UNIVERSITY OF CALIFORNIA

Stratigraphy, Tectonics, Paleoclimatology and Paleogeography of Northern Basins of Brazil

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by

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ABSTRACT

Stratigraphy, Tectonics, Paleoclimatology and
Paleogeography of Northern Basins of Brazil

by

Mário Vicente Caputo

Paleozoic basins in northern Brazil contain thick sequences of sedimentary rocks, including diamictites. Because several different geological environments may generate diamictites a study of tectonism, stratigraphy, paleoclimatology and paleogeography was made in order to deduce the processes involved in their origin.

A large part of northern Brazil strata is underlain by metavolcanic and metasedimentary sequences steeply folded and metamorphosed during many tectonic events from about 3,600 to 1,000 m.y. ago. Northeast Brazil was also affected by the Brazilian tectonic cycle from about 700 to 450 m.y. ago. The pre-basin weak zones and resulting trends are responsible for the shape and geometry of 3 huge intracratonic basins developed during Paleozoic time: the Solimões, Amazonas and Parnaíba basins.

The three basins had a similar geologic development during Paleozoic times; from Ordovician to Early Carboniferous time only clastic rocks were deposited, and from Late Carboniferous to Permian time carbonate and evaporites were also laid down.

Tectonism that affected basins is related to uplift and collapse

that preceded the break up of Pangea and subduction activity along the Solimões Basin, in the western side of the South American continent.

Climate has influenced the characteristics of each formation. Paleolatitudes based on paleoclimatic indicators such as tillites, eolian sands, coal, bauxite, red beds, evaporites, limestone, fauna and flora, changed from polar and circumpolar to equatorial during Phanerozoic times. Glaciation was recorded in Ordovician-Silurian, Late Devonian and Early Carboniferous times.

A Late Devonian glaciation left a clear imprint as shown by sedimentary facies. Diamictites with striated, faceted and polished pebbles; rhythmites with dropstones; erratic boulders; striated pavements and deformed sandstones document glacial conditions.

Study of the migration of glacial centers based on the available literature and new data from Brazil shows that they closely follow published paleomagnetic wander data and that there is a close relationship between all Paleozoic glaciations and the Brazilian glaciations. Ice centers moved from northern Africa to southwestern South America from Late Ordovician to Early Silurian time. From Mid-Silurian to early Late Devonian time no record of glaciation is known. In Late Devonian time intermittent glaciation initiated again in central South America and, from Late Devonian to Late Permian time ice centers migrated toward Antarctica across South America and South Africa. The Devonian and Ordovician-Silurian glaciations together with the Permo-Carboniferous glaciations may all have primarily resulted from the shifting position of the Gondwana continent with respect to the South Pole.

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CHAPTER. INTRODUCTION

The main purpose of this study is to examine the climatic effect in the sediments deposited in the northern basins of Brazil. This study focuses on the geology of the Pimenta Bueno, Solimões, Amazonas and Parnaíba basins of northern Brazil in some detail. The geology of the Marajó, Sergipe-Alagoas and Jatobá grabens (Figures 1 and 2), are examined in the context of the entire stratigraphic sequence and paleoclimatic setting. Although paleoclimatic factors are not easy to evaluate, this study examines their effects on the development of the sedimentary rocks present in the region. Glaciation, turbidity currents, tectonism, debris flow and other processes have been regarded as responsible for the generation of diamictites found in northern Brazil. To understand the diamictite genesis an in-depth study of tectonism, paleogeography, and paleoclimatology was necessary in order to deduce which processes were involved in their origin.

In the past, many investigators concluded that tectonism controlled facies patterns and the introduction of coarse clastics into the Amazonas and Parnaíba basins. However, high clastic supply to a basin may be due to tectonic uplift, regression, or

Figure- 1 South America Index Map

climatic change. The highest rates of erosion occur in glacial and periglacial regions, therefore, sediment production may be extremely high in periglacial situations, when fluvial activity flushes debris from glaciated areas following the melting of ice; regressive sequences due to a drop in sea level independent of tectonics are also generally composed of thick, coarse clastic suites. It is therefore unwise to consider that any increase in coarse clastic supply to a sedimentary basin is necessarily related to tectonic uplift in source areas, or tectonic activity in the basin itself. Sedimentological, biological and paleomagnetic data are the tools that enable geologists to attempt paleoclimatic reconstructions. There are many lithic paleoclimatic indicators that can be used to interpret the general position of a sedimentary basin in relation to past latitudes. Lithic indicators of paleoclimatology as well as the description of stratigraphic units will be used for such purpose.

The spatial distribution of glacial rocks through time is important in unraveling world climatic history, and in deducing whether there have been relative motions between continents. Since the general shape and geographic position of the continents is of major importance for paleoclîmatic reconstruction, it is useful to study the effect of paleogeographic reconstruction fur-

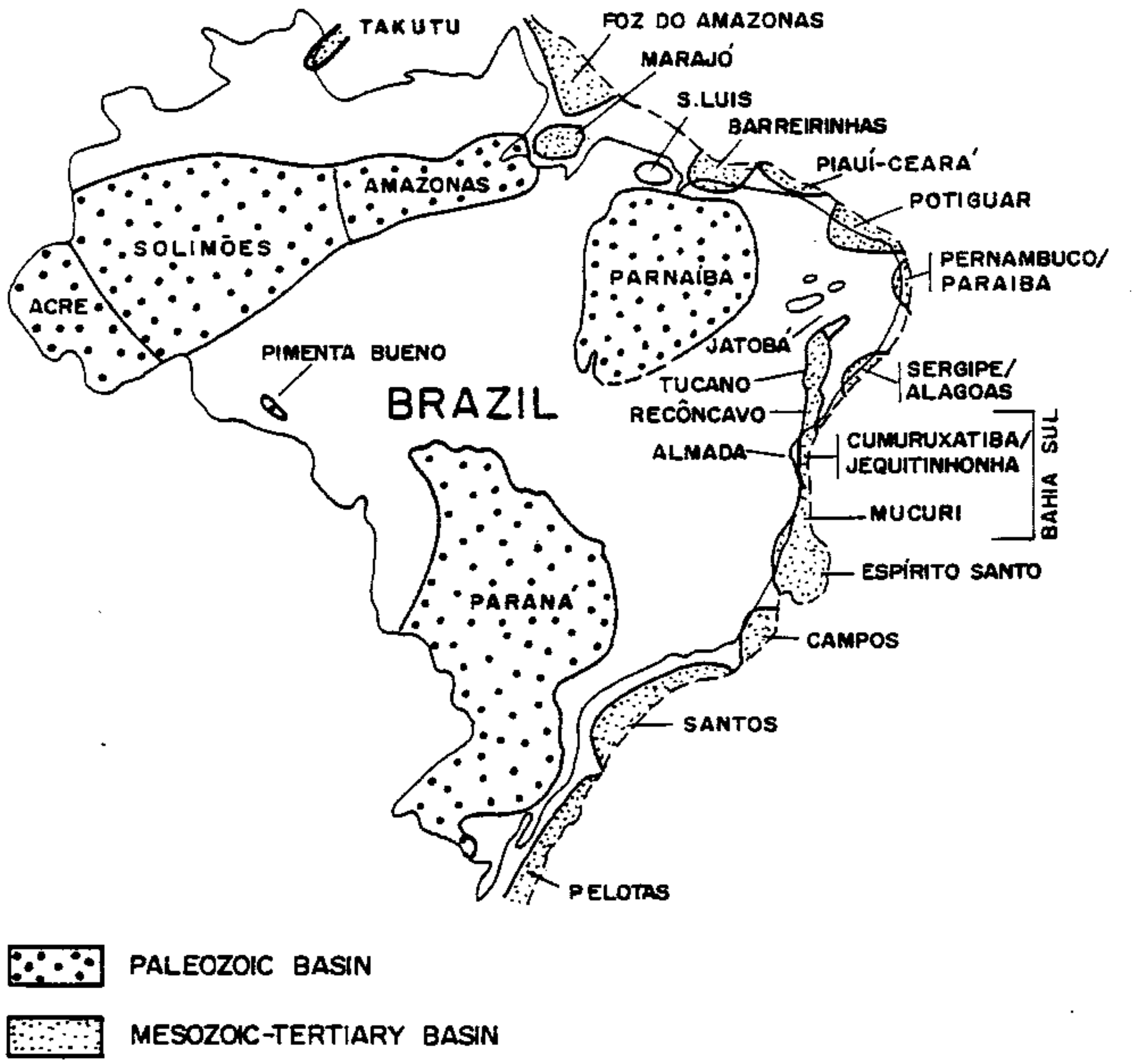


Figure – 2. Brazilian sedimentaryPhanerozoic
basins índex map according to Petrobras

nished by plate tectonics upon paleoclimatological studies (Spjeldnaes, 1981). Intercontinental stratigraphic correlations also offer some evidence for worldwide climatic interpretation.

This study also intends to put an end to the long debate about the existence or non-existence of glaciation in the Late Devonian time in northern Brazil.

Glacigenic beds of other South American and African basins are also discussed according to their age and geographic distribution in order to determine whether or not ice-centers migrated systematically through time in Western Gondwana.

SCOPE AND METHODS OF STUDY

Some Paleozoic sections across the Amazonas and Parnaíba basins were examined and described during many field surveys since 1962, when I started to work for Petrobrás. The study of this broad area has shown that the three sedimentary basins had a similar geological evolution. Then, areas of interest were re-examined in order to develop this study. All sections where previous investigators had mentioned supposed glacial features were examined in order to confirm or dispute the presence of a Devonian glaciation. I have studied the sedimentary pile of each basin surface and sub-surface beds in order to analyze the effect of climate and tectonism on each formation from Early Paleozoic to Recent time.

Descriptions of Pimenta Bueno Basin formations as well as Parnaíba Basin Mesozoic and Cenozoic formations should be credited to previous investigators mentioned below. I was not able to examine these units. They were described here because they exhibit characteristics of paleoclimatic, paleotectonic and paleogeographic importance.

Environmental interpretations were made by me in order to outline the paleogeography of north Brazil and understand the effect of sea-level changes on the sedimentary facies.

Rock and time-correlation of formations were made in order to verify the influence of the tectonism on sedimentary rocks of this huge area. In the second chapter I describe the most common paleoclimatic lithic indicators and in the third chapter I discuss criteria for identification of glacigenic rocks in order to reach paleoclimatic interpretations. In the fourth chapter I discuss the tectonic development of South America from up to Early Paleozoic time based on the available literature in order to understand the tectonic setting of the Paleozoic basins. In the next chapter I discuss the origin and development of the basins. In the following chapters I describe the formations of each basin based on field work, cores, cuts, sidewall samples, electrical logs, minor sedimentary structures of cores of more than 250 wells studied by me at Petrobras paleolab and at the University of Santa Barbara. Petrographic studies of glacigenic sedimentary rocks were also done. A study of tectonism from Ordovician to Recent time was made based on the study of the stratigraphy and the nature, distribution, thickness, pinch out, presence of unconformities, facies, facies-changes, fossil content of sedimentary rocks and type of magmatism.

At the end of this study I discuss the Late Devonian glaciation controversy and present evidence for this glaciation, and different Paleozoic glaciations are reviewed here in order to recognize any relationship between their development.

Finally, the study draws some conclusions with respect to geological consequences on a worldwide basis related to icecap waxing, waning and migration.

CHAPTER 2. CRITERIA FOR PALEOCLIMATIC STUDIES

Paleoclimatology is the study of ancient climates of the earth, where the data come from all sources, but especially employing geological methods. Instruments of modern climatologists are useless to the paleoclimatologist, who must acquire paleoclimatic data by studying rocks, minerals, chemical elements, and fossils of fauna and flora. Climatological data for the recent past are more reliable than those acquired from the remote past. For example, plants and animals of the Tertiary age are more similar to the present ones than those from Paleozoic times. The absence of land plants in Early Paleozoic time probably caused large differences in weathering and sediment transport. This causes problems in environmental and climatic interpretation since sediments look as if they had been formed in areas of high relief (Spjeldnaes, 1981) or colder climates. The effect of living organisms in changing sediment characteristics on the planet has been very important during the course of time. The absence of laterites or bauxites, indicative of wet equatorial climates, in lower Paleozoic or older rocks, may be related to the absence of land plants in those ancient times (Spjeldnaes, 1981).

The high development of gramineous plants in Cenozoic times may have changed weathering rates on land and the sediment supply to the basins. The above example shows that the number of climatic indicators become reduced with time and that paleoclimatologists therefore face considerable limitations in obtaining reliable reconstructions.

Paleoclimatic studies were of fundamental importance in establishing the continental drift theory at he beginning of the twentieth century and in providing pieces of evidence for plate tectonics studies. The position of past polar regions obtained from lithic indicators also can furnish a reference frame for mobilistic reconstructions of continents. Lately, geochemical and paleomagnetic determinations have given great support in paleoclimatic studies. Paleomagnetic data provide a reference frame which can be matched with geological information making climatic reconstrutions more reliable.

All methods of paleoclimatic studies have limitations, but the integration of many sorts of evidence may converge to a single climate for a given region at a given time (Nairn, 1961).

The climatic imprint is clearer in areas of weak tectonism than in areas where there has been much tectonic disruption. Therefore, paleoclimatic investigations are more fruitful in intracratonic basins as is the case in northern Brazilian basins.

Some formations which formed under special climatic conditions, as in a periglacial setting, may be interpreted as a result of relatively high tectonic activity, but a knowledge of the climate during their deposition may reveal their true origin.

The inferences which may be obtained from paleoclimates, concerning the latitudes of the continents in the past, are important for a better interpretation of paleogeography and depositional environments, and also for locating formations of economic value.

The most important indicators of paleoclimates are "climate sensitive sediment types", such as limestone, evaporites, red beds, bauxite, aeolian sandstone, coal, tillite, and faunas and floras. Phosphates, feldspars, micas and other minerals also have paleoclimatic importance.

In addition, the association, presence or absence of certain climatic indicators are significant too in deciphering past climates. For example, absence of limestone, dolomite evaporite and red beds and presence of gray beds on cratonic areas may indicate cold climatic conditions. This study will discuss the most important climatically controlled sediments.

LIMESTONE

Most limestones result from biochemical processes and, to a lesser extent, from chemical processes, and they form in clear and shallow marine and lagoonal environments rich in salts (Wilson, 1975). Figure 3 shows the distribution of recent marine carbonate deposits (after Fairbridge, 1964). In the past, shelf carbonate sedimentation was regarded as being a low-latitude phenomenon.

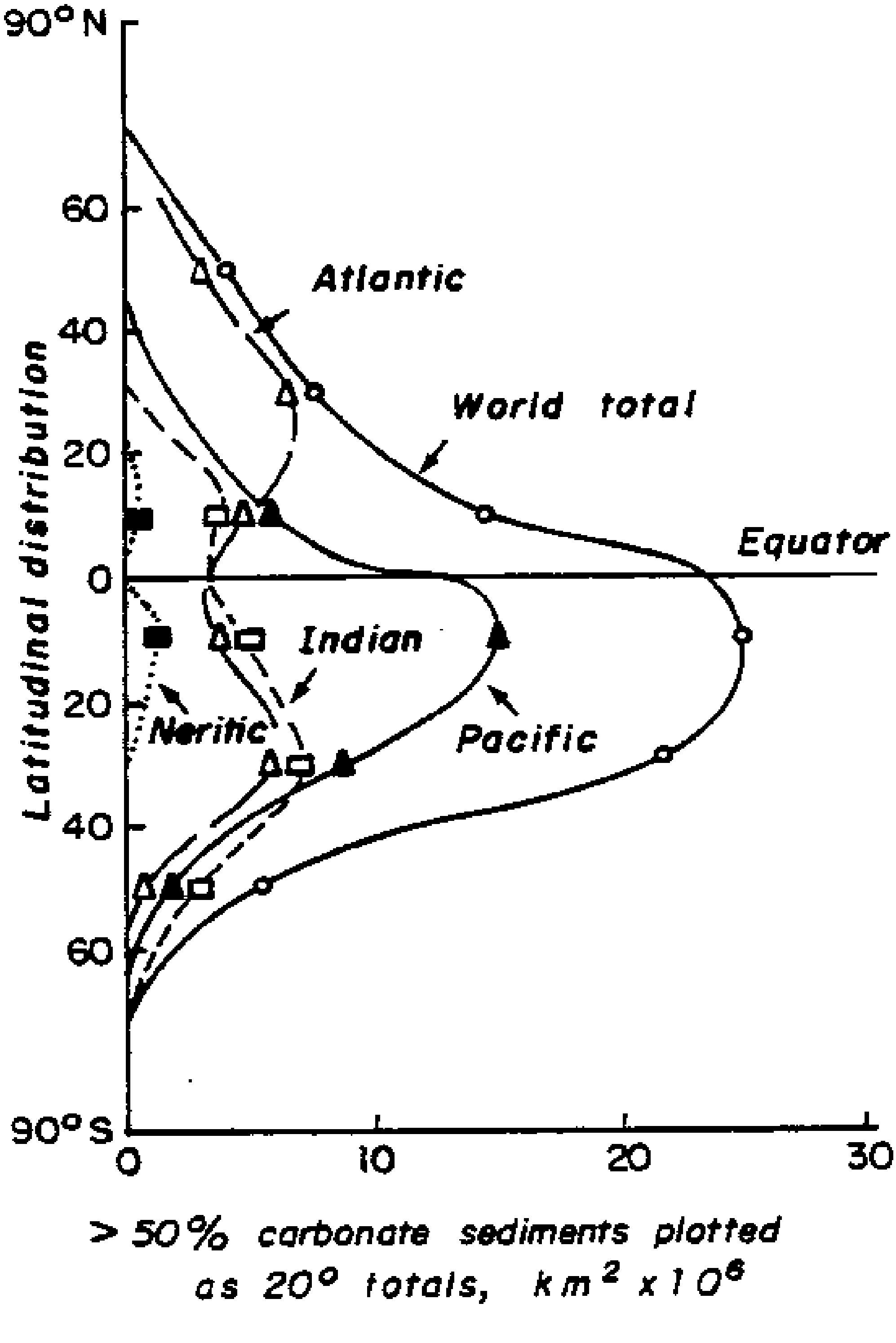


Figure 3. Distribution of recent marine carbonate deposits (after Fairbridge, 1964)

However, although high carbonate productivity occurs in tropical areas today, carbonate sedimentation also takes place in mid and high latitude areas (Figure 3).

The composition of the latter sediments are distinct and can easily be distinguished from tropical carbonates. Lees (1975) pointed out two major associations of skeletal carbonates related to climate: one is a warm-water assemblage (chlorozoan) in which hermatypic corals and calcareous green algae are characteristic components. The other is a temperate-water assemblage (foramol) in which the main components are foraminifera, mollusca, bryozoa, calcareous red algae, and barnacles. In tropical areas, ooids, grapestones, peloids, reef boundstone and lime accumulations are confined to lagoonal and marine environments. In mid-latitudes or temperate-water, lime mud, organically derived, is rapidly dissolved which leaves the carbonate with no micritic constituents (very fine subcrystalline textural components). The different textural composition of carbonates, therefore, allows one to regard them as indicators of warm or temperate climates. It is not always possible to tie warm-water carbonates to low latitudes and temperate and cold-water carbonates to mid and high latitudes because in non-glacial times the warm carbonate belt may expand to about 55°-60º away from the equator.

Moreover, in non-glacial periods the oceanic circulation may outflow from tropical restricted marginal or interior seas, may move toward the poles beneath the normal marine water that moves toward the equator (Brass et al., 1982). The salty bottom water circulation induces higher water temperatures, differences in chemical composition, pH, oxygen and CO_2 content in comparison with present bottom cold water circulation. These differences may have a very important influence upon aquatic fauna and flora.

At times of intense glaciation, warm-water carbonate areas may shrink to about 20-25º of latitude (Figures 4, 95) away from the equator (Fairbridge, 1964). In marginal or interior seas close to the poles, low temperature and low salinity due to high inflow of fresh water is not conducive to the accumulation of carbonates, but in the ocean, cold-water carbonates may be deposited. The presence of dropstones and rock fragments with a wide range of sizes associated with appropriate biota in carbonates is indicative of cold water deposition under arctic and subarctic conditions (Rao, 1981).

Reef-forming hermatypic corals have long been considered to be climatically controlled. Modern reef-forming corals require temperatures above 21ºC and are presently limited to latitudes less than 30ºC. The temperature range they tolerate is 18 to 36ºC with optimum range between 25º and 29ºC (Habicht, 1979). However, the coral belt may be enlarged, reduced or eliminated depending on the prevailing climatic conditions (Figures 4, 95). The use of biological indicators is based on the knowledge of the ecological requirements of recent species. Spjeldnaes (1981) provided a useful study of organisms for paleoclimatic reconstructions.

EVAPORITES

Evaporites are rocks resulting from the precipitation of salts where evaporation exceeds total inflow from run-off and from underground sources. Evaporites may be marine or non-marine in origin. Evaporation excess causes higher surface water salinity in the open ocean, evaporite deposition in restricted marginal or interior seas, and deserts on land.

It is reported that evaporation exceeds precipitation over the oceans between 5º and 35º S and between 15º and 40º N, while precipitation exceeds evaporation between 5º S and 15º N. The earth's thermal equator is displaced northward with respect to the geographic equator due to the extensive ice-cap in Antarctica and small land distribution in the southern hemisphere (Frakes, 1979).

High evaporation rates are intimately related to atmospheric circulation in the north and south subtropical zones of high pressure. Here, part of the dense cold and dry air sinks, whereas part of it moves toward the equator (easterly trade winds), and

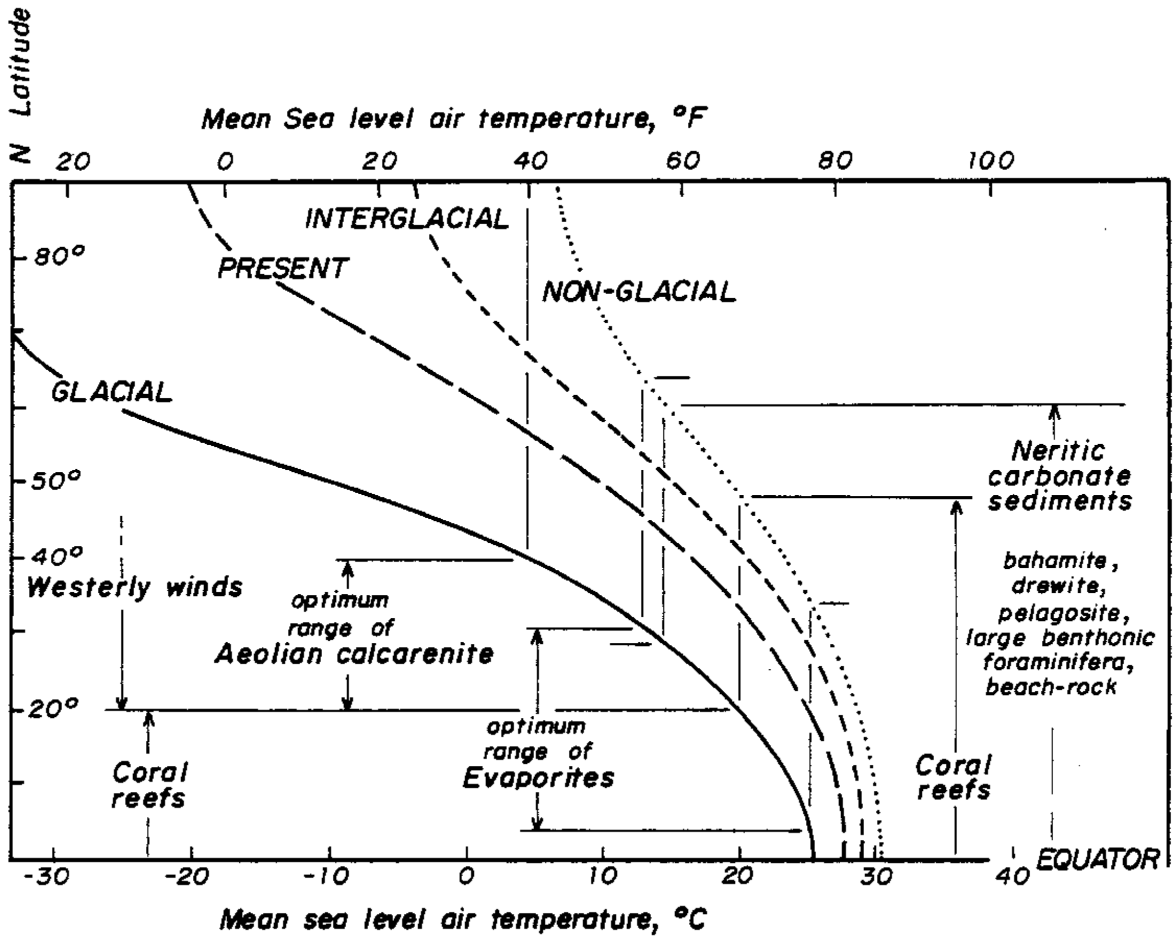


Fig. 4 Surface mean temperatures for the present, glacial, nonglacial, and interglacial periods (after Fairbridge 1964)

yet part moves into high latitudes (westerly trade winds}. The dry air picks up moisture and warmth on its way toward the equator (trade winds) where it expands, rises and then cools delivering heavy equatorial rainfall. Part of the air which rises goes directly to the polar region, and part goes to the tropical zones of high pressure, thus closing a partial cycle. There are also two belts of low pressure at about 60º latitude where the returning cold air from the poles (polar front) forces up the warmer air from the subtropical zones of high pressure causing high precipitation. The front itself moves as these two air masses push back and forth.

This is the atmospheric circulation picture of today, but when there is an extensive monopolar or bipolar glaciation, climatic belts may change. When a worldwide refrigeration occurs, both evaporation and precipitation should be reduced in equatorial areas where arid and semi-arid climates may predominate (Fairbridge, 1972). In nonglacial periods, when a higher average temperature prevails, climatie belts are very difficult to assess, but high evaporation and precipitation zones should be wider or another belt of lower pressure could be established near the free-ice.polar.areas.

The studies by Gordon (1975) and Drewry et al. (1974) of present and past evaporite distribution show bimodal frequency with few evaporites in the assumed paleoequatorial zone and two belts of abundant evaporites between 10 and 50 degrees of the paleo-equator. This equatorial evaporite minimum may help to establish paleoequatorial zones in worldwide paleoclimatic studies. The distribution of ancient evaporites fits better the model of plate tectonics with lateral shifts of continents. The assumption by Meyerhoff and Meyerhoff (1972), based on ancient evaporites, that continents have not moved in the past makes it difficult to interpret ancient evaporite belts (Gordon, 1975; Zharkov, 1981).

Gypsum precipitates in the sea where the brine concentration ranges between 2 and 12 times its normal salinity and halite starts to form after 91.7% of the original seawater is evaporated (Borchert and Muir, 1964 in Habicht, 1979). Since 1.382 g of $CaSO_4$ and 29.696 g of NaCl result from evaporation of one liter of seawater, one should normally find twenty times more NaCl than $CaSO_4$ in the evaporite deposits. The observed deficit of NaCl in most evaporites may be due to the outflow of denser salty currents from the basin of precipitation to open oceans, as occurs in the Mediterranean Sea today. When the sea entrance is very narrow or when total basin isolation takes place, NaCl predominates over $CaSO_4$. The volume of the evaporite deposits cannot indicate climatic severity or the amount of evaporated water due to the outflow of bottom denser brines. The volume of anhydrite may furnish a better idea of the minimum amount of evaporation than halite does. The presence of evaporites in many basins at the same time may indicate that a large amount of heat was transferred from the equator to the poles as vapor.

RED BEDS

A red bed is a sandstone, siltstone or mudstone made of detrital grains set in a reddish-brown mud matrix or cemented by precipitated reddish-brown ferric oxide (Van Houten, 1964). Continental red beds are found in a wide range of sedimentary environments, such as alluvial fans, rivers, flood plains, deserts, lakes and deltas (Turner, 1980). Conditions for red bed formation are a warm or hot climate under semiarid, arid or humid settings It is difficult to distinguish between red beds formed in humid dry climates, but most of them were deposited at relatively low latitudes within 40 degrees of the equator. In order to distinguish between dry and moist climates, Walter (1974) stressed that dry-climate red beds are associated with aeolian sands, desert fluvial sediments, and evaporites formed in playa lakes and inland sabkhas, while moist-climate red beds are inter-bedded with coal strata. Some red-beds do not show direct evi-dence of the nature of the climate, in the depositional environment, but suggest sedimentation in hot conditions. In special cases, middle latitude red beds are found, Carboniferous tillites from Bolivian basins are red and may result from the alteration of a thin loose unstable rock veneer left in shield areas during the previous Late Devonian glaciation.

Red beds are especially suitable for paleomagnetic studies and many paleomagnetic surveys have been carried out in them. Turner (1980), however, pointed out that due to diagenesis the results my be misleading. He recognized three general kinds of magnetization: type A magnetizations include essentially a single component of magnetization and may appear with zones of either normal or reversed polarity; type B magnetizations are multicomponent and were recorded over long time ijntervals, encompassing at least one geomagnetic field reversal; type C magnetizations are those in which the original magnetization was replaced, therefore, their present magnetizations have no relation to the magnetic field during the deposition of the red beds. Each type reflects increasing red bed diagenesis.

BAUXITE

The aluminous (bauxite) and/or ferruginous laterites form under pedogenic conditions controlled by a well-defined climate, with relatively high atmospheric temperature and sufficient rainfall, occurring in equatorial and tropical areas. Virtually all rocks which contain aluminous silicate may be converted into bauxite. Porous texture of the mother rock, plateau topography and, the existence of a variable water-table favoring infiltration drainage through the sediments, with springs near the foot of the plateau slopes, are factors which lead to the destruction of aluminous silicates by hydration and hydrolysis and liberation of the alkaline and alkaline earth ions and of all or part of the silica (Nicholas and Bildgen, 1979)

In depressions, destruction of silicates also occurs, but inadequate drainage does not permit their removal, while in rugged topography rock decomposition is reduced by lack of infiltration through the sediments. Organic acids from luxuriant vegetation may play an important role, but one known pre-Devonian deposit exists, the Precambrian(?) bauxite of Bokson, U.S.S.R, mentioned by Nicholas and Bildgen (1979), which may challenge the importance of vegetation cover in bauxite genesis.

Nicholas and Bildgen (1979) have demonstrated that bauxite distribution belts are very useful in reconstructing continental positions on Earth. According to them the distribution of ancient bauxite deposits is best understood in the context of lateral shifting of the continents from Devonian time on, and when continents are appropriately displaced the ancient bauxite deposits show notable fit to the present-day model. However, a problem exists with bauxite deposits; their age is very difficult to determine because they represent a type of rock alteration that takes place in substrates of any age. The time of bauxitization needs to be limited by the ages of the underlying and overlying deposits and other paleogeographic methods.

Nicholas and Bildgen (1979) stressed that bauxite deposits must have a close relationship to orogenic belts and volcanism, and bauxite mother rocks must belong to the group of under-saturated volcanic rocks (basalt, andesite, nepheline syenite, etc.). In my opinion tectonics is not a condition in bauxite genesis because Tertiary Amazonian bauxites are far in time and place from any mobile belt, and almost concomitant volcanism is not necessary either because there are no Tertiary or Late Cretaceous volcanic extrusions in the area.

Clays resulting from the erosion of the Guyana and Guaporé shields were converted into bauxite on the Alter do Chão and mother rock may have been very diverse because the Guyana and Guaporé shield source areas are composed of rocks with much petrographic variety. The huge amount and wide areas of bauxite deposits favor the idea that kaolinitic clays, siltstone and argillaceous sandstone strata of Alter do Chão were weathered "in situ" after deposition under the geological conditions already explained.

COAL

Coal deposits resulting from land plants accumulate in swampy areas covered with very shallow water or peat bogs with a water level which rises into the vegetation cover. Flat topography, abundant rainfall and poor drainage are the most important factors in coal accumulation. Plant growth must be faster than plant decay. Any plant remains may be converted into coal, under any moist climate such as tropical, temperate, boreal or frigid, except in polar extremes where the moisture is nearly always frozen (Schopf, 1973). (Six months of darkness at the poles may not be so restrictive for plant life in polar areas, but the frozen water inhibits plant development). It is possible for plants to grow at the poles under higher temperatures than those that exist there at present. Therefore, polar light is adequate to sustain plant life on the poles if the temperature is not below freezing point all year round. If plant-bearing deposits are found above glacigenic sedimentary deposits it means that the climate has ameliorated in the areas up to a level that plants can develop. At present, in Alaska, in the wake of glacial retreat, plants are growing nearly everywhere.

Coal should be most abundant in the wet tropical equatorial belt and in the two high latitude humid belts, but in semi-arid regions some coal can occur due to a high water-table, so it is very difficult to tie coal deposits to the latitudes where they were deposited. Regions of most extensive modern peat deposition are in fluvial, deltaic and coastal depositional environments. In fluvial environments peat deposition takes place in floodplains and oxbow lakes (abandoned meanders). In delta plain environments peat deposition occurs in interdistributary bays, lakes and in abandoned distributary arms. In coastal environments peat deposition is present in marsh fringes.

Although coal deposits have little value as paleoclimatic indicators, floras which make up coals are good indicators because plant distribution is primarily controlled by climate, which varies with latitude (Schopf, 1973). The major vegetation zones and the climate under which they grow show a symmetrical arrangement since successive cooler zones occur at higher latitudes on opposite sides of tropical belts (Axelrod, 1963).

Plant characteristics indicative of tropical, subtropical and cold climates have been known for a long time (White, 1913 in Schopf, 1973). Cold and frigid climate plants are characterized by growth rings. Growth rings from very cold area trees are very thin while cold temperate area trees´ growth rings are thicker due to larger and faster wood growth in milder climates. Schopf (1973) regarded thick growth rings found in wood incorporated in Permian tillites from Antarctica (85º S) as an indication of a non-polar position of that continent in Permian time. In my opinion, in interglacial or non-glacial time intervals, when a general climatic amelioration takes place, vegetation may advance in the wake of the ice retreat as it presently does in the Arctic region, and may develop on glacial sediments.

AEOLIAN SANDS

At present about 36% of the earth's total land area is located in semi-arid, arid and extremely arid regions. In the past, sediments deposited in such regions could have formed a significant portion of the sedimentary record (Opdyke, 1961). However, geologists have failed, in many cases, to recognize aeolian derived sediments in the geologic column. The widest deserts are located along the Tropic of Cancer and the Tropic of Capricorn, but they may occur as far north as 45º in Asia or at the equator, where the Andean belt causes a rain-shadow on the western South American coast.

Alluvial fans, wadis (ephemeral desert rivers), playa deposits, and continental and marine evaporites are normally associated with aeolian sands. Additional minor features are the presence of ventifacts, pebbles with brown to black coatings called "desert varnish" and caliche accumulations. Scattered "milled-seed" grains may impose a bimodal character to the aeolian sands which normally are well sorted, very fine, and well rounded with high sphericity. The major criterion for identification of desert sand is the large-scale and relatively high angle of cross-stratification sets (Walker and Middleton, 1979). Associated interdune deposits composed of a wide variety of sediment types and structures reflecting deposition under wet, damp and dry conditions are good indicators of sand seas (Kocurek, 1981). On the northern Brazilian coast, north of the Parnaíba Basin, a modern coastal dune field presents a large number of interdune lakes indicating a high water table in the area.

The average dip direction of he cross-strata of paleodunes gives the sense of sand transport and consequently the direction of paleowind responsible for the sand deposition (Bigarella, 1973a). The interpretation of the general paleowind circulation pattern may indicate an area north or south in relation to the equator. The area may be located in a region where westerly or easterly wind patterns predominate or between them (wheel-round wind area), although the wind belts' position and width may have changed during glacial and non-glacial periods. For example, in glacial periods the wet equatorial zone is reduced and the trade wind belts advance towards the equator (Fairbridge, 1972), eliminating a large extent of tropical forests. In Pleistocene time a large width of the Amazonian forest was extinguished due to the effects of glaciation.

TILLITE

Tillite is a genetic term for a rock formed from lithified till which is a nonstratified, non- or poorly sorted sediment with particles ranging from clay- to boulder-size, carried or deposited by a glacier. The glacial environment is characterized by dominance of huge ice masses as a geologic agent, but the environment is varied, so there are many different types of deposits.

Continental glaciers cover mountains, plateaus, and valleys, concealing the entire country except for the highest steep peaks. About 10% of the earth's surface is presently covered by glacial ice, but during the Pleistocene glaciation, maximum glacial extent was about 30% (Flint, 1971) with a probable ice volume 3 times larger than that at present. Today the glacial environment is restricted to areas around the north and south polar regions and high mountains in lower latitudes above the snow line. Near the equator, valley, glaciers occur at elevations of about 4,750 m or higher. In Borneo, at about 6º latitude and 4,102 m elevation, Pleistocene till deposits extended to 2,700 m, related deposits to about 1800 m and outwash deposits to the coast (Meyerhoff and Teichert, 1971).

According to Flint (1961) a clear distinction should be made between extensive ice-sheets and smaller glaciers discontinuously

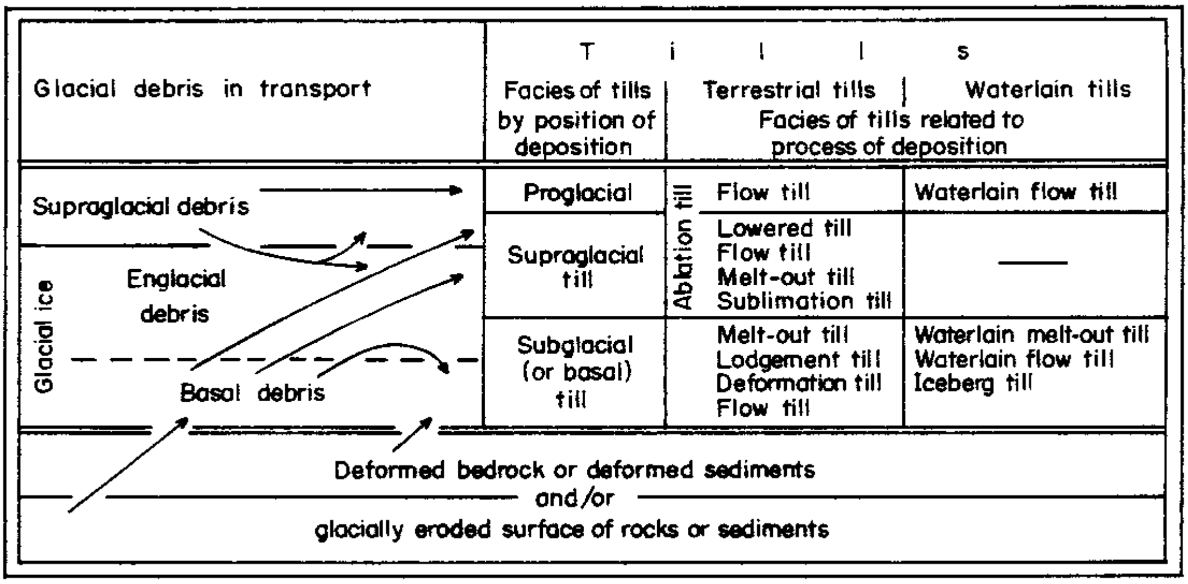


Figure 5 Provisional genetic classification of tills according to their position and process of deposition and the relation of tills to glacial debris in transport and the substratum (after Hambrey and Harland, 1981.

occupying highland areas. Around highland glaciers, foothill areas may experience continual warm temperatures. If glacial sediments reach the sea, they accumulate in fjords, and sedimentation covering extensive areas is unlikely.

Glaciers form when more snow falls during the winter than melts and evaporates in the summer. They move both downslope and upslope because the force of gravity is transmitted over the whole ice body; the glacier actually spreads out slowly. In the past many tillites were identified only on the basis of rock texture, but since the 1950´s and 1960´s geologists have become aware that the actual supposed tillites may not indicate a glacial origin, because similar rocks could be formed by debris flow, proximal turbidity currents and mass flow movement in general (Crowell, 1957). Under such circumstances, descriptive terms were created such as diamictite (Flint and others, 1960b), mixtite (Schemerhorn, 1966) and tilloid (Pettijohn, 1957). These terms have no genetic implication, but describe rocks with textural characteristics similar to those of tillites. Therefore, sedimentary units composed of rocks considered to be tillite on the basis of their texture had to be re-studied, mainly because they did not have any significance in documenting climatic history and related paleogeography. The figure 5 shows a provisional genetic classification of tillites and related sediments, illustrating the relationship of debris in transport in the glacier and the process of deposition prepared by Hambrey and Harland (1981).

Investigators such as Harlan and others (1966), Crowell (1957, 1964, 1983), Schwarzbach (1963, 1964), Schemerhorn (1966), Frakes and Crowell (1967), Spencer (1971), Hambrey and Harland (1979, 1981) have established criteria for identifying a glacial origin. Ancient glaciations are recognized by identifying both glacial geomorphic forms and associated sedimentary features related to the glacial activity. Since glaciation has a worldwide influence in the geologic process, indirect evidence found away from the glaciated area may also aid in establishing glacial activity at a given time.

Tillites record the advance and retreat of ice-sheets that result from climatic change in high and moderate latitudes. They do not define a distinct and permanent and climatic zone and their lateral extent depends on the intensity of the glaciation with which they are associated (Drewry and others, 1974). Therefore, it is desirable to determine its full extent in order to infer the severity of climate at a given time.

CHAPTER 3. CRITERIA FOR THE IDENTIFICATION OF GLACIGENIC DEPOSITS

Till-like deposits can form under diverse circumstances. However, their field relationships may allow discrimination between various origins. Crowell (1964) explained the possible origins for till-like deposits, as summarized below:

(1) deposition by glaciers as till;

(2) downslope movement or slumping in marine and nonmarine environments;

(3) debris-flows, both subaqueous and subaerial;

(4) mixing and down-current movement caused by the impact of strong turbidity currents;

(5) milling and mixing within and beneath giant slide blocks which grade continuously in size up to immense thrust plates;

(6) volcanic mud flows (lahars);

(7) talus debris along escarpments, both subaqueous and subaerial;

(8) selective weathering or alteration of conglomerate in place; and it may be added;

(9) tectonic melanges from subduction zones, diapirs, and Broad fault zones.

As seen above, sediments covering the entire gravel- to clay-size range are of importance as climatic, tectonic and sedimentary environmental indicators (Schermerhorn, 1966). Because tillites display a texture that is common in rocks from diverse environments, other features must be observed in order to establish the rocks´ glacial nature. In many cases, large areas are covered by diamictites (or mixtites) but, because critical features for identification are not observed, the glacial origin of such sediments cannot be demonstrated. Before the nature of the beds is known it is therefore better to make use of descriptive terms rather than genetic ones. Two descriptive terms are currently widely accepted.

The first term, diamictite, was proposed by Flint and others (1960a,b) for an indurated rock formed from an essentially non-sorted, non-calcareous, terrigenous deposit composed of sand and/or larger particles immersed in a muddy matrix. They did not specify the amount of muddy matrix in diamictite, meaning that graywackes may be included and sandy rocks or sandy tillites free of a muddy matrix are excluded. The second term, mixtite, was coined by Schermerhorn (1966) for mixed coarse- to fine-sediments with a wide range of grain sizes, and characterized by a sparse to subordinate coarser fraction composed of clasts of all sizes and shapes, immersed in a matrix made up of varying proportions of sand, silt and clay. The term diamictite has been more widely used although mixtite is more inclusive in describing non-sorted clastic sedimentary rocks that are made up of a wide range of fragment sizes, and generated in variable environments.

The term diamictite is widely used in Brazil and has priority, therefore it is here used for encompassing those rocks that are made up of a wide range of particle sizes.

The genetic term till was first defined by Geikie (1863) to describe a "stiff clay full of stones varying in size up to boulders produced by abrasion and carried on by the ice-sheet as it moved over the land". Till was also defined by Francis (1975) as a sediment deposited by or from glacial ice without the intervention of running water. Penck (1906 in Du Toit, 1953) was the first geologist who used the word tillite to describe lithified till or boulder clay of a glacial origin, occurring in the Karoo Basin, South Africa. Since tillite is a genetic term it should only be used when the glacial nature of the formation under discussion has been established

According to Hambrey and Harland (1981) the term tillite may include the lithified equivalents of:

(1) Terrestrially deposited till of various types;

(2) Till deposited by a grounded ice-sheet or glacier in a marine environment;

(3) Material deposited by a floating ice-sheet or glacier;

(4) Material resulting from deposition by floating icebergs into marine sediments of other sources (excluded by many

authors);

(5) Material deposited from ice floes and obtained from beaches or rivers (excluded by many authors);

(6) Material deposited by stranded icebergs;

(7) Material deposited by ice in marine environment and subjected to large-cale mass movements.

Tillite deposits are normally non- to poorly sorted, with a great range of fragment sizes. The clasts may be rounded, sub-rounded, angular or faceted or “flatiron shaped”, and some may carry striations. Tillites from a continental glaciation should cover a great areal extent and contain far-travelled clasts. A charac-teristically high variety of clast types is consistent with broad source areas. Some clasts may be of fragile stones, such as shale pieces (Boulton and Deynoux, 1981). The matrix may contain any proportion of sand-, silt- and clay-particles. The particles may be picked up directly from the substrate or may result from rock material mechanically reduced to sand-, silt- and clay-sizes.

This material is originally made up of unstable minerals (rock flour), so fine grained that it is hard to identify under the petrographic microscope.

If the glacier substrate consists of clay, the corresponding tillite may contain much muddy matrix, whereas if the glacier sub-strate consists of sand-sized grains, the resulting tillite may have a sandy matrix as was observed in Ordovician tillites from northern Africa (Fairbridge, 1969, 1970a,b). Moreover, when ice-sheets reach the shoreface and offshore areas rich in muddy sediments, which are exposed because of sea water withdrawal due to ice build-up on land, tillites may change from having a sandy to a muddy facies. A subsequent fast transgression due to ice melting may protect the glacial rocks in the basin area from erosion, while tillites exposed in shield high regions may be removed by weathering agents.

Hambrey and Harland (1981) developed a table which includes all kinds of recognizable tillites. This table is presented in Figure 6.

Although this classification may be useful, it is very hard to identify the different glacial subenvironments in the geological record.

In this study, the size of the area (three huge sedimentary basins and large grabens) and the nature of the investigation allows one to study only some critical features. These may enable the recognition of a glacial origin for some formations and to establish a Devonian ice age. Limited data and exposures do not permit discrimination between minor glacial subenvironments.

DIRECT EVIDENCE FOR GLACIATION

Direct evidence for glaciation may be based on the presence of boulder pavements, striations and related features, periglacial

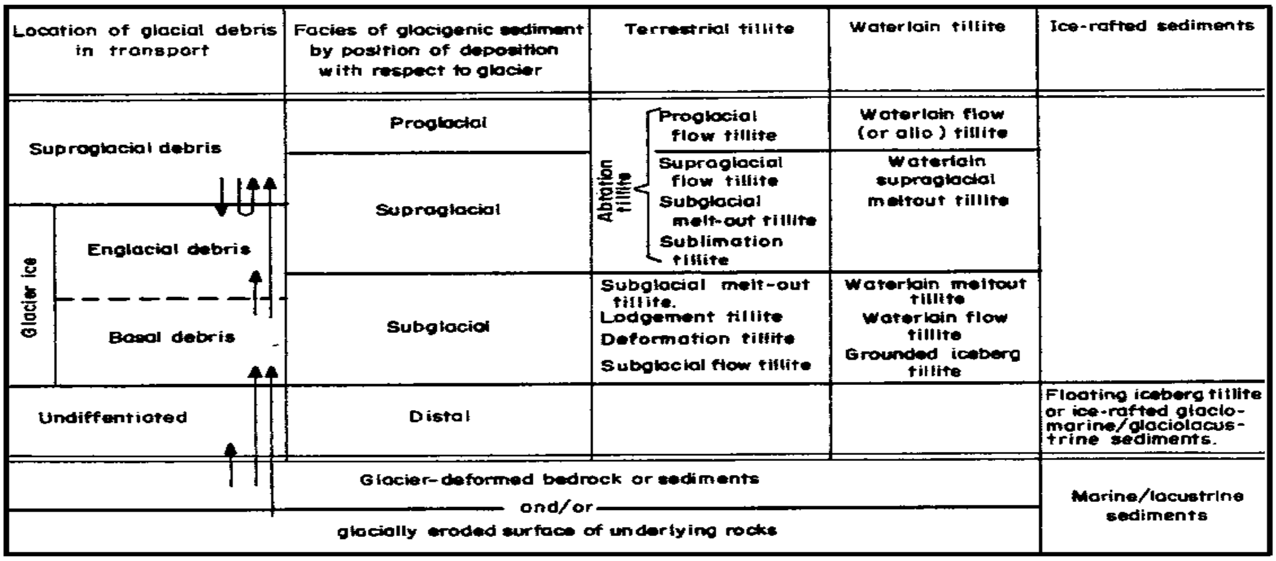


Figure 6 Provisional genetic classification of tillites and related sediments, illustrating the relationship of debris in transport in the glacier and the process of deposition. (after Hambrey, and Harland, 1981)

glaciolacustrine and glaciomarine deposits and loess, patterned ground and other features. A discussion of certain of these critical features follows:

STRATA AND BOULDER PAVEMENT

In some places older rocks or strata and also boulder beds previously deposited in front of a glacier may constitute the ice-sheet substrate, so during a new glacial advance, or readvance, the top of the boulder bed is levelled by the ice motion, resulting in striated boulder pavements (Harland and others, 1966).

STRIATIONS AND RELATED FEATURES

Glaciers commonly scour into bedrock beneath them, leaving striations upon it which are due to attrition between clasts included in the ice body and the glacier rock substrate. In general, where the bedrock consists of soft material, grooves and striations have high relief, but where the bedrock consists of hard rock, striations are less visible and grooves less common. In some places, low hills are abraded by glacial action resulting in features known as "roches moutonnées". Chatter marks and crescentic gouges are common features on striated pavements. Striated pavements may also result from a tectonic or mass-flow origin, but the pattern is more uniform and a layer of gouge-like material is probably present. Glacially striated pavements if they are overlain by diamictite deposits, a tectonic origin is excluded and if the mixtite deposits contain striated clasts, and show great extent, a glacial origin is likely. Moreover, if the substrate is horizontal over a large area, tectonic or debris-flow origin is unlikely. Striated pavements and "roches moutonnées" overlain by tillites with striated clasts furnish irrefutable evidence for terrestrial glaciation (Crowell, 1983).

PERIGLACIAL DEPOSITS

The peripheral zone to the glacial ice, where mean annual temperatures are below 0º C, is called the periglacial zone. The term periglacial was introduced by Lozinsky (1908 in Washburn, 1979) to designate the climate and the climatically controlled features adjacent to the Pleistocene ice-sheets. Subsequently, the term was extended to areas adjacent to older ice-sheets.

Large amounts of sediment are commonly carried beyond the ice by rivers and wind. Sediments are laid down in wide plains in coalescent fans. In this environment braided streams predominate due to high loads, seasonally variable discharges, steep slopes and coarse sediment (Fahnestock, 1963). Glaciofluvial outwash deposits lose their glacial imprint, and striations on clasts are rarely preserved. These sediments are generally composed of gravels and sands, and may be poorly- to moderately-sorted or even well-sorted (Francis, 1975). In North America, Quaternary outwash deposits extend from the outer limit of related ice-sheets to the Gulf of Mexico, more than 1,000 km downstream along the Mississippi River (Flint, 1975). Similarly, in the Soviet Union, Quaternary periglacial outwash deposits extend comparable distances via the Volga River to the Caspian Sea (Flint, 1975).

Flint (1975) also pointed out that the presence of fresh plagioclase feldspar and biotite, pebble- and cobble-sized erratics with glaciated shapes, and the high proportion of angular sand-sized grains of quartz, as well as grains with specific surface microtextures, are all indicative of glaciation. As enlarged upon below, in the Parnaíba Basin, Kegel (1953) described faceted stones and large amounts of angular sand-sized quartz and feldspar grains in conglomerate and sandstone beds of the Lower Serra Grande Formation (Group). These sediment characteristics are consistent with a glaciofluvial origin for part of the Serra Grande Group of the Parnaíba Basin.

GLACIOLACUSTRINE DEPOSITS

Lakes are frequently developed in periglacial areas, for many reasons: damming of river courses by glacial bodies or debris, reversal of regional slope due to the peripheral isostatic depression caused by ice loading and development of irregular topography by glacially deposited or eroded landforms (Flint, 1971). Deltaic deposits may be formed at lake margins and varved sediments are laid down in the centers of glacial lakes. A varve is a type of rhythmic sedimentary deposit which is laid down mechanically under the influence of seasonal changes during one year. It consists of couplets: a lower coarse member made up of silt and very fine sand and an upper member made chiefly of clay. The clay member is usually dark and rich in organic content. This makes a characteristic contrast between light and dark bands. The thickness of each varve ranges from less than 1 cm to, rarely, 75 cm (Flint, 1975). The coarser member itself is composed of many graded laminae. The fine member is normally graded but the contact with the underlying member is commonly sharp. Dropstones, ranging from sand-size upwards may occur in the section. According to Edwards (1980), varves are a consequence of two mechanisms:

(1) Sediment-laden stream water is denser than lake water, so the coarse sediment is transported as a density underflow (Gustavson, 1975). The occasional development of ripple cross-lamination in the coarser layer indicates deposition by a bottom traction current. These may or may not be related to turbidity currents.

(2) Strong seasonal variations in run-off and winter ice

cover on lakes lead to the deposition of coarse sediment during the summer and fine sediments from suspension during the winter.

Varves are formed in fresh water lakes; in general, clay flocculation in salt water prevents the formation of marine varves.

DROPSTONES

The presence of dropstones in varves, if striated and faceted, indicates glacial origin without the need for microscopic examinatión (Flint, 1975). Stones falling from an ice raft will puncture the laminae beneath and even splash up fragments of the substrate as they penetrate (Crowell, 1983). Later sediment then drapes over the top of the stones. In the sea, glacial debris can be rafted by an ice shelf, sea ice or by icebergs.

Three distinct types of sediments may develop around glaciers which reach the sea (Boltunov, 1970; Edwards, 1980). Beneath the extremity of the glacier a large amount of debris (unstratified till) is laid down on the sea floor without reworking. Beyond this, a region develops with little coarse material dropped under the glacier or carried away by icebergs (stratified till). In an outer zone normal marine sedimentation occurs with sporadic dropstone released from floating ice. In a marine environment the presence of dropstones indicates the existence of glaciation, although the ice centers may be far away, whereas in a lacustrine environment dropstones indicate that the margins of ice-sheets were nearby at the time of varve deposition. Dropstone may also be deposited by beach or river ice of middle latitudes.

PATTERNED GROUND AND OTHER FEATURES

The presence of fossil patterned ground, ice wedges sandstone, sandstone dikes, and cast involutions characterize periglacial conditions, where the ground is continuously or periodically frozen. Permanent frozen ground, or permafrost, was defined by Müller (1947) as a thickness of soil or other superficial deposit, or even of bedrock, at a variable depth beneath the surface of the earth in which a temperature below freezing point exists continually for a long time.

Patterned ground is a group term for more or less symmetrical forms such as circles, polygons, nets, steps, and stripes developed under intensive frost action. The large tundra polygons or ice-wedge polygon formed by processes of contraction in extremely cold conditions are only found in permafrost regions (King, 1976, p. 15). Frost cracking in a permafrost milieu is commonly accompanied by growth of ice wedges. The ice wedge grows as a result of surface water or groundwater filling the fissure and freezing. Later the ice melts and cracks may be filled with extraneous material resulting in silt, sand or pebbe wedges or dikes.

Washburn (1979) stressed the importance of ice wedges when they are found in fossil form as paleoclimatic indicators. Contraction cracking on a small scale may form in areas where winter conditions are severe and freeze-thaw activity is common. The crack pattern tends to be polygonal hexagonal similar to that occurring in volcanic rocks with a dominant angular intersection of 120º.

Periglacial involutions are contortions of bedding and interpenetration of one layer by another. This type of penecontemporaneous deformation structure comprises disturbed, distorted, or deformed sedimentary layers. Penecontemporaneous deformation may arise in a variety of ways such as ice shove, mass movement, turbidity currents, differential loading, ice undermelt and tectonism (Embleton and King, 1975). Although most of these structures are not typically periglacial in origin, it is important to know their genesis in order to identify their depositional deformation. Some deformational structures may be incorrectly identified, damaging a good environmental interpretation. This task is not easy. Load-structures, for example, are sole markings generally preserved on the lower side of the sand overlying a mud layer. On the surface, they appear as swellings, varying in shape from slight bulges, to deep or shallow rounded knobby bodies, to highly irregular protuberances (Reineck and Singh, 1980). Generally, such bulge load structures vary in size from a few millimeters to several decimeters (Potter and Pettijohn, 1963).

The structures' size is very important, because some large deformation structures, considered as load structures by many investigators, may not be so. Other interesting features are convolute bedding or convolute folding that in the past were regarded as distinctive of turbidity currents. Convolute bedding is a structure showing marked crumpling or complicated folding of the laminae of a rather well-defined sedimentation unit (Potter and Pettijohn, 1963). Even though layers may be strongly folded they are remarkably continuous; faulting and slippage are not normally associated with convolutions (Reineck and Singh, 1980). These characteristics are important for recognizing convolute structures.

In turbidites, in the division C of the Bouma sequence (Bouma, 1962), ripples, wavy or convolute laminae may be present. Because of the presence of convolute bedding in the Bouma sequence, it was early regarded as typical of turbidite sequences, but convolute bedding has also been identified in tidal flat, river, delta and lacustrine environments (Reineck and Singh, 1980).

Undermelt structures occur when stagnant ice, before melting, is covered by sediments. Sediments collapse over an irregular surface showing deformation structures. In this case a similar type of sediment may occur above and below the irregular surface, suggesting that the structure is not due to loading.

Distorted bedding, showing decollement-like folding and small-scale faulting, is a widespread feature in glaciolacustrine and glacio fluvial sediments (Reineck and Singh, 1980). This may be produced by an overriding mass of ice causing the sediment to slump.

The identification of these main glacial and periglacial criteria discussed above may indicate the presence of widespread glacial activity at a determined time in the geological record of a basin.

INDIRECT EVIDENCE FOR GLACIATION

LOESS

Loess is a nonconsolidated sediment, commonly nonstratified, consisting chiefly of quartz silt with subordinate fine sand and clay, deposited primarily by wind (Flint, 1975). Loessite, or lithified loess, is siltstone with at least two-thirds of its particles silt sized.

Due to high temperature gradients in glaciated areas anti-cyclonic winds are very vigorous. During the winter, river beds are exposed to wind activity, so ventifacts, sand dunes and loess deposits may develop around the ice-sheets. Sand dune deposits commonly occur relatively close to source areas, but loess material may be found far away from the ice-margins. Loess deposits mantle topographic highs and lows and decrease in grain-size and thickness downwind away from the source (Smith, 1942). According to Flint (1975), Pleistocene loess covers a total area of about 3.5 million square kilometers in Europe and North America. Wide areas are also covered by loess in South America (Patagonia) and Asia. Older loess deposits are seldom recognized in the geological record, but may have covered large areas during glacial events as they did during the Pleistocene glacial episode.

Edwards (1979) recognized Late Precambrian glacial loessites in north Norway. In western New York State, the Late Devonian Portage shale shows a loess-like section which attracted the attention of Twenhofel (1932, pp. 81, 175). He suggested that the marine mudstone (Nunda Shale) could be wind-derived. (It would be interesting to date the Nunda Shale more precisely in order to determine whether it is correlative with the Mid-Famennian glacial event in northern South America.) Black carbonates rich in silt-size particles may be formed in temperate regions in times of widespread glaciation as a result of strong wind activity generated in glaciated areas.

REGRESSIONS AND TRANSGRESSIONS

Indirect evidence of glaciation and cold temperatures includes rapid sea-level changes due to the exchange of water between ice-sheets and oceans. Sea-level changes due to glacial episodes are difficult to recognize in Phanerozoic times, but during Late Ordovician and late Cenozoic times widespread regressions and high clastic supply correlate with glacial events. Sea-level fluctuations as recorded by Permo-Carboniferous cyclothems have also been related to glacial and interglacial episodes. Eustatic changes in sea level may be caused by many different phenomena and have been frequently discussed in the literature (Hallam, 1977; Crowell, 1978, 1983).

MASSIVE BIOTIC EXTINCTION

Massive biotic extinctions have been attributed to a great number of diverse causes, and different extinctions may have different causes, but the Late Ordovician-Early Silurian and Eocene-Early Olgocene extinctions (Herman, 1981) may correlate with glacial episodes. In Eocene-Early Oligocene time, Antarctic ice-caps reached the shore as deduced from glaciomarine sediments around Antarctica (Frakes, 1978). The Late Devonian mass extinction may correlate with a cooling event in South America (Copper, 1977). Even the Late Cretaceous biotic extinction could somehow be related to a sharp climatic deterioration. The Permo-Triassic extinction could correlate with reversal of oceanic circulation due to a general warming up of the planet and widespread regression induced by the Hercynian orogeny that increased the oceanic capacity. That is, the subduction of oceanic ridges and the formation of mountains (orogenesis) in areas where previously existed continental shelves increased the oceanic capacity causing regression. Cold bottom water (approximately 3ºC at the present time) in polar and sub-polar marginal seas probably did not exist at the end of the Permo-Carboniferous glaciation; instead warm salty water, formed by high evaporation in low latitude marginal seas, may have played a fundamental role (Brass and others, 1982). In Late Permian time, 42 % of the total Paleozoic evaporites were laid down (Zharkov, 1981). This large amount of evaporites suggests an abundant production of warm salty water at that time. Low oxygen and CO_2 content, different chemical composition and bottom water temperature (possibly as high as 15ºC in Permo-Triassic time as it was in the Cretaceous) may have had an important role in causing faunal extinctions.

PALEOGEOGRAPHIC RECONSTRUCTION

During the Pleistocene glaciation, glaciers reached middle latitude areas and narrowed the warm climatic belt (Figure 4), while during times of low or nonglacial development the warm climatic belt may have reached high latitudes, as deduced from the presence of tropical plants at 60 degrees N in Miocene times (Frakes, 1979). These latitudinal changes in climatic belts may be determined from lithic and biologic climatic indicators. The presence of supposed glacigenic sediments in high latitude areas may confirm the glacial origin of such rocks.

PALEOMAGNETIC DATA

During the past few decades paleomagnetic data have become more accurate and reliable and they are now a powerful tool for paleocontinental reconstructions. If paleomagnetic paleopole positions are consistent with paleogeographic reconstructions based on lithic and biologic climatic indicators, times of glaciation may be determined under the premise that tillites are found in high latitudes. This model seems to work well for Phanerozoic glacial events, but not for many of the Precambrian glacial events. Precambrian tillites appear to exist in both low and high latitudes (Harland, 1964a,b). In this case, it may be found either that the low latitude Precambrian tillites are related to altitude, or that the paleomagnetic poles were not coincident with paleogeographic poles, or that the magnetic have been misinterpreted.

CHAPTER 4. TECTONIC DEVELOPMENT OF SOUTH AMERICA FROM EARLY PRECAMