests that around this period (≈33,000 years B.P.) drier conditions prevailed.

After the 'palynological hiatus' (subzone 11W2) *Weinmannia* is significantly more dominant, especially when one takes the high v/p-value of 1.63 into account, and suggests an increasing humidity. It replaces other Andean elements like *Podocarpus*, *Viburnum* and Solanaceae. *Quercus* is present which means that the basin was still situated somewhere in the lower parts of the Andean forest belt. There is a change in the individual contribution of the local elements; Gramineae, Compositae and ferns were replaced by Begoniaceae and aquatic elements like *Typha* and *Hydrocotyle*. Also the presence of *Amphitrema* (*A. flavum*) shows a positive correlation with increasing humidity (Casparie 1972).

During zone 11X (≈32,000 to ≈24,000 years B.P.) a *Weinmannietum* covered the hills with a considerable contribution of other Andean forest elements such as *Clusia*, *Myrica*, *Hedyosmum* and *Rapanea*. A similar association is described by Rangel & Lozano (1986) between 2500 m and 3000 m altitude along the humid slopes of the Central Cordillera nearby the study area. The disappearance of *Quercus* and Urticales indeed suggests a shift of the forest belt to a higher elevation compared to the previous period. The contribution of the herbs decreases towards the end of the zone and some even decline to almost zero. The same holds for several fungal types.

Elements of the lower Andean forest belt increased from ≈24,000 to ≈20,000 years B.P. The sharp increase of Myrtaceae and *Ilex* (subzone 11Y1) seems to be contradictory as *Ilex* is more common in the high Andean forest (Fig. 32H) whereas Myrtaceae are more common below 2000 m altitude (Fig. 32F). However, *Ilex* is found in conjunction with *Eugenia* (Myrtaceae) in the relatively dry *Quercus humboldtii* forests (Lozano & Torres 1984) and with *Ugni myricoides* (Myrtaceae) in other slightly more humid *Quercus* forest types (Kappelle 1987).

Possibly we are dealing with a *Quercus* forest with initially *Hieronima*, *Viburnum*, *Ilex*, Myrtaceae and Palmae type 4. The fern *Elaphoglossum* and the parasitical Loranthaceae are often found in combination with *Quercus* (Kappelle 1987). During subzone 11Y2 (≈22,000 years B.P.) *Ilex* and Palmae were replaced by *Clusia*, Papillionidae and Leguminosae. The increase of Ericaceae might be explained by somewhat drier conditions. A *Hedysomum* community mixed with *Quercus* and *Weinmannia* appears around 20,000 years B.P. (subzone 11Y3). *Clusia* completely disappears. Rangel *et al.* (in prep.) describe the presence of *Hedyosmum* communities between 2500 and 1700 m altitude in the Central Cordillera. Other *Hedyosmum* communities are found at an altitude of 2600-2700 m. The presence of *Weinmannia* and *Quercus* suggests we are dealing with a forest type nowadays present around 2600 m altitude. The pollen rain is dominated by *Alnus* meaning that alder trees reestablished in the local vegetation after a long period of absence. The local presence is confirmed by a large piece of *Alnus* wood at the transition from zone 11Y to zone 11Z which caused an interruption in the pollen record. The intercalations of clayey and sandy laminae indicate that the site came in reach of floodwater.

Around 19,000 years B.P. (zone 11Z) *Myrica* pollen suddenly appeared in relatively high quantities and *Alnus* falls back. Van der Hammen & González (1960a) found alternating high percentages of *Alnus* and *Myrica* at the highplain of Bogotá during the Holocene and explained them to wetter and drier periods, respectively: When *Alnus* dominates, the area must have been subjected to more frequent inundations during the year, whereas the dominance of *Myrica* would point to drier conditions. Therefore, the presence of *Myrica* in zone 11Z may be ascribed to dry climatic conditions around the Pitalito Basin. This is coherent with palynological studies from higher elevations (*e.g.* Van Geel & van der Hammen 1973; Kuhry 1988). They report the presence of a very cold and extremely dry period from c. 22,500 years B.P. to c. 14,000 years B.P.: the Fuquene Stadial.

With the exclusion of *Alnus* and *Myrica* from the pollen sum (not shown) it appears that, compared to the previous phase, especially elements of the higher Andean forest belts increased (*Weinmannia*, *Viburnum* and *Rapanea*) which indicates a lowering of the temperature. Typical sub-Andean and tropical elements like Papillionidae, Palmae, Myrtaceae and *Hieronima* disappear. A *Myrica-Weinmannia* forest is nowadays present around 3000 m altitude. The presence of *Quercus* might be explained by the very dry conditions during this period. Possibly some small patches of *Quercus* forest were present. According the palynological data from the Pitalito Basin the vegetation type had an open character characterized by *Myrica* scrubs and the forest stands, presumably made up by trees of *Quercus* and *Weinmannia*, were scarce.

Around 17,000 years B.P. pollen registration came to an end. The nearly absence of pollen in the upper 180 cm of the core coincides with an inorganic sandy/clayey layer (=marker horizon; page 78). The lack of pollen is explained by loss due to corrosion as the few pollen grains present (mainly of *Myrica*) were in a very bad condition. Apparently, the wetlands were choked with sediments and fell frequently dry causing the formation of gypsum and the cracking of clay. The presence of vertical cracks favoured the penetration of oxygen into the soil profile leading to pollen deterioration. As pollen of *Myrica* is susceptible to corrosion in river clay soil (Havinga 1971) it is assumed that *Myrica* shrubs dominated during this period. It seems that climatic conditions were comparable with the previous period.

The vast amount of monolete fern spores indicates that ferns may have dominated in the local vegetation stand. Similar vegetation stands are actually present in the La Coneca area at sites where the peat layer is thin or absent (Van Straaten 1989). As fern spores are less susceptible to corrosion then pollen (Havinga 1984) they might be overrepresented in the pollen assemblage.

5.2.3 CORE PIT 2

Core PIT 2 is situated in the swampy area of La Coneca in the northwestern part of the basin (Photos 2 & 10). The La Coneca region is the only part of the basin where peat of Holocene age has been found. Notwithstanding the presence of several artificial channels the site is badly drained. The site is used as grassland and in the near surrounding there is cultivation of bananas. For a description of the local vegetation for the La Coneca area one is referred to Van Straaten (1989) and to section 2.4.2. For the legend and interpretation of the stratigraphical column of core PIT 2 one is referred to Fig. 28. This section was analyzed palynologically by L.F. Herrera.

5.2.3.1 STRATIGRAPHY

The total length of core PIT 2 measures 1100 cm (Fig. 28). Except for the upper 445 cm the lithological sequence is comparable with core PIT 11 (Fig. 27): a thick accumulation of peaty material with intercalations of clayey and sandy layers. A differentation between both cores occurs at the 445 cm level where a 110 cm thick compact inorganic sandy/clayey layer is present. From the obtained ^{14}C datings it can be concluded that this layer was deposited between c. 19,000 and 7,000 years B.P. In several locations this very conspicuous layer is present at the surface. The fact that in the La Coneca area this layer is found at depth is ascribed to tilting of the basin floor towards the northern basin margin (section 4.2). This makes La Coneca an unique area as only here Holocene organic-rich deposits are present.

It can be concluded that from 335 cm downwards the same comparable types of sediment are present as were described from PIT 11, starting with the inorganic compact clayey layer. From both sites it appeared that in this marker horizon pollen is absent and that monolete psilate fern spores are abundantly present. For practical reasons it was decided to analyze only the upper 335 cm of core PIT 2 which is considered to be complementary to the palynological data revealed by core PIT 11.

0-105	cm	Peat; dark brown to brown (wet). Densely rooted from 0 cm to 30 cm. From 80 cm many plant rests; abrupt transition to:
105-135	cm	Humic clay; dark grey (wet). At the base plant rests; abrupt transition to:
135-215	cm	Peat; dark brown to brown (wet). Horizontal stratification; rootlets at the top. From 175 cm an increase of plant rests; abrupt transition to:
215-220	cm	Clayey peat; dark brown (wet). Abrupt transition to:
220-275	cm	Clay; dark grey (moist). From 262 to 275 cm small black nodules (ø 1 cm) are present (organo-mineral complexes ?); abrupt transition to:.
275-290	cm	Peat; dark brown (wet). Small black nodules are incorporated. Compound very weak and watery; abrupt transition to:
290-335	cm	Peaty clay; green/brown (moist). From 315 cm to 335 cm an increase of the sand-size with at the base a thin layer (2 cm) of angular gravel (ø 1 cm).
335-405	cm	Clay; green-grey (dry). Angular blocky structure; very compact character. From 367 to 375 cm a layer of humic clay. At the base a 5 cm thick hornblend and biotite-enriched sandy layer is present.

5.2.3.2 DATING

Absolute dating From the analyzed interval only one sample was selected for ^{14}C dating. The surface (0 cm level) is assumed to have an age of 0 years (Table 7). These two datings in combination with the cumulative pollen concentration renders it possible to calculate ages for each level.

Table 7.
The obtained absolute datings of core PIT 2.

Laboratory numbers	Type of dating	Depth below surface (cm)	Age (years B.P.)
—	—	0	0
GrN-13993	C-14	306-310	7150 ± 100
GrN-15234	C-14	461-475	19,370 ± 190

Relative dating To date core PIT 2, the same methodology was applied as has been described for core PIT 11. For the upper 335 cm the pollen concentration at each analyzed level and the total estimated amount of accumulated pollen grains were calculated. For the latter calculation, a multiplication factor was used corresponding with the sample interval. The results are depicted in Fig. 36.

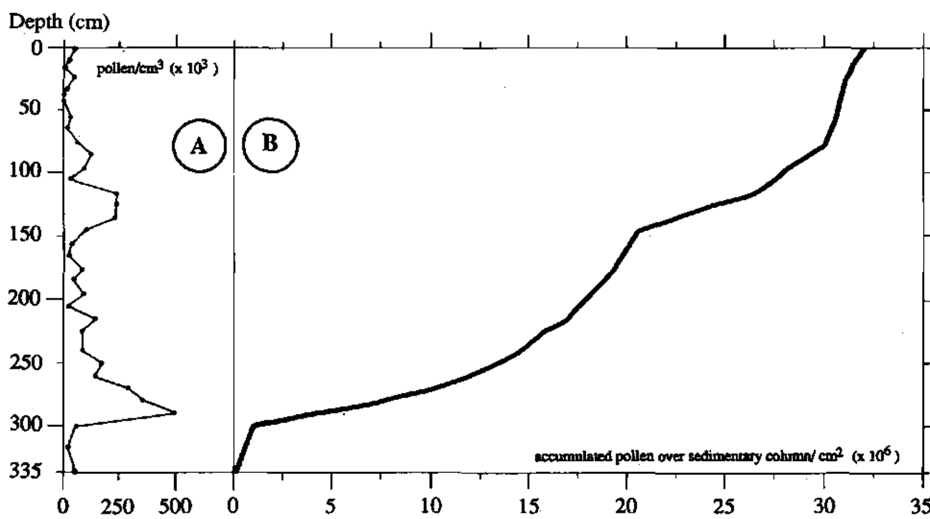


Fig. 36.
(A) Pollen concentration curve and (B) cumulative pollen concentration of pollen sum elements of core PIT 2.

The relation between these absolute datings and the cumulative pollen concentration is described by a simple linear regression line (Fig. 37A) given by the equation:

$$Y_i = 31{,}905{,}929 - 4321\, X_i \qquad (3)$$

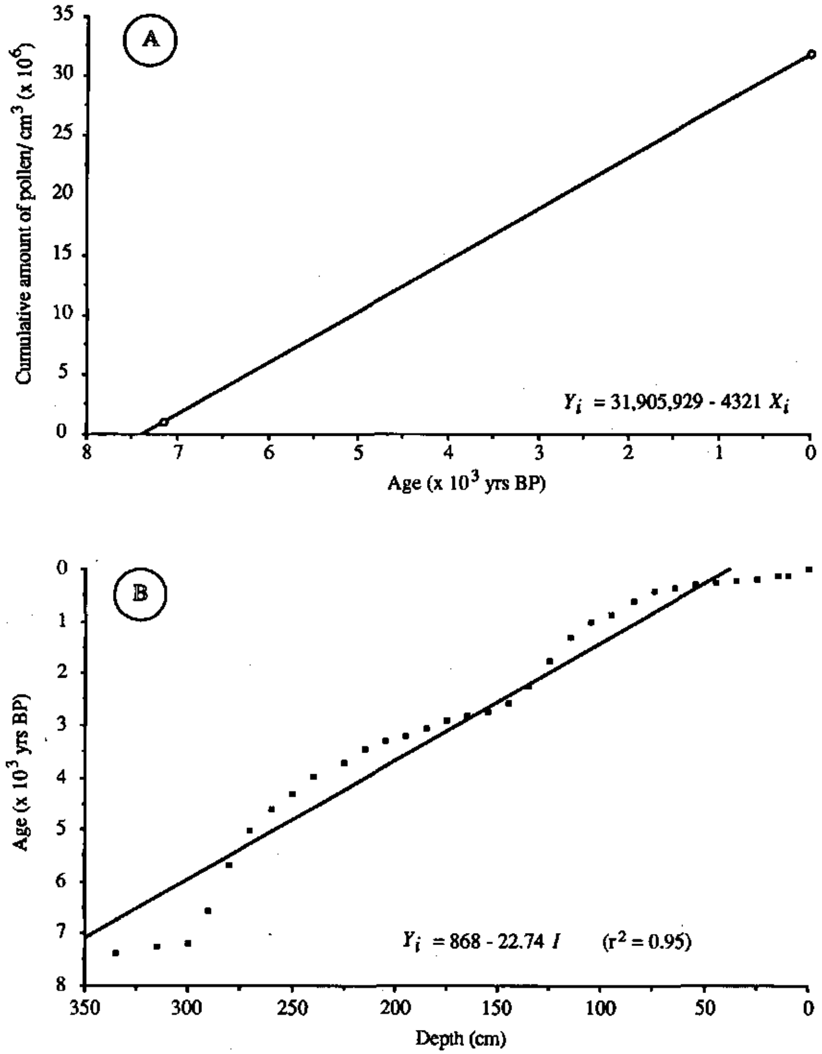


Fig. 37.

A/ The relationship between the obtained ^{14}C datings (X-axis) and the cumulative pollen concentration (Y-axis) of core PIT 2 expressed by a linear regression model.

B/ The relationship between depth (X-axis) and the calculated ages (Y-axis) of core PIT 2 expressed by a linear regression model. The Y-values are obtained by using equation (3).

The estimated calculated age for each specific level is illustrated in Fig. 37B. The relation between these ages and the depth is given by the following equation which was obtained by a simple linear regression:

$$Y_i = 868 - 22.74\, I \qquad (4)$$

Equation (4) yields an approximate sedimentation rate of c. 0.45 m/ 1000 years for the Holocene deposits.

5.2.3.3 PALYNOLOGICAL ZONATION

The same type of palynological diagrams and curves were constructed as earlier have been described for core PIT 11 (Appendix V). These diagrams are based on the same grouping of arboreal taxa as described in Table 6 with the assumption that arboreal pollen taxa did not play a role in the local vegetation stands.

DESCRIPTION

Zone 2X (335-307.5 cm)

This two-spectral zone is characterized by high percentages of the Andean forest elements (c. 85%) and the presence of *Alnus* (c. 65%). The other groups show low scores (less than 5%). The arboreal pollen percentages are low (40%) and decline further to c. 15% at the end of the zone.

Hedyosmum (c. 80%) is the absolute dominating element in the pollen sum. The curve of *Quercus* shows a progressive increase towards the end of the zone. Melastomataceae are present with high percentages (mean of c. 45%). Several local elements such as Caryophyllaceae, Gramineae and *Polygonum* appear in the upper part of the zone together with the aquatic element *Typha*. This appearance coincides with the increase of Gramineae (45% to 500%) and Labiatae and the decline of Compositae, Melastomataceae and the treeferns (*Alsophila* and Cyatheaceae). The percentages of the fungal remains are low.

Zone 2Y (307.5-110 cm)

The tropical/lower sub-Andean elements show a sharp increase to c. 40%. The Andean forest elements decrease to a mean of c. 40%. Goup IV increases to a mean of c. 10%. On the whole the presence of arboreal pollen is low (minimum c. 10%; maximum c. 30%).

Subzone 2Y1 (307.5-245 cm)

The sharp decline of *Hedyosmum* coincides with the appearance of numerous other pollen taxa especially those of Group I and II. The most important are: *Acalypha*, *Cecropia*, Proteaceae and Urticaceae/ Moraceae for Group I. The curve of *Quercus* reaches a maximum. The Andean forest elements are presented by *Rapanea* and Solanaceae. The presence of *Clethra* explains the rise of Group IV. Melastomataceae and the local elements Caryophyllaceae, Gramineae and *Typha* decrease whereas other local elements such as Chenopodiaceae, *Grammitis* and *Paepalanthus* reach maxima in the upper half of the zone together with the aquatic elements Oenotheraceae and

Botryococcus. The curve of the monolete fern spores shows a sharp increase. Several elements reach a maximum at the 280 cm level. *Hypericum* and *Dalea* appear in the upper part of the subzone. *Zea mais* appears for the first time at the 250 cm level. *Spirogyra* appears and shows a progressive increase.

Subzone 2Y2 (245-165 cm)

There is a slightly increase of several Andean forest elements such as *Clusia*, *Hedyosmum*, Meliaceae, *Vallea* and *Weinmannia*. The curves of the tropical and sub-Andean elements show a general decline. *Hedyosmum* decreases in the upper half of the zone in favour of *Quercus* and *Clethra* (c. 12%). *Acacia* appears with c. 15%. Begoniaceae and Cyperaceae increase in the upper part of the zone whereas *Polygonum*, *Typha* and some fern spores decline (a.o. *Elaphoglossum* and *Grammitis*). Other important local elements are *Hydrocotyle*, Labiatae and *Rumex*. *Botryococcus* disappears towards the end of the subzone. Several representatives of the Zygnemataceae (*Debarya*, *Spirogyra* and *Mougeotia*) attain maxima.

Subzone 2Y3 (165-110 cm)

Cecropia decreases and the curve of Urticaceae/Moraceae increases somewhat. *Miconia* and *Vallea* reach their maxima at the expense of *Hedyosmum*. Melastomataceae attain high values (c. 60%). The decline of Begoniaceae at the end of the zone coincides with an increase of Gramineae and Labiatae. Compositae are well presented with c. 60%. *Zea mais* is not recorded. Representatives of the Zygnemataceae show a peak around the 140 cm level together with a maximum of the fungi *Neurospora*.

Zone 2Z (110-0 cm)

A sharp increase of the tropical/lower sub-Andean elements (c. 50%) at the expense of the Andean forest elements (c. 35%) characterizes this zone. The elements of Group IV increase to c. 10%. Initially, the arboreal pollen percentages are low (c. 10%) but attain relatively high values in the upper part of the zone (50%).

Subzone 2Z1 (110-75 cm)

The main contributors to Group I are Urticaceae/Moraceae with c. 30%. Other tropical and sub-Andean elements are *Acalypha*, *Cecropia*, *Trema* and Palmae. The curve of *Quercus* shows a progressive decline whereas *Hierinoma*, *Alchornea* and Myrtaceae increase. *Hedyosmum* rises at the cost of *Miconia* and *Vallea*. *Acaena*/*Polylepis* is present with c. 10%. Numerous herbaceous elements increase or reappear a.o.: *Acacia*, *Piper*, Caryophyllaceae, Cyperaceae and Gramineae. Cyatheaceae, *Elaphoglossum* and monolete fern spores are the main representatives of the ferns. *Pediastrum* reaches a maximum of 20% at the transition to subzone 2Z2. *Zea mais* reappears with c. 4%.

Subzone 2Z2 (75-0 cm)

Acalypha, *Cecropia* and *Trema* increase whereas *Quercus*, *Hieronima* and Myrtaceae decrease to zero. *Miconia*, Solanaceae and *Vallea* are the main contributors to the Andean forest elements. *Weinmannia* as well as *Rapanea* show a progressively increase. Gramineae and Cyperaceae decline sharply. Compositae, *Pilea* and *Piper* are the main representatives for the local elements. *Acacia* (c. 35%) and *Eleocharis* (c. 40%) show a peak around the 25 cm level. Aquatic elements such as *Hydrocotyle*, *Typha* and the algae *Debarya* and *Spirogyra* are present. Cultivates as *Zea mais*, *Manihot esculenta* and *Psidium* are present.

INTERPRETATION

The pollen record starts at the 335 cm level, after the deposition of an inorganic sandy/clayey layer which was deposited during the dry and relatively cold Fuquene stadial (zone 11Z).

Zone 2X starts approximately at 7350 years B.P. and ends around 7200 years B.P. The forests around the basin were dominated by *Hedyosmum* trees in combination with *Quercus*, Proteaceae and possibly Melastomataceae. *Hedyosmum* grows both in the Andean and sub-Andean forest and is even present in the warm tropical zone (*H. scaberrimum;* Rangel *et al.*, in prep.). It is also an indicator for the instability of the vegetation. Rangel *et. al.* (*l.c.*) report the presence of *Hedyosmum* trees especially between c. 1700 m and 2500 m altitude. The absence of *Weinmannia* and the presence of tropical elements suggest that we are dealing with a sub-Andean forest with some tropical elements. *Alnus* was still present nearby the site. *Alsophila* and other Cyatheaceae are important constituents in the forest. The local vegetation was made up by Caryophyllaceae, Gramineae, Labiatae and *Polygonum*. They grew on the peaty soils with a possibility to anchor their roots in the underlying compact clay. Some present-day examples are *Hyptis* sp. (Labiatae), *Polygonum punctatum* and some grasses like *Panicum helobtum* and *Leersia hexandra* which are found along the outer zones of the swamps where their roots lie in the reach of underlying clayey deposits (Van Straaten 1989). Melastomataceae may also have been present as local element. Van Straaten (*l.c.*) reports a high coverage of *Tibouchina anderssonii* in the actual swamps of the La Coneca area. Salomons (1986) describes the possibility of some caryophyllaceous taxa to grow in the understory of an Andean forest or on bare slopes or peaty soils (*e.g. Cerastium* sp. and *Stellaria cuspidata*). The presence of *Typha* indicates local wet conditions.

Alnus dissappears around 7200 years B.P. from the surroundings of the Pitalito Basin together with *Hedyosmum*. A similar fall of *Hedyosmum* and *Alnus* is known from the section Laguna de Palacio at 2000 m altitude (Van der Hammen 1974; Kuhry 1988). This fall however, is dated around 11,000 years B.P. A fall of *Alnus* is also registered in the Merenberg area (alt. 2500 m) around 6500 years B.P. (Piñeros 1988) indicating that this change may not merely be a local phenomenon. The sharp decline of *Hedyosmum* and the simultaneous rise of numerous tropical/sub-Andean elements like *Acalypha*, *Cecropia*, Urticaceae/Moraceae and Proteaceae reflects an amelioration of the climate (subzone 2Y1; ≈7200 to ≈4100 years B.P.). These elements are most common in the lower sub-Andean forest belt between c. 1200 and 1800 m altitude. An *Inga*-*Cecropia* forest with *Quercus* is found along the eastern slopes of the Magdalena Valley between 2000 m and 2300 m altitude. However, *Quercus* stands are also reported from lower altitudinal levels (Lozano & Torres 1974). A mixed forest with tropical/lower sub-Andean elements was possibly present around the Pitalito Basin. *Rapanea guianensis*, *Rapanea ferruginea* and also the tree *Clethra* are common elements in the *Quercus* forests. In this context therefore, it is assumed that *Rapanea* as well as *Clethra* are representatives of the sub-Andean forest zone. This also explains the relative high percentages of both pollen taxa in the actual pollen rain (zone 2Z). There is a substantial change in the local peatbog vegetation: all elements of the previous period (zone 2X) show a decline. The local vegetation is characterized by a cyperaceous reed swamp, with stands of Begoniaceae (*e.g. Begonia patula* ?), Labiatae, *Piper*, Chenopodiaceae, Gramineae and ferns like *Grammitis*. The isolated peak of Cyperaceae, Begoniaceae and *Piper* at the 280 cm level might have to do something with the watery peat layer (275-290 cm). *Piper* sp. is nowadays found at the rim of wet places (Van Straaten, in prep.). Remarkable is the presence of elements nowadays growing around the forest limit such as *Hypericum* and *Paepalanthus*. It is not very likely that the pollen and spores of these 'upper-belt' taxa were deposited after long-distance dispersal by water as the clayey material in which they are found probably originates from the hills directly north of the Pitalito Basin (page 81). Possibly an azonal páramo-like vegetation established in La Coneca due to edaphic conditions. The presence of *Selaginella* and Oenotheraceae (probably *Ludwigia*), in combination with open water elements like the algae *Botryococcus*, *Pediastrum* and *Spirogyra*, indicate the presence of standing water. Of importance is also the first occurrence of *Zea mais* pollen at the 250 cm level. This level has a calculated age of c. 4300 years B.P. which means that the first records of agricultural activity of prehistoric man in the Pitalito Basin antedates the known prehistoric settlements by about 1500 years (section 2.5).

From ≈4100 to ≈2800 years B.P. (subzone 2Y2) the hills were still forested by a mixed forest with tropical as well as sub-Andean elements: *Cecropia*, Palmae, *Alchornea*, *Quercus*, *Hedyosmum*, Meliaceae, Solanaceae, *Vallea*, *Clethra* and *Rapanea*. Compared with the previous period, the floristic composition of the forest changes in favour of *Hedyosmum* and Meliaceae. A maximum of *Botryococcus* indicates that stagnant water was still present. So does the presence of fossil spores of *Mougeotia* sp., *Spirogyra* and *Debarya*, all Zygnemataceae. The Zygnemataceae produce spores in stagnant, shallow, mesotrophic fresh water which warms up quickly (Van Geel 1978). The pool was surrounded by elements from the hygrosere: Gramineae (possibly *Panicum helobtum*, *Erianthus trinii*, *Coelorachis aurita*) and Labiatae (*Hyptis* sp.?) were abundantly present in combination with *Polygonum*, *Piper* and some ferns (Cyatheaceae, *Elaphoglossum*). Other elements of the hygrosere are *Hydrocotyle* and *Rumex*. The presence of some of these elements may be ascribed to the underlying clayey layer in which these plants could anchor their roots. As in zone 2X, *Typha* is found immediately at the beginning of peat development after the deposition of clayey material. Besides the records of *Zea mais* pollen, pollen grains of *Phaseolus* spp. (beans) are found at the 215 cm level (c. 3500 years B.P.). The appearance of *Acacia*-type around 3300 years B.P. is remarkable as most species of this genus were introduced by man for economical reasons after the Spanish conquest c. 1,550 years A.D. (Pérez 1978; Torres 1983). A possible explanation is that an autochtonous species inhabited the site.

During subzone 2Y3 (≈2800 to ≈1100 years B.P.) sub-Andean/Andean forest elements increase. The most important elements are *Quercus*, *Miconia*, *Vallea* and Solanaceae. The somewhat further decreasing percentages of tropical elements suggest slightly lower temperatures and a mixed sub-Andean/lower Andean forest occurs on the slopes of the hills. *Vallea* is nowadays limited to the Andean forest belt between approximately 2300 m and 3800 m altitude and prominent at the clay soils of the highplain of Bogotá (alt. c.

2600 m) in the edaphically bound *Xylosma-Duranta-Vallea* forests (*e.g.* Van der Hammen & González 1960). Similar edaphic conditions might have prevailed in the Pitalito Basin leading to the local presence of *Vallea*. Hooghiemstra (1984) reports the presence of *Miconia-Vallea* forests in the early (Pliocene) Andean forest types at an altitude ranging between c. 2000 m and 2500 m altitude just below the forest line. Human influence also may have played a role as *Vallea* might be dominant in degraded vegetation types (Hooghiemstra *l.c.*). In the local vegetation stand Cyperaceae are replaced by Compositae. Begoniaceae and indicators of open water show a sharp decline halfway subzone 2Y3 whereas Gramineae increase. It might be indicative for the desiccation of the pool. Melastomataceae other than *Miconia*, might have been present in the local stand as well as in the forests surrounding the basin.

The last 1100 years (zone 2Z) are characterized by the increase and establishment of tropical elements around the Pitalito Basin. The disappearance of *Quercus* around the 75 cm level (c. 450 years B.P.) is probably caused by human activity (deforestation). Deforestation probably lead to an overrepresentation of tropical elements as several of these 'warmer' taxa like *Acalypha*, *Cecropia* and *Trema* appear as secondary taxa in the open spaces. Rangel & Lozano (1986) describe a secondary forest type which replaced a *Quercus* forest. It is characterized by the presence of Euphorbiaceae, *Cecropia, Miconia, Hedyosmum* and *Pilea*. The local vegetation shows an initial rise of Caryophyllaceae, Gramineae, *Paepalanthus, Piper* and *Elaphoglossum*. A similar association was found in zone 2X and was partly ascribed to the presence of a clayey layer in which roots could anchor. *Myriophyllum, Polygonum, Typha* and Cyperaceae at the base of zone 2Z indicate the presence of small pools. This association was followed by an *Eleocharis* reedswamp community (subzone 2Z2) with Compositae (*Baccharis nitida* ?) and some ferns. Puzzling is the presence of *Acaena/Polylepis* pollen at the 85 cm level (c. 650 years B.P.). *Acaena* and *Polylepis* occur almost exclusively around the forest limit or higher upslope where they form isolated patches in the grass páramo (Cleef 1981). The following two explanations can be given for the presence of its pollen:

1/ Pollen was deposited after long-distance transport. Grabandt (1985) provides an indicator for pollen production and dispersal for several taxa: the so-called MBV (=mean background value). This value is the mean of all pollen percentages of those relevés in which the respective taxa are not present. The MBV for *Acaena/Polylepis* is 0.76 which is a value that does not exclude the possibility of long-distance dispersal. It is remarkable however, that pollen of *Acaena/Polylepis* remains restricted to one spectrum only and does not occur earlier, not even in the sediments of core PIT 11 which are assumed to be deposited during relatively cold climatic conditions.

2/ A bird visited *Acaena/Polylepis* stands in the higher altitdunal zones and left its excreements in the Pitalito Basin.

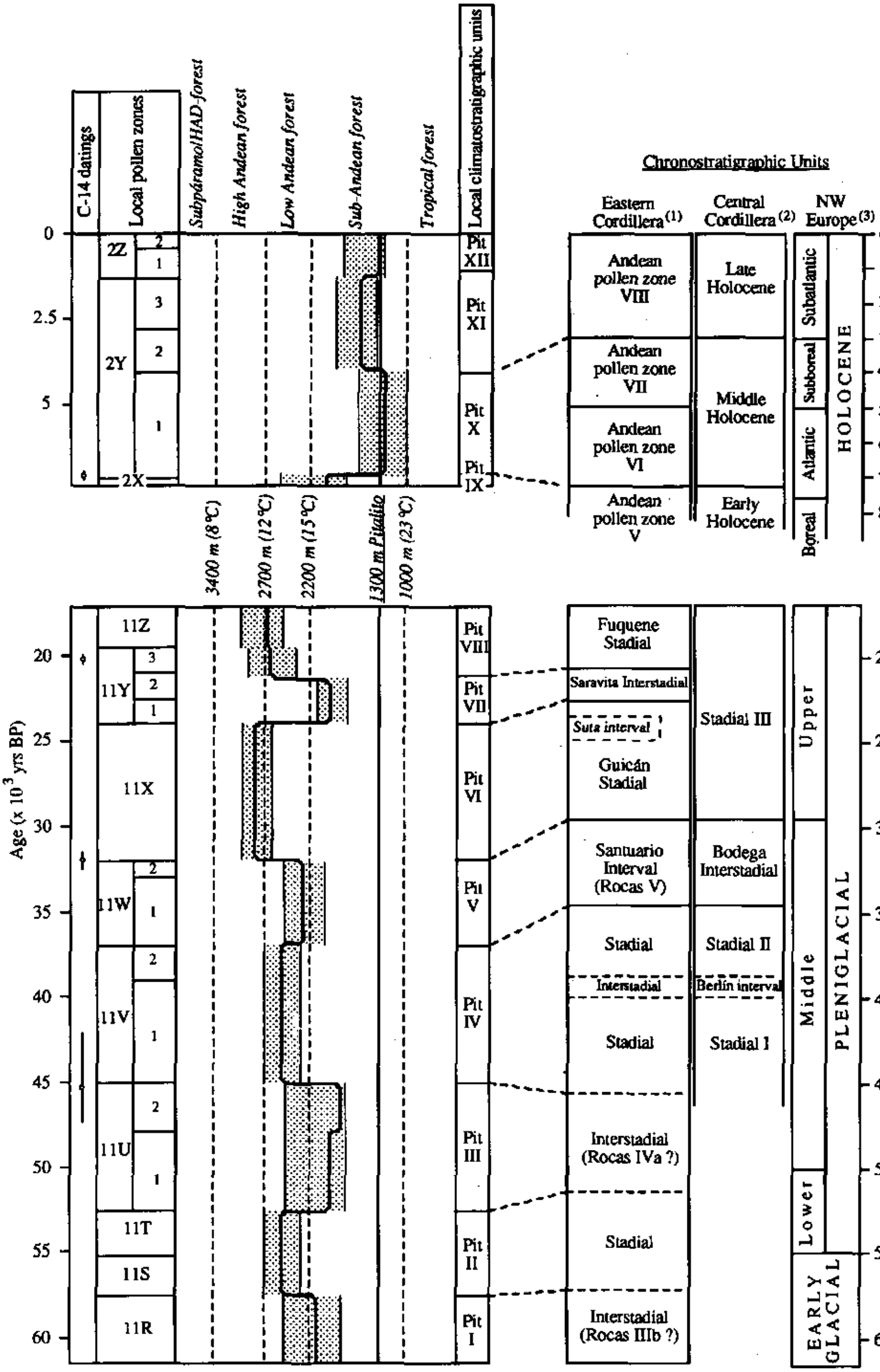
Besides *Zea mais* at least two other products were cultivated in the basin plain: *Manihot esculenta* (yuca) and *Psidium* spp. (Guayaba). The peak in the curve of *Acacia* spp. (Mimosaceae) at the 45 cm level (c. 250 years B.P.) might have to do something with cultivating activity. Several species of this genus have economical use and are imported.

5.3 Correlation and conclusions

A compilation of the regional vegetation development deduced from the pollen records PIT 11 and PIT 2 is shown in Fig. 38. The vertical shifts of the zonal forest belt at the altitude of the Pitalito Basin and the changes in the floristic composition are shown in the diagram. The local climatostratigraphic units, indicated in Roman figures, are correlated with the results of palynological studies from other parts of the Colombian Andes. Especially the correlation for the interval earlier than 45,000 years B.P. is tentative as this part is beyond the limit of thrustworthy ^{14}C dating.

From the pollen record it appears that Andean forest covered the surrounding hills of the Pitalito Basin during a considerable part of the Last Glacial. Its floristic composition varied from a *Weinmannietum*, a mixed Andean forest type and a lower Andean *Quercetum*. The latter of these three forest types occurs at the lowest elevation and is nowadays found in the area between approximately 2200 m and 2700 m altitude. This means that the sub-Andean/Andean forest boundary was lowered at least about 1000 m compared to the modern situation. Accepting an average lapse rate of 0.60°C/100 m altitude this would mean that the average annual temperature lowered about 6°C during some periods in the Last Glacial.

The palynological record starts with unit Pit I (Fig. 38), a relatively warm phase during the Early Glacial. The mean annual temperature was about 4-5°C lower than today. Based on the calculated ages this unit may be contemporaneous with the Rocas IIIb Interval (≈60,000 years B.P.; Schreve-Brinkman 1978). However, she recorded at that time temperatures that are comparable with the present ones. Unit Pit II represents a colder phase during which temperatures declined. The mean annual temperature around the Pitalito Basin probably was about 13°C which is c. 7°C lower than modern temperature. Until yet, the existence of this colder phase is not well described. Schreve-Brinkman (1978) describes a relatively cold and wet phase of about the same age at the El Abra corridor (subzone IIe) at an altitude of 2570 m. According to her data, temperature was about 7°C lower then the modern temperature which would be consistent with the data from the Pitalito Basin. Unit Pit III is characterized by an increase of *Quercus* and represents a warmer phase although temperature was still about 4°C lower than at present. At the end of this period the temperature possibly increased further with c. 1°-2°C. Van der Hammen & González (1963) and Van der Hammen (1974) reported for the highplain of Bogotá a change to somewhat drier conditions at the end of the Early Glacial which can partly explain the increase of *Quercus*. Drier conditions however, are contradicted by Schreve-Brinkman (1978) who describes for the same area an albeit poorly recorded period of relatively warm and extremely wet conditions around that time (Rocas IVa Interval). During the succeeding unit Pit IV mean annual temperatures declined with 7°C compared with the present-day situation. It correponds with a colder period which has also been registered in the Central Cordillera (Stadial I & II; Salomons 1986) as well as in the Eastern Cordillera (Van der Hammen *et al.* 1981). Salomons (1986) calculated a lowering of the timberline with c. 1300 m compared with the modern situation implying that the mean annual temperature had dropped with 8°-9°C as compared with today.



(1): Van der Hammen & González, 1960, 1965a and 1965b; Van Geel & Van der Hammen, 1973; Schreve-Brinkman, 1978; Van der Hammen et al., 1981; Kuhry, 1988. (2): Salomons, 1986. (3): E.A. Koster, 1980; E. Kolstrup, 1980.

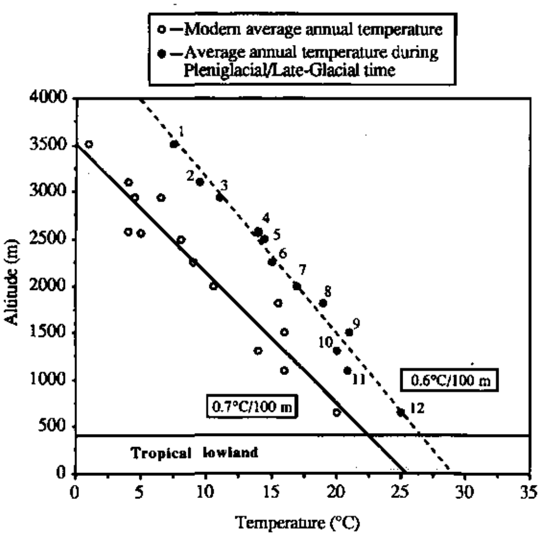
Glaciers descended to an altitude of 2500-2200 m in the Cocuy area (Eastern Cordillera; Van der Hammen *et al.* 1981) and to an altitude between 2700-3400 m in the Tolima/Ruiz area (Central Cordillera; Thouret 1983). Possibly this period of maximum glacial extent is also recorded around the highplain of Bogotá (Morainic complex 1; Helmens 1988) and is partly ascribed to a higher amount of precipitation during the earlier stadials of the Last Glacial (Van der Hammen *et al.* 1981). The increase of *Weinmannia* at the end of unit Pit IV might possibly be ascribed to a somewhat colder climate and/or more humid conditions. The warmer Berlín interval around 38,000 years B.P. (Salomons 1986) is not registered in the Pitalito pollen diagram. Unit Pit V represents a somewhat warmer phase with possible drier conditions. Compared to the preceeding phase, the mean annual temperature possibly rose with approximately 2°C. This amelioration has been recorded in both mountain chains (Salomons 1986; Schreve-Brinkman 1978). During unit Pit VI the mean temperature in the Pitalito Basin declined to approximately 8°C below the present-day level. This period is possibly contemporaneous with the Guicán Stadial (Van der Hammen *et al.* 1981) and the start of Stadial III (Salomons 1986). The Guicán Stadial comprises the Suta Interval as described by Van Geel & van der Hammen (1973). They calculated a mean annual temperature of about 10°C below the present-day level at the altitude of the highplain of Bogotá and suggest drier conditions. This cold phase is interrupted around 24,000 years B.P. by a brief period with higher temperatures (unit Pit VII). This warmer phase has not been recorded in the Central Cordillera (Salomons 1986). Around 21,000 years B.P. (unit Pit VIII) elements increased which are nowadays present around 3000 m altitude. If only a fall in temperature was responsible for the lowering of the forest belts with about 1700 m, this would mean a decrease of the average annual temperature with c. 10°C compared to the actual situation. However, such a decline of temperature must be considered as a maximum as also the drier conditions may have favoured the settlement of *Myrica* scrubs at low elevations. A decrease of c. 6-8°C is therefore more likely. Unit Pit VIII is contemporaneous with the beginning of the very cold and dry Fuquene Stadial which lasted from c. 22,500 to c. 14,000 years B.P.(Van Geel & van der Hammen 1973; Kuhry 1988). According to Kuhry (1988) *Myrica* scrub páramo was present around 2000 m altitude on the eastern slopes of the Magdalena Valley at that time. This would mean that páramo vegetation was able to reach the crests of the mountains surrounding the Pitalito Basin. Van der Hammen (1974, 1981, 1989) stated that during that same period dry tropical woodland ascended along the eastern slopes of the Magdalena Valley to an altitude of about 2000 m and that the open páramo vegetation was in contact with open xerophytic vegetation, leaving no forest at all. This means that the xerophytic-like vegetation, actually present in the Upper Magdalena Valley near Altamira (Fig. 1), could migrate towards the south and came in the reach of the Pitalito Basin. The Pitalito Basin became 'sandwiched' between an ascending xerophytic type of vegetation and a descending scrub páramo vegetation.

Fig. 38.
Changes in the floristic composition of the zonal forest around the Pitalito Basin during the last 62,000 years and its correlation with more regional registered fluctuations in the Colombian Andean region. The approximate average annual temperature around the Pitalito Basin at a specific time interval and its upper and lower limit are represented by the bold line and a stippled area, respectively.

The high percentages of *Myrica* pollen can be ascribed to the nearby presence of a páramo scrub vegetation, a xerophytic-like vegetation or both as this element shows a wide vertical distribution due to its pioneer qualities in the Andean regions (Rangel & Lozano 1986). The presumed presence of a sparse vegetation cover is consistent with the fact that around the same time the eastern flood basins received clastic material due to a change in alluvial architecture (section 4.1.2). The sparse vegetation cover in combination with a dry climate in which torrential rains may prevail, made the unprotected bare soil vulnerable to erosion. *Myrica* is a well-known pioneer on unstable soils (Salomons 1986). Sediment-loaded water was irregularly supplied to the river system which lead to frequent overflooding of the river banks.

After an interruption in the palynological record of about 10,000 years, pollen registration started again around 7300 years B.P. in the La Coneca area. Core PIT 2 was obtained from this area and represents a part of the Holocene period. The pollen diagram starts with the registration of the latest phase of a relative cold period (unit Pit IX) which ended around 7200 years B.P. This period is contemporaneous with the end of the Andean pollen zone V which lasted from 9000 to c. 7500 years B.P. (*e.g.* Van der Hammen & González 1960, 1965a, 1965b) and marks the transition from the relatively cold Late Glacial to the warmer Holocene. Due to the wide ecological range of the genus *Hedyosmum* it is difficult to come at a conclusion about the prevailing climatic conditions. The temperature was possibly some 4°C lower than today. Around 7200 years B.P. the modern values of temperature had been reached and tropical/sub-Andean taxa settled around the basin. The somewhat higher temperature between 7200 and 5000 years B.P. (subzone 2Y1) has been reported by many authors for the Central Cordillera (*e.g.* Melief 1985; Piñeros 1988; Bakker & Salomons 1989) as well as for the Eastern Cordillera (Van der Hammen & Gonzalez 1960b; Schreve-Brinkman 1978; Kuhry 1988). It is referred to as 'Bioclimatic Optimum' or 'hypsithermal' and is contemporaneous with the Andean pollen zones VI and VII. The climate was about 2-3°C warmer than today and more humid. The last 1100 years (unit Pit XII) are characterized by an increase of tropical elements which must partly be ascribed to human influence.

Interesting is the decline of temperature with altitude during Pleniglacial time in northern South America. According Van der Hammen (1979, 1989) a steeper thermal gradient would be present during glacial times as the depression of temperature may be modest in the tropical lowlands of South America (c. 3°C below c. 400 m altitude) whereas at higher altitudinal levels the temperature decline lies in the order of 6-8°C. The latter is consistent with data obtained from the Colombian Andean regions above c. 2500 m altitude (*e.g.* Van Geel & van der Hammen 1973; Van der Hammen 1974; Schreve-Brinkman 1978; Van der Hammen *et al.* 1980/1981; Hooghiemstra 1984; Melief 1985; Salomons 1986; Kuhry 1988; Piñeros 1988). These data together with the modern average annual temperature and temperature gradient (approx. 0.60°C/100 m altitude) are shown in Fig. 39. Less data are available from the lower parts of the Colombian Andean region. Cleef *et al.* (in prep.) report from an area at c. 2250 m altitude near Manizales (Central Cordillera) a maximum temperature decline of c. 6°C around 45,000 years B.P. Van der Hammen (1974, 1981) and Kuhry (1988) describe the lowering of the forest line



SOURCES

1/ La Ciega (3510 m); Van der Hammen et al. (1980/1981).
2/ Alsacia (3100 m); Melief (1985).
3/ TPN 40, TPN 21 (both 2940 m); Salomons (1986).
4/ Fuquene (2580 m), El Abra (2570 m), Sabana de Bogotá (c. 2550 m); Van Geel & Van der Hammen (1973), Schreve-Brinkman (1977), Kuhry (1988).
5/ Merenberg (2500 m); Piñeros (1988).
6/ Otoño-Manizales Enea (2250 m); Cleef et al. (in prep.).
7/ Pedro Palo (2000 m); Van der Hammen (1974).
8/ Libano (1820 m); Salomons (1986).
9/ El Dorado (1500 m); Monsalve (1985).
10/ Pitalito (1320 m); Bakker (this study).
11/ Mera (1100 m); Liu & Colinvaux (1985).
12/ La Yeguada (650 m); Piperno et al. (1990).

Fig. 39
Variation of modern and Pleniglacial/Late-Glacial mean annual temperature with altitude in the northern Andean region.

to 2000 m altitude along the dry eastern slopes of the Magdalena Valley (section Laguna Pedro Palo). Xerophytic vegetation seems to be in direct contact with a *Myrica* scrub páramo. The forest limit possibly lowered some 1300 m due to dry and cold climatic conditions and indicates a lowering of temperature of c. 6°-7°C at this altitude. Salomons (1986) reports the presence of *Quercus* forests at an altitude of 1820 m (section El Libano; Central Cordillera) during the Upper Pleniglacial. He proposed a lowering of about 500 m for these forests but more probably this figure lies in the order of 600 to 800 m as nowadays these forests are present between 2500 and 3100 m altitude. Monsalve (1985) describes a palynological record from an approximate altitude of 1500 m (section El Dorado; Central Cordillera). *Quercus* forests dominated at this altitude between c. 40,000 and 10,000 years B.P. together with *Hedyosmum* and *Ilex*. This may possibly indicate a lowering of the Andean forest of at least 700 (more probably 900 to 1000 m) and a minimum temperature decline of c. 4.5°C. This present study shows a temperature decline of about 6°C during a considerable part of the Last Glacial around the Pitalito Basin (1300 m). Colinvaux and co-workers (1987a, 1987b, 1987c) and Liu & Colinvaux (1985) calculated a minimal depression of mean annual temperature with c. 4.5°C around 30,000 years B.P. for the upper limit of the present Amazonian rainforest in Ecuador (alt. c. 1100 m). This figure might be flattered as these calculations are mainly based on the findings of fossil *Podocarpus* wood and pollen at this elevation. Gentry (1986) reports *Podocarpus* from west Colombian lowland forests and Van der Hammen (oral information) found *Podocarpus* in the western and very wet Amazone Basin. Therefore, the presence of *Podocarpus* at low elevations does not necessarily claim a significant lowering of temperature. Piperno *et al.* (1990) raised cores from the Panamese Lake La Yeguada (alt. 650 m) for a.o. pollen analysis. They report that, during the Late Glacial, arboreal taxa were present around this lake which are currently found at c. 1500-2500 m altitude. Based on these findings, they calculated for this period a temperature depression of c. 5°C compared to modern temperature.

The assumption for a modest decline in temperature in the tropical lowlands are based on the extrapolation of oceanic surface temperature during the ice-age (CLIMAP 1976) as most pollen diagrams of the tropical South American lowlands only register fluctuations in precipitation (*e.g.* Wijmstra & van der Hammen 1966; Van der Hammen 1974, 1983; Absy 1979; Bradbury *et al.* 1981) without deducing data about thermal conditions. The report shows areas in the mid-latitude ocean with surface temperatures around 18,000 years B.P. lying within 2°C of modern temperature. These glacial sea-surface temperatures are deduced from a.o. plaktonic biota. Such biota might be influenced by local upwelling features leading to the presence of relatively cold water at the sea surface. In such a case 'cold' planktonic groups may be present in areas which were not affected by a decline of temperature. So the extrapolation of oceanic surface temperature to land temperature might be hazardous as water masses are controlled by oceanic circulation. Rind & Peteet (1985) uniformely lowered the CLIMAP sea-surface temperatures by 2°C to produce a better fit to the continental data set of o.a. the Andean regions of Colombia. Furthermore, the paucity of data of sea-surface conditions may prevent us from a reliable conclusion. Fig. 39 shows data of the average annual temperature in the Northern Andean region during the Pleniglacial time varying with altitude. These data reveal a somewhat steeper temperature gradient during the ice-age compared to the modern time. Extrapolation of these data to lower elevation suggests that in the tropical lowlands the average annual temperature may have declined with some 3°-4°C during full glacial time. In order to evaluate the problem, more data from low altitudinal levels are required but it seems that at low elevations the glacial temperature depression is less compared with that at high altitudes and that the thermal gradient may have been in the order of 0.7°C/100 m.

The presence of pollen and spores from taxa occurring around or above the forest line may be explained in most of the cases by local edaphic conditions. In recent studies of the present vegetation Rangel & Lozano (1986) and Rangel & Lozano (in prep.) encountered local stands of a páramo-like vegetation at an altitude of 2300 m in the Merenberg area in the Central Cordillera (nearby the Pan de Azucar volcano; Fig. 8). This azonal vegetation type was found under local marshy conditions. About 60% of the elements recorded in this open type of vegetation are characteristic for the páramo belt: *Espeletia hartwegiana* subsp. *centroandina*, *Hypericum lancioide*, *Sphagnum magellanicum*, *Blechnum columbiensis*, *Paepalanthus columbiensis*. The zonal vegetation is made up of a *Quercetum* with *Hedyosmum*. They ascribe the presence of these páramo elements to edaphic conditions (a badly-drained area, low pH) which impede the growth of trees so that a more open type of vegetation can establish at lower elevations. From this study it is obvious that similar edaphic conditions prevailed in the Pitalito Basin during the Pleniglacial. As edaphic conditions play a decisive role in the occurrence of this type of vegetation this páramo-like vegetation must be considered as azonal. It can not be excluded however, that pollen of these 'upper-belt' elements has been deposited after long-distance dispersal by air or water. The latter is especially possible if pollen is present in clay-rich strata.

The coincidence of changing climatic conditions and the increase of minerogenic sediments due to a change in alluvial architecture around 20,000 years B.P. is without reasonable doubt; the minerogenic influx increased at least over the entire eastern flood basin and was not merely a local event. However, one have to consider the possiblity that also other strata then the well-documented marker horizon might be positively related to a change in climatic conditions. For instance, the minerogenic layer in core PIT 11 between 1110 and 900 cm depth clearly coincides with an increase of 'warmer elements' between c. 52,500 and 45,000 years B.P. (unit Pit III) and might be ascribed to an ameloriation of the climate. The inorganic sediments are not correlative with minerogenic strata of other boreholes from a lithostratigraphical point of view. A chronostratigraphical and biostratigraphical correlation of the sediments would be benificial in order to find out whether we are dealing with a local event of increased sedimentary influx inherent to the river regime (autocyclic control), or that extrabasinal parameters like tectonic adjustments or climatic conditions (allocyclic control) are responsible for this change in lithology.

5.4 Description of pollen and (asco-)spore types

Several identified and unindentified pollen and (asco-)spore types are published here and illustrated in the following photoplates. A few types, described and illustrated earlier, were included when providing additional information. Pollen was identified with the aid of the collection of recent pollen, present at the Hugo de Vries laboratory, University of Amsterdam, and the following publications: Heusser (1971), Van Geel & van der Hammen (1973), Van Campo (1974), Markgraf & D'Antoni (1978), Bonnefille & Riolet (1980), Hooghiemstra (1984), Salomons (1986) and Kuhry (1988). The description of the pollen types as given below is mainly after Iversen & Troels-Smith (English version, 1980), Erdtmann (1952) and Faegri & Iversen (1975). The suffix 'type' is used because the taxon under discussion morphologically resembles the fossil type. As a rule measurements are based on 5 to 10 pollen grains. When it was not possible to reach this amount it is indicated (n). Of length and breadth the mean values as well as their standard deviations (between brackets) are given. Of all other measurements only the mean values are reported. When it was not possible to find at least five grains, only the mean values are given. Pollen types which could not be identified as belonging to any natural taxon are designated by their aperture-formula followed by the sculpture type (*e.g.* C_3P_3 ret=tricoloporate, reticulate). If not otherwise indicated, all scales are x1000. The enumeration of the microfossil types is continuous with that of Hooghiemstra (1984), Salomons (1986), Kuhry (1988) and Wijninga & Kuhry (in press).

POLLEN TYPES

Type 643: Leguminosae. Plate I.
Pollen grains tricolporate, radially symmetrical, prolate to oval, amb circular; tectate, reticulate. Colpi long and narrow with costae increasing towards the pori. Porus distinct, meridionally elongated (lg. c. 2.1 µm). Exine c. 1 µm thick. Polar area medium. Lumina isodiametrical. Size: 18.0(0.5) x 14.5(1.5) µm.

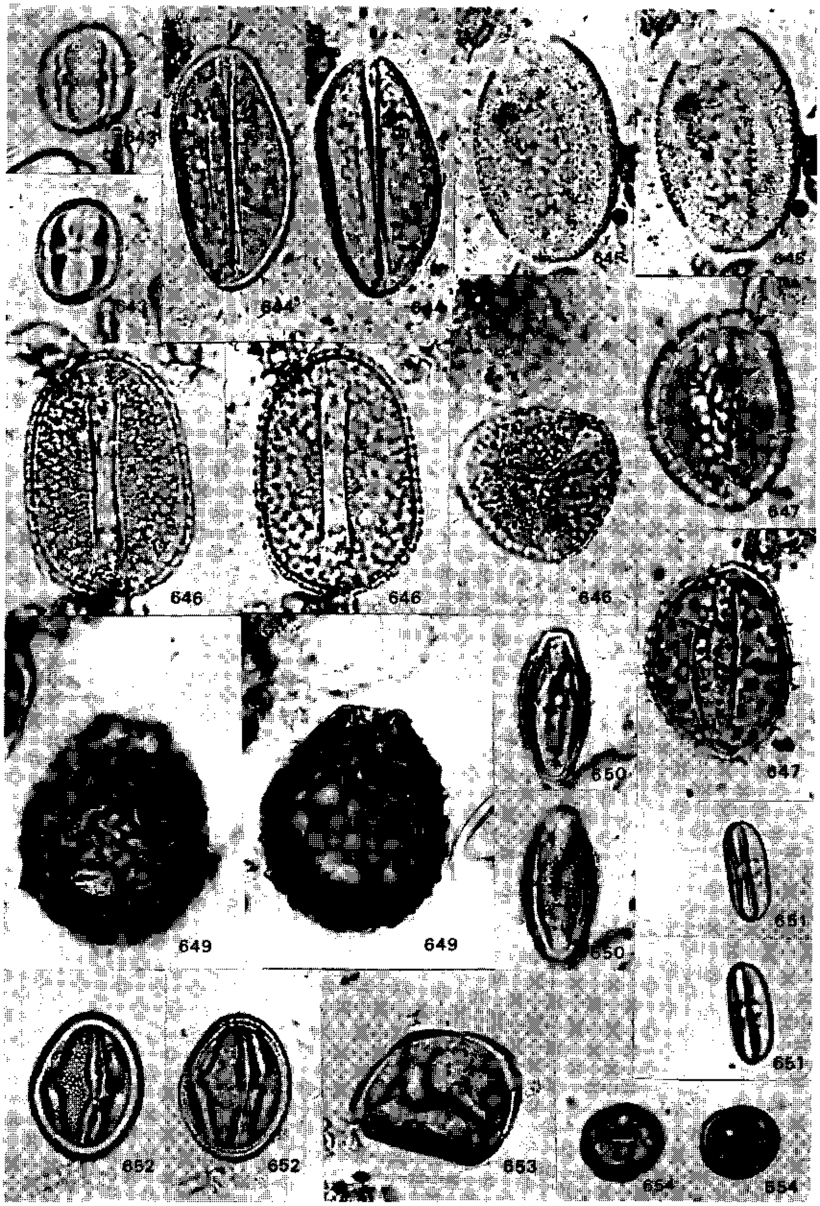
Type 644: Palmae-type 1. Plate I.
Pollen grains monocolpate, bilaterally symmetrical, amb elliptical; tectate, scabrate to verrucate. Colpus (c. 0.5 µm wide) as long as equatorial diameter, psilate margins. Exine c. 1.1 µm thick. Size: 40.8(1.8) x 23.7(4.4) µm.

Type 645: Palmae-type 2. Plate I.
Pollen grains monocolpate, bilaterally symmetrical, amb elliptical; semi-tectate, reticulate. Colpi c. 21.6 µm long. Exine c.1.3 µm thick. Size: 36.4(1.1) x 25.5(1.7) µm.

PLATE I

Type 643: Leguminosae. Type 644: Palmae-type 1. Type 645: Palmae-type 2. Type 646: Palmae-type 3. Type 647: Palmae-type 4. Type 649: Papillionidae. Type 650: *Phyllanthus cf. caroliensis* (Euphorbiaceae). Type 651: Begoniaceae. Type 652: *Clusia* (Clusiaceae). Type 653: Campanulaceae. Type 654: *Bunchosia cf. pseudinitida* (Malpighiaceae).

PLATE I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Type 646: Palmae-type 3. Plate I.
Pollen grains monocolpate or trichotomocolpate, bilaterally symmetrical or radially symmetrical, amb elliptical or sub-triangular with convex walls; semi-tectate, reticulate. Colpus c. 30.5 μm long and 4.2 μm wide, psilate margins. Exine c. 1.5 μm thick. Luminae polygonal shaped (c. 2 μm wide). Size: c. 41.2(2.1) x 28.5(1.3) μm.

Type 647: Palmae-type 4. Plate I.
Pollen grains monocolpate, bilaterally symmetrical, amb elliptical; intectate, echinate. Colpus of moderate length (c. 26.5 μm long, 6.6 μm wide) with ragged margins. Echinae (length 2.5 to 3.2 μm, diam. at base 1.5 to 2.1 μm) are evenly distributed and seated in pockets, perpendicular on exine. The necks of the echinae are sharply pointed and do not bend or hardly so. Exine c. 1.3 μm thick. Size: 32.3(1.7) x 26.1(0.9) μm.

Type 648: Deleted

Type 649: Papillionidae. Plate I.
Pollen grains triporate, subsphaeroidal to oblate, radially symmetrical; semi-tectate, coarse-reticulate. Pori circular (diameter c. 8.0 μm). Exine c. 1.4 μm thick. Lumina more or less sphaeroidal and isodiametric (c. 6 μm wide). Relatively small muri (0.5 μm). Size (n=1): c. 38 x 32 μm.
Occurred in PIT 11 at depth 12.45 m.

Type 650: *Phyllanthus cf. caroliensis* (Euphorbiaceae). Plate I.
Pollen grains tricolporate, prolate, radially symmetrical, amb circular; tectate, scabrate. Colpi long with costae, endopori elliptical. Exine c. 0.8 μm thick, thickening towards the pori and the poles to c. 1.5 μm. Size (n=1): c. 25.2 x 11.5 μm. Lit. Punt & Bentrop (1974).

Type 651: Begoniaceae. Plate I.
Pollen grains tricolporate, prolate to perprolate, bilaterally symmetrical, amb circular; semi-tectate, striate. Exine c. 0.5 μm thick. Size: 18.2(1.7) x 8.3(0.9) μm.

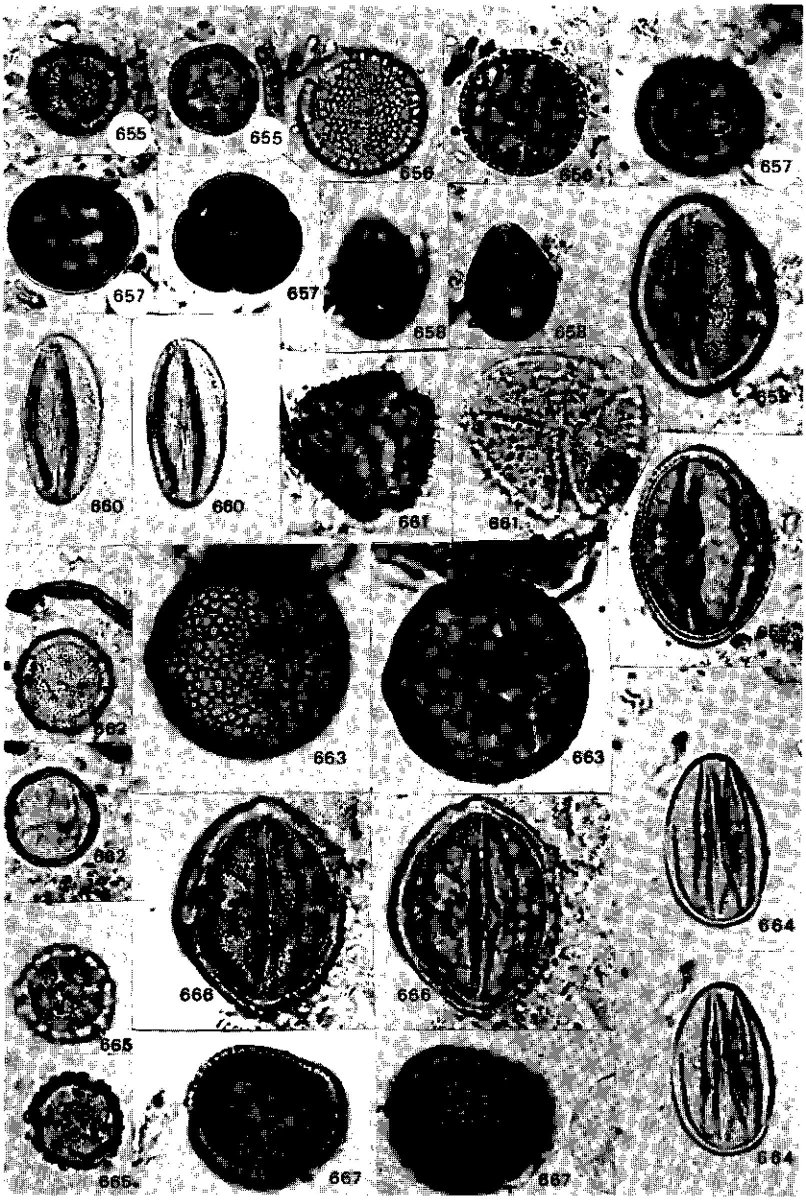
Type 652: *Clusia* (Clusiaceae). Plate I.
Pollen grains tricolporate, prolate, radially symmetrical, amb circular; tectate, micro-reticulate to reticulate. Colpi long with costae. Endopori distinct (diameter c. 3 μm). Exine c. 2 μm thick. Size 25(1.7) x 17.6(2) μm.

Type 653: Campanulaceae. Plate I.
Pollen grains peri(5)porate, circular, radially symmetrical, amb circular; intectate, micro-

PLATE II

Type 655: Menispermaceae. Type 656: cf. Menispermaceae. Type 657: *Norantea*-type (Marcgraviaceae). Type 658: Proteaceae. Type 659: C_3P_3 scabr. Type 660: C_3P_3 ret. Type 661: C_3P_3 ech. Type 662: C_3P_3 microret. Type 663: C_3P_3 ret. Type 664: C_3 psi. Type 665: C_3P_3 verr. Type 666: C_3P_3 scabr. Type 667: C_3(?)P_3 ret.

PLATE II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echinate. Large pori (diam. c. 5.3 μm) with indistinct annuli. Exine c. 2 μm thick. Echinae are sparsely distributed and indistinct. Equatorial diameter (n=3): c. 37.5 μm. Included in the curve of Pollen Div.

Type 654: *Bunchosia cf. pseudinitida* (Malpighiaceae). Plate I.
Pollen grains peri(9)porate, circular, radially symmetrical, amb circular; intectate, psilate. Pori well rounded (diameter c. 2μm). Exine c. 1 μm thick. Equatorial diameter c. 13 μm. Included in the curve of the Malpighiaceae. Occurred in PIT 11 between 9.75 and 10.95 m.

Type 655: Menispermaceae. Plate II.
Pollen grains tricolporate, radially symmetrical,.subsphaeroidal, amb triangular; semi-tectate, reticulate. Colpi long with distinct and smooth margin. Endopori circular (diam. c. 2 μm) and indistinct. Muri narrow. Lumina more or less angular and irregular (diam. c. 1.5 μm). Exine c. 1.7 μm thick. Short but distinct columellae. Size: 16.0(1.5) x 15.0(2.1) μm. Lit. Thanikaimoni (1984): Plate 67.

Type 656: cf. Menispermaceae. Plate II.
Pollen grains tricol(por?)ate, subsphaeroidal to circular radially symmetrical, amb circular; semi-tectate, reticulate. Colpi long and distinct. Endopori indistinct with ragged margins. Exine c. 1.5 μm thick. Lumina more or less sphaeroidal, differing in size (1.5 μm< ø <3.5 μm), decreasing towards the colpi. Size: 21.5(1.6) x 22.5(1.3) μm. Lit. Thanikaimoni (1984): Plate 67.

Type 657: *Norantea*-type (Marcgraviaceae). Plate II.
Pollen grains tricolporate, radially symmetrical, oblate, amb circular; tectate, psilate, tectum with perforations. Colpi short and distinct. Endocolpi narrow with costae. Polar area very large. Exine c. 2.5 μm thick. Columellae indistinct. Size (n=3): 21.5 x 21.0 μm. Lit. Punt (1971).

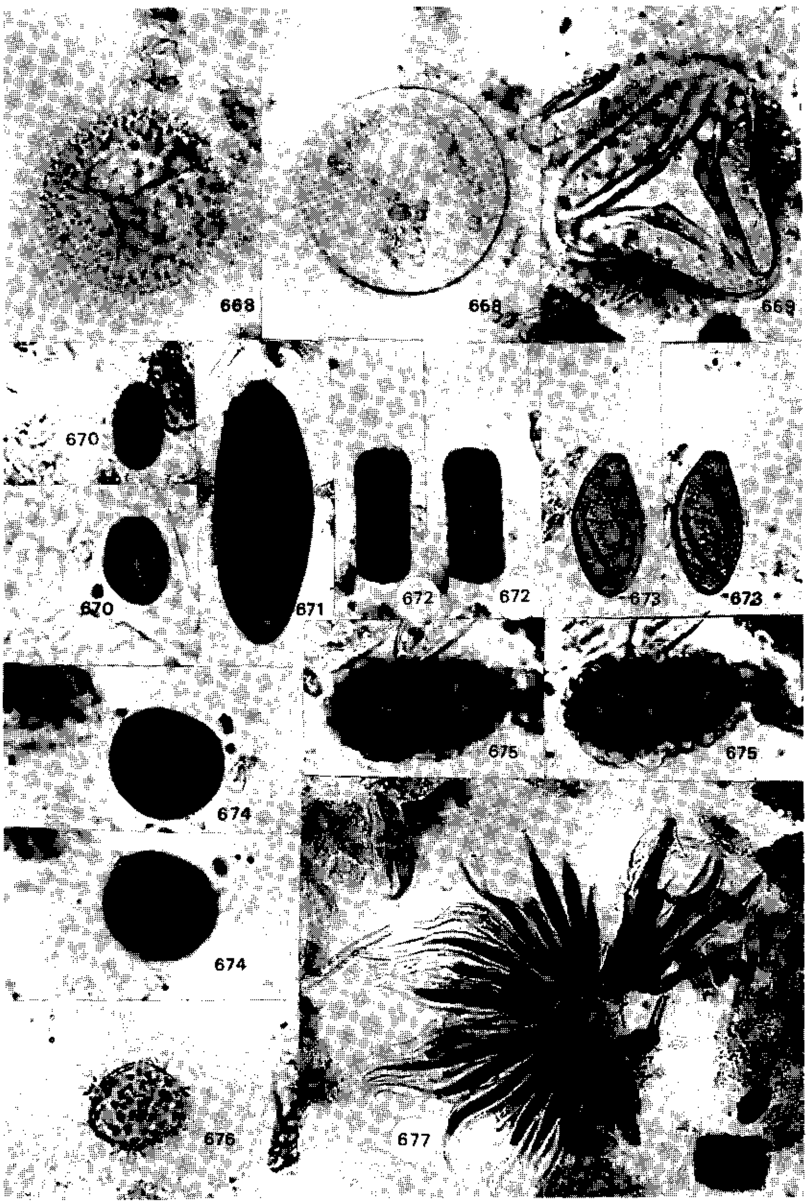
Type 658: Proteaceae. Plate II.
Pollen grains triporate, subsphaeroidal to oblate, radially symmetrical, amb circular; tectate, psilate. Pori subsphaeroidal (diam. c. 3.5 μm) with ragged margins. Exine c. 1.2 μm thick, thickening towards the pores (c. 2 μm). Size (n=1): c. 15.6 x 20.0 μm. Included in the curve of 'other pollen'

Type 659: C_3P_3 scabr. Plate II.
Pollen grains tricolporate, prolate to subsphaeroidal, radially symmetrical; tectate, scabrate. Colpi long with broad costae. Endopori distinct (c. 4.0 μm) with costae. Exine c. 1.1 μm thick, thickening towards the poles c. 1.7 μm due to longer columellae.

PLATE III

Type 668: Trilete spore (x750). Type 669; Adiantaceae-type. Type 670: Fungal spore. Type 671: Fungal spore (x500). Type 762: Fungal (?) spore. Type 673: Fungal spore. Type 674: Ascospore. Type 675: Ascospore. Type 676: Unknown micro fossil. Type 677: Peltate leaf hair (x500).

PLATE III



Collumellae distinct and regularly distributed. Nexine and columellae resp.thicker, higher towards the poles. Thick tectum. Size: 30.5 (2.0) x 21.5 (2.0) µm.

Type 660: C_3P_3 ret. Plate II.
Pollen grains tricolpate, perprolate, radially symmetrical; tectate, reticulate. Colpi long and narrow. Polar area small. Exine c. 1 µm thick. Size: 27.0 (1.6) x 12.5 (0.8) µm.

Type 661: C_3P_3 ech. Plate II.
Pollen grains tri- or syncolporate, radially symmetrical with straight sides and blunted angles, amb angular; tectate, echinate. Colpi long with indistinct pori. Polar area small. Echinae (2 µm long) irregular in size, sharp pointed ends, relatively broad, irregular-rounded base (1 µm) and are densely distributed. Exine c. 1 µm thick. Columellae indistinguishable. Size (n=4): 25.0 x 23.0 µm.

Type 662: C_3P_3 microret. Plate II.
Pollen grains tricolporate, radially symmetrical, subsphaeroidal to depressed oval, amb circular; semi-tectate, micro-reticulate(?). Colpi narrow and moderately long. Polar area large. Pori indistinct. Exine c. 1.1 µm thick. Size: 16.3(1.8) x 15.8(1.7) µm.

Type 663: C_3P_3 ret. Plate II.
Pollen grains tricolporate, radially symmetrical, subsphaeroidal, amb circular; semi-tectate, reticulate. Colpi of moderate length with costae increasing towards the pori. Broad colpi transversalis. Exine c. 1.6 µm thick, thickened at the inter colpia (c. 2.1 µm). Muri almost as wide as the lumina diameter. Lumina approximately isodiametrical. Equatorial diameter (n=3): c. 44.0 µm.

Type 664: C_3 psi. Plate II.
Pollen grains tricolpate, radially symmetrical, prolate to oval, amb circular; tectate, psilate. Colpi very long with costae. Polar area small. Exine c. 1.5 µm slightly increasing towards the poles (1.8 µm). Size: 30.0(0.6) x 17.5(2.4)µm.
Included in curve of Pollen Div. Occurred in PIT 11 at depths 4.05 m and 10.65 m.

Type 665: C_3P_3 verr. Plate II.
Pollen grains tricolporate, radially symmetrical, subsphaeroidal, amb semi-angular; intectate, verrucate. Exine c. 0.5 µm thick. Verrucae more or less circular (diameter c. 2 µm) and c. 2 µm high. Equatorial diameter (n=1): c. 16 µm.
Included in curve of Pollen Div. Occurred in PIT 11 at depths 3.15 m and 3.75 m.

Type 666: C_3P_3 scabr. Plate II.
Pollen grains tricolporate, radially symmetrical, prolate to perprolate, amb circular; tectate, scabrate. Colpi elongate with broad costae increasing towards the pori to c. 2 µm. Porus circular (diam. c. 4 µm). Polar area small. Exine c. 2 µm thick. Distinct columellae (c. 1.5 µm high). Size (n=1): c. 35 x 27 µm.
Included in curve of Pollen Div. Occurred in PIT 11 at depth 3.60 m.

Type 667: C_3(?)P_3 ret. Plate II.
Pollen grains tri(col?)porate, radially symmetrical, subsphaeroidal amb semi-angular; semi-tectate, reticulate. Exine c. 2.2 µm thick. Distally branched columellae. Muri as broad as the lumina. Equatorial diameter: 26.7(2.7) µm.
Included in curve of Pollen Div. Occurred in PIT 11 at depths 3.75 m, 7.20 m and 10.80 m.

(ASCO-)SPORE TYPES

Type 668: Plate III.
Spores trilete, radially symmetrical, amb circular; clavate to bacculate. Laesurae almost extending to the equator. Clavae/bacculae 3.5(0.9) µm high. Sclerine c. 2-3 µm thick. Equatorial diameter: c. 52(7) µm.

Type 669: Adiantaceae-type. Plate III.
Spores trilete, radially symmetical, amb circular to subtriangular convex; psilate. Laesurae almost extending to equator, margo c. 3 µm wide. Sclerine c. 2 µm thick, thickening towards the equator to c. 9 µm. Equatorial diamter c. 43(5) µm. Lit. Murillo & Bless (1974): Plate 4 and Hooghiemstra (1984): Plate 36

Type 670: Plate III.
Fungal spores ellipsoidal, one-celled, dark brown. Ornamented with 10-15 µm longitudinal grooves over the whole spore length (the exact number is difficult to observe; some of the grooves anastomose and do not extend over the whole spore length).Size c. 16 x 10 µm.

Type 671: Plate III.
Fungal spores (conidia?) 9-septate, not constricted at the septa, light-brown, smooth surface. Size c. 86 x 31 µm. Lit. Hooghiemstra (1984): Plate 44.

Type 672: Plate III.
Fungal (?) spores cylindrical, one-celled, dark-brown. With c. 10 longitudinal wide furrows (the exact number is difficult to observe). Size: c. 21 x 9 µm.

Type 673: Plate III.
Fungal spores one-celled, ellipsoidal, brown. Surface covered with numerous small pits. Obliquely truncate at the proximal end provided with a sunken pore. Size 15-29 x 8-14 µm. Lit. Pals *et al.* (1980): Plate II.

Type 674: Plate III.
Ascospores globose to ovoid, one-celled, brown. Surface showing a finger-print like striation. Size c. 19 x 17 µm

Type 675: Plate III.
Ascospore uniseptate, light-brown. Surface with anastomosing epispore 2-5 µm high.

Size (excluding the epispore) c. 29 x 15 µm.
Lit. Melief (1985): Plate III.

Type 676: Plate III.
Unknown micro fossil, globose to ovoid, wall c. 0.5 µm thick. Ornamented with 1-2 µm high echinae, c. 2 µm apart. Size (excluding the echinae) c. 17 x15 µm.

Type 677: Plate III.
Peltate leaf hair, diameter c. 150 µm.

CHAPTER 6

SYNTHESIS

Alluvial architecture and the degree of interconnection of channels are controlled by a variety of interrelated parameters that may be either allocyclic (extrabasinal) or autocyclic (intrabasinal) in character. Allocyclic controls include tectonic adjustments in the depositional basin and/or its source area, and climatic conditions in both sediment provenance areas and the basin of deposition. These factors work in concert to regulate variations in stream discharge, sediment size, valley slope, vegetation cover and ultimately alluvial architecture. Autocyclic control, on the other hand is inherent to the river regime and includes channel migration and avulsive events. In many alluvial suites it is not possible to identify a single dominant parameter controlling alluvial architecture. Information concerning both geological and (paleo)climatological parameters is often uncertain. Nonetheless, knowledge of these factors is critical to a meaningful evaluation of their contribution to overall architectural motifs.

This study reveals two types of allocyclic controls: tectonics and climate. These affect the sedimentation processes in the Pitalito Basin on two different scales.

TECTONIC CONTROL

The Pitalito Basin affords a typical example of a strike-slip controlled intramontane basin:
- The basin is asymmetrical with a shallow western part (c. 300 m deep), a deep eastern part (c. 1200 m deep) and with its structurally deepest part close to the active, northern, strike-slip margin;
- An asymmetrical margin topography with an active northern margin and a relatively passive southern margin;
- Small, debris-flow-dominated alluvial fans along the active margin and large streamflow-dominated fans along the inactive southern margin;
- The basin infill is characterized by thick sedimentary sections compared to the area, the abrupt lateral facies changes and the stable position of these changes in time;
- Axial infilling of the basin, subparallel to the strike-slip basin margin;
- A migration of the depocenter in time in the opposite direction to basin extension.

The Pitalito Basin developed as a result of extension along a fault junction (Fig. 17). It is the first basin of this type which is described in detail in Colombia. The geological structures in and around the Pitalito Basin are genetically related to the north-south trending Garzón-Suaza fault. It appears that in the Pitalito Basin tectonics plays a most prominent role in basin morphology and governs the large scale distribution of the sediments: coarse-grained sediments in the shallow western part of the basin and fine-grained sediments in the deep eastern part of the basin.

CLIMATIC CONTROL

The effect of climate on sedimentation patterns is more difficult to recognize than the effects of tectonics: on the one hand thermal conditions varied considerably during a part of the Last Glacial and, with the exception of the cold and dry Fuquene Stadial (≈22,500 to ≈14,000 years B.P.), may have had little effect on fluvial dynamics. On the other hand, no full glacial/interglacial cycle has been exposed so that the effects of large-cyclic, climatic fluctuations are not known. The glacial/interglacial cycles may lead to the deposition of different sediment types alternating in depth. Such alternations might be detected by the geoelectrical survey. However, as the duration of a glacial or interglacial period lies in the order of several tens of thousands of years and the estimated average sedimentation rate in the Pitalito Basin lies in the order of 0.30 m/1000 years, this would mean alternating layers in the order of 15-20 m thick. Usually, such relatively thin layers can not be detected by the vertical electrical soundings, especially when they are present at greater depth. Interesting, in view of the deeper positioned sediments, are the outcrops of tilted coarse-grained and lignitic basin sediments along the northern strike-slip fault and along the fault in the northwestern part of the basin (Photo 6). They are made up of coarse channel deposits, finer grained overbank deposits and organic flood basin deposits. Their stratigraphical position is unknown as absolute dating was not possible and pollen was absent. It seems however, that organic-rich fluvial sediments might have had a wider distribution in former times and were also present in the western part of the basin.

Somewhere between ≈17,000 and ≈7000 years B.P. infilling processes made place for erosional processes and the northeastern directed flow of the fluvial system changed to a northwestern one. These dramatic changes of the fluvial dynamics are ascribed to allocyclic controls. As these changes are contemporaneous with the Fuquene Stadial it is tempting to ascribe them to climatic control. It is estimated that the decline in temperature in the registered period between ≈60,000 and ≈20,000 years B.P. varied between c. 3° and 8°C. By means of these variations in temperature several relatively warm and cold periods could be distinguished. During such periods the zonal vegetation shifted up and down and obviously also affected higher altitudinal zones in the source area (2800 m altitude) and thus the upstream part of the Guachicos river. Nowadays the source is situated in the Andean forest belt and the river descends to the lower sub-Andean forest belt. During the Last Glacial however, the forest line oscillated between c. 3200 m and c. 2200 m altitude during the relatively warm and cold periods, respectively. This means that the river passed through a variety of ecological environments: during the colder periods the source area was located in an open páramo vegetation belt and the river descended to an Andean forest belt. During warmer periods the river passed through a high-Andean forest and descended to the sub-Andean forest belt. It appeared that the type of sediments did not alter significantly during the transition from a relatively cold to a warmer period and *vice versa*. From this, one may conclude that the architecture of the distal part of the fluvial system remained relatively unchanged between ≈60,000 and ≈20,000 years B.P.

Around 20,000 years B.P. however, the clastic influx in the eastern flood basins increased considerably due to a climatic change. The climate became dry and mean annual temperature around the Pitalito Basin may have decreased considerably, possibly with as much as 8°C. Apparently, some threshold in the climatic conditions was surpassed causing a change in the sediment distribution and the alluvial architecture. The temperature decline reached comparable values as in the previous period before 20,000 years B.P. in which no significant changes in the alluvial architecture in the Pitalito Basin were observed. Therefore, as far as the climatic control is concerned, the amount of mean annual precipitation might be considered dominant for this significant change in the sedimentation patterns. An open and sparse vegetation cover established probably along the entire course of the Guachicos river due to the combination of dryness and temperature decline. Unprotected bare soils are easily eroded and the increased sediment input into the river system in combination with peak floods may lead to frequent overbank flooding. According to studies made around lat. 5°N (Van der Hammen 1974, 1981, 1989; Kuhry 1988), the open páramo vegetation was locally in direct contact with the open (sub)xerophytic vegetation along the western slopes of the Eastern Cordillera.

In fact, these climatic changes only explain the deposition of minerogenic material on top of the peat and do not explain the 90°-turn of the paleocourse nor the halt of basin infill processes. In section 4.2 some arguments are put forward for a climatic cause of these changes. Other phenomena however, like the presence of faults along the modern course of the Guarapas river and tilting of the basin floor towards the north and, possibly, west, point to tectonics as the possible responsible control for these changes. It is possible that comparable significant alterations in the alluvial architecture took place in former times. One may speculate that during such periods sediment deposition may have come to a halt leading to hiatuses in the sediment record. Basin infill may have started again due to for instance increasing rates of subsidence outpacing the rates of incision. But no such hiatuses could be detected with the geophysical methods employed.

The presence of an open vegetation type in and around the Pitalito Basin about 20,000 years B.P. leads to increasing erosion in the surrounding mountainous areas and to the deposition of debris in the form of a c. 2 m thick, poorly-sorted sediment layer on top of organic material. Locally, this sediment layer contains gypsum. Jungerius (1976) also reports an increase of landscape instability for the Middle Magdalena Valley during the Last Glacial (his Phase 6). He describes the presence of gravelly slope deposits and stonelines on top of paleosols and ascribes their genesis to mass-wasting processes during the dry equatorial climate. During this time the reduction in vegetation cover left the soil unprotected against denudation processes. He also stated that the mass-wasting processes acted mainly on superficial soil material which is in agreement with the theory of a sheet wash erosion around the Pitalito Basin. Van der Hammen (1978) reports the deposition of colluvial and windblown 'loessic' sediments in the El Abra corridor in the highplain of Bogotá (c. 2570 m high) during the Fuquene Stadial. During earlier Last Glacial stages a temperature decline is recorded comparable with that of the Fuquene Stadial (Schreve-Brinkman 1978; Δ 8°-10°C) without evidence of a similar decrease of mean annual precipitation. Van Geel & van der Hammen (1973) and Kuhry (1988) report relatively humid conditions before the start of the Fuquene Stadial. All this may indicate that also at higher elevations, just as in the Pitalito Basin, a decrease of the mean annual precipitation is the dominant factor in changes in the sedimentation and erosion patterns.

FURTHER RESEARCH

The Pitalito Basin developed as a result of strike-slip movement along the Garzón-Suaza fault. This fault forms a part of the so-called Eastern Andean Frontal Fault Zone (EAFFZ; Fig. 2A): a megafault system which can be traced from Ecuador, via the Garzón-Suaza fault and the eastern side of the Colombian Eastern Cordillera, to the Merida Andes in Venezuela. The fault zone separates the tectonically active Andean block from the rest of South America (Pennington 1981). The Andean block moves NNE with respect to the South American plate so that transcurrent faults with a dextral sense of movement can be expected along this fault zone. According Pennington's theory (1981) both, the transcurrent faults in the Merida Andes as well as those along the eastern front of the Eastern Cordillera, belong to the same megafault system and became most active since the Pliocene (Mattson 1984). It is interesting therefore, to note that the presence of Late Pliocene pull-apart basins are common elements along the transcurrent faults in the Merida Andes (Schubert 1980) and that several transcurrent faults are recognized in northeastern Colombia (*e.g.* Kellogg 1984; Boinet *et al.* 1985). However, documentation about the presence of transcurrent faults in the southern extension of the Eastern Andean Frontal Fault Zone, more specific around the present study area, is scarce, and especially the documentation about the presence in this area about pull-apart basins. This present study and other recent studies (Diederix, in prep.; Van der Wiel, in prep.) reveal the existence of dextral strike-slip movements along the Garzón-Suaza fault. Therefore, besides the Pitalito Basin, also other Late Cenozoic intramontane and intermontane basins with features inherited by strike-slip movements may be expected in the Eastern Cordillera, for example: the Algeciras Valley, the Neiva, Girardot and Honda Basins (Van Houten 1976) and the highplain of Bogotá (c. 800 m deep; Hooghiemstra 1984). Based on new data the basin infill of the highplain of Bogotá yields an age of c. 1.3 Ma at the 340 m depth level (Hooghiemstra, oral information). Assuming a continuous record this implies an approximate sedimentation rate of ≈ 2.6 cm/100 years over this period. This figure is in the same order as the estimated sedimentation rate in the Pitalito Basin over the last 60,000 years (≈ 2.5-3.0 cm/100 years). Another possible example of a pull-apart basin might be the Magdalena-Cauca-San Jorge Basin in northern Colombia. The basin contains 3,000 to 8,000 meters of mainly Tertiary sediments (Case *et al.* 1984) and is delineated by several transcurrent megafaults (*e.g.* Dolores-Romeral fault, Santa-Marta fault). By means of ^{14}C dates from bore holes drilled to depths of 55 m a sedimentation rate of about 3.8 mm/year over the last 7500 years (HIMAT 1977) is calculated. Such high sedimentation rates are typical for strike-slip basins. However, in this case sedimentation might also be influenced by mechanisms such as Pleistocene erosion and sea-level rise.

Older pull-apart basins may be expected in the western Andean regions along for instance the Dolores-Romeral and the Palestina megafault system. These fault systems exhibit right-lateral displacements starting around the Cretaceous/Tertiary boundary. The Dolores-Romeral fault is still active (Kellogg *et al.* 1985; James 1985) and exercise regional control over the geomorphological features in the Cauca Valley.

A final remark concerns the possible role for the sediments in the Pitalito Basin as a sensible recorder of the climatic history in the low montane regions of tropical South America. Hooghiemstra (1984) elaborated one of the longest terrestrial pollen record of the world: a 357 m long core from the highplain of Bogotá (2570 m altitude). This (Funza) record shows a sequence of vertical shifts of the zonal vegetation belts in time and is interpreted as the effects caused by numerous glacial/interglacial cycles in the high Andean region. The Funza record was recently matched with a new deep-sea core obtained from the Eastern Pacific near Panama (ODP 677; Shackleton, unpublished). Matching combined with K-Ar dating of the intercalated volcanic ashes reveal that the base of the record is about 1.3 million years old at 340 m depth (Hooghiemstra, oral information). Based on these new data the recognized cyclic climatic changes from the Funza record were also tuned with astronomical data. This leads to the recognition of cycles of long and short duration. Actually, a new c. 540 m long core from the same area is elaborated.

The basin infill of the Pitalito Basin may offer partly similar possibilities as the sediments in the highplain of Bogotá, such as:

- a thick sediment pile is present, especially in the northeastern part of the basin;
- in the eastern part of the basin also the sediments at depth are probably made up by fine-grained, fluvio-lacustrine sediments;
- possibly no large hiatuses are present in the record;
- absolute time control is possible as most probably volcanic ashes orginating from the Central Cordillera (*e.g.* Nevado del Huila, Puracé volcano), are deposited in the basin;
- the organic sediments possibly guarantee a good pollen preservation;
- a good documentation about the origin of the sediments is made available by this study;
- the infrastrucure offer a good accessibility to the terrain.

A long pollen record from the Pitalito Basin would not mean a copy of data already known from the highplain of Bogotá. Due to its relatively low elevation (1300 m altitude) the pollen record may reveal the effects of climatic changes on vegetation and on sedimentation patterns in the low montane regions. It also may form a link between the well-documented climatic fluctuations in the high Andean regions and the known climatic fluctuations in the tropical lowlands. A long Pitalito pollen record would gain reliability by matching it with the Funza record and the available deep-sea cores. Furthermore, the effects of the late Andean orogeny (starting ≈5 Ma ago) on vegetation development may be recorded in the pollen assemblages as well. No such long pollen record is known from the low mountainous areas in the tropics.

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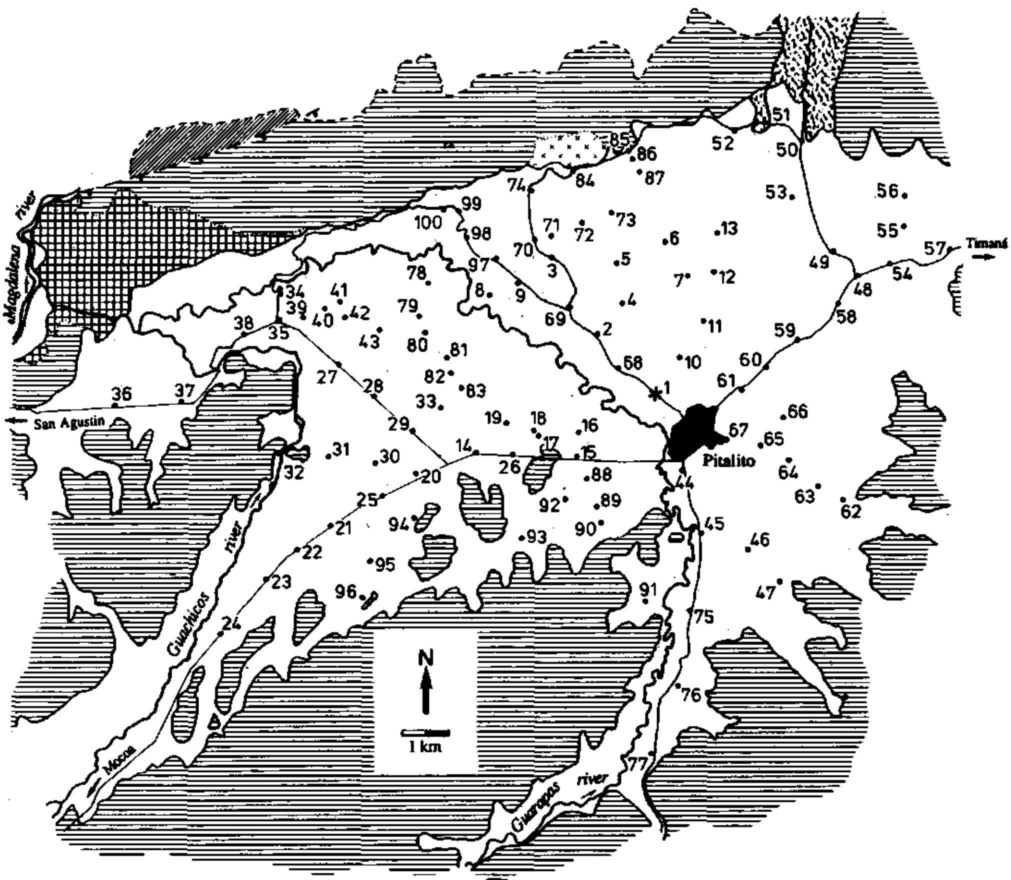
APPENDICES

Appendix I & V (at the back cover)

APPENDIX II

POINT	ALTITUDE (m a.s.l.)	G obs (mgal)	G corr (mgal)	POINT	ALTITUDE (m a.s.l.)	G obs (mgal)	G corr (mgal)
1*	1270	0.00	0.00	51	1313	2.87	11.30
2	1274	-3.23	-2.46	52	1296	6.77	11.21
3	1284	-6.40	-3.27	53	1278	-1.92	-1.39
4	1267	-5.93	-6.59	54	1270	3.36	2.24
5	1278	-7.88	-6.07	55	1268	6.08	4.31
6	1280	-8.73	-6.46	56	1298	1.51	6.47
7	1280	-8.79	-6.63	57	1280	7.67	8.72
8	1285	-3.38	-0.09	58	1267	-2.47	-3.15
9	1276	-4.57	-3.25	59	1268	-3.87	-4.40
10	1273	-3.55	-2.81	60	1272	-3.75	-3.23
11	1268	-7.36	-7.79	61	1275	-2.70	-1.53
12	1275	-8.53	-7.47	62	1285	8.74	12.33
13	1271	-8.81	-8.64	63	1280	4.93	7.28
14	1290	4.35	8.98	64	1275	1.55	2.71
15	1286	-1.20	6.75	65	1275	0.05	1.20
16	1284	-0.64	6.91	66	1278	-1.17	0.67
17	1287	2.84	11.07	67	1276	0.81	2.16
18	1288	2.26	10.78	68	1270	1.17	1.20
19	1287	-1.23	7.09	69	1278	-4.78	-3.00
20	1295	-1.75	8.31	70	1281	-5.97	-3.67
21	1303	-3.07	8.75	71	1283	-6.87	-4.17
22	1313	-3.95	10.16	72	1277	-7.96	-6.66
23	1320	-5.97	9.83	73	1278	-8.61	-6.96
24	1315	-2.28	12.45	74	1300	-3.44	3.14
25	1296	-2.24	7.99	75	1280	4.08	12.85
26	1291	3.78	12.95	76	1280	5.19	14.02
27	1296	4.77	10.72	77	1280	5.90	14.80
28	1293	-2.32	7.67	78	1284	0.81	3.77
29	1291	-2.90	6.73	79	1284	1.14	4.29
30	1290	-2.81	6.54	80	1290	0.20	4.55
31	1293	-2.69	7.23	81	1288	0.07	4.11
32	1282	4.86	12.32	82	1291	0.14	4.81
33	1288	-3.47	5.30	83	1294	0.63	6.07
34	1270	-0.19	4.43	84	1290	0.39	4.72
35	1279	0.30	7.07	85	1306	0.68	8.46
36	1299	0.65	11.89	86	1309	0.71	9.16
37	1289	1.48	10.51	87	1289	1.08	5.05
38	1278	1.99	8.53	88	1275	3.09	4.29
39	1286	1.22	9.48	89	1282	4.58	7.32
40	1279	-1.63	5.03	90	1292	9.54	14.71
41	1282	-2.68	4.64	91	1271	9.27	9.72
42	1285	0.29	8.40	92	1280	5.91	8.24
43	1286	0.88	10.32	93	1285	6.95	10.57
44	1274	5.53	6.48	94	1299	4.09	10.86
45	1279	6.28	8.43	95	1302	3.18	10.61
46	1284	4.24	7.46	96	1320	0.01	11.44
47	1280	6.67	9.14	97	1277	-3.86	-2.34
48	1270	-1.08	-1.28	98	1283	-1.89	0.88
49	1270	-3.47	-4.60	99	1295	0.32	5.83
50	1302	4.69	10.48	100	1321	-3.36	7.89

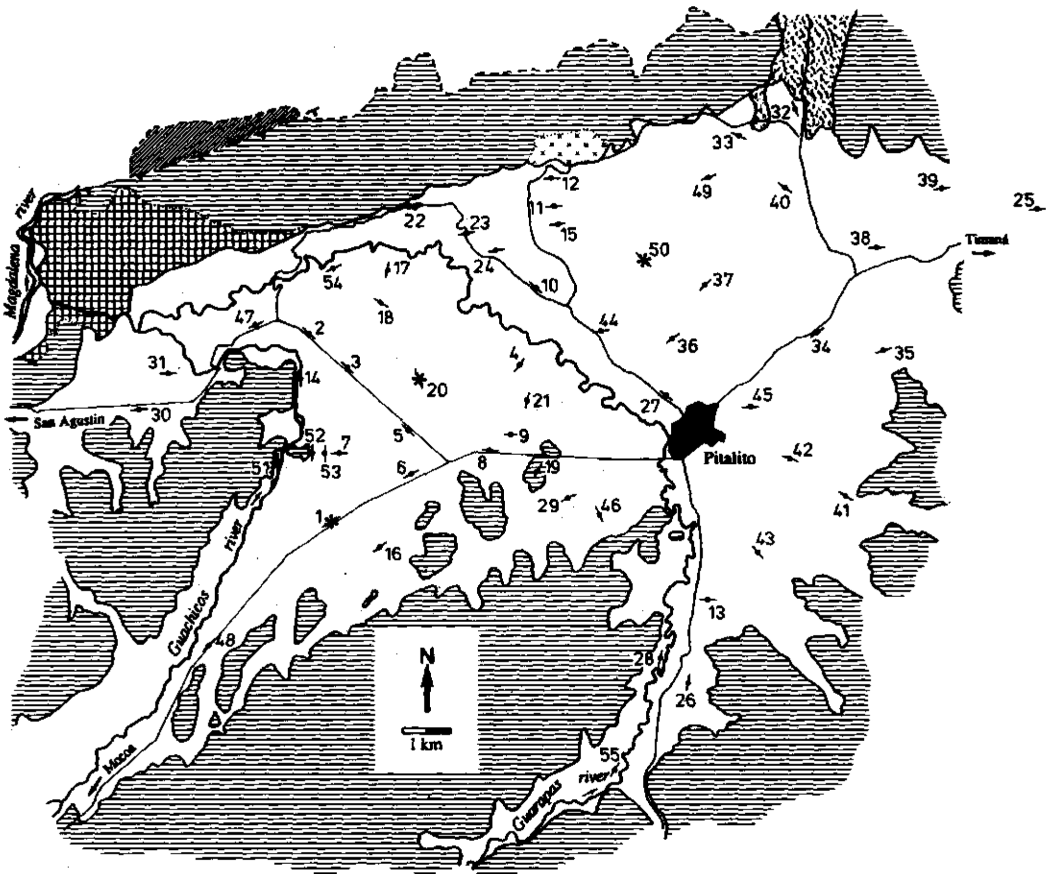
Position of the gravimetrical observation points, their elevation and the observed and measured values of gravity at these points. The basepoint (no. 1) is indicated with *.



Magdalena river
San Agustin
Timaná
Pitalito
Guachicos river
Mocoa
Guarapas river
N
1 km

Appendix III

Position and orientation of the vertical electrical soundings (VES).


Magdalena river
San Agustin
Pitalito
Timaná
Guachicos river
Guarapas river
N
1 km

Appendix IV

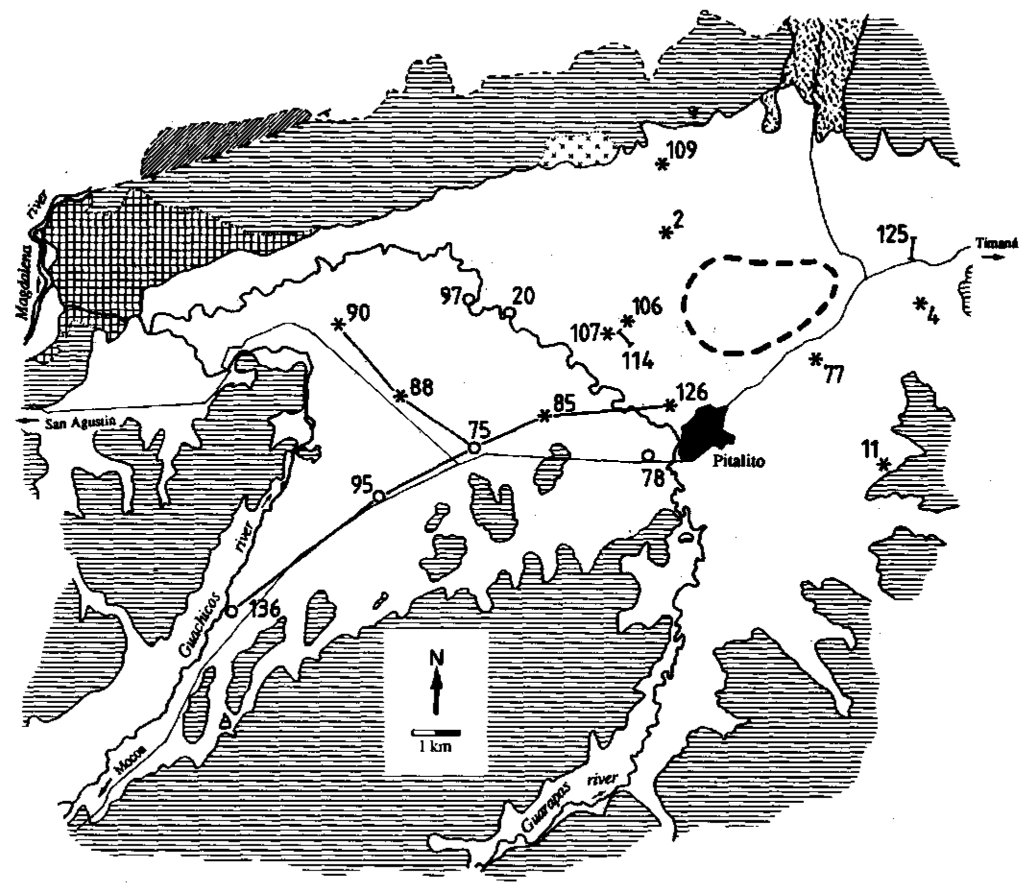
Position of the illustrated lithological sections and borings.

Circles: Outcrops

Stars: Borings

Lines: Sections

Broken solid line: Position of the diatomaceous layer



Magdalena river
San Agustín
Timaná
Pitalito
Guachicos river
Mocoa
Guarapas river
N
1 km
109
2
125
4
97
20
106
107
114
90
88
85
126
77
75
95
78
11
136

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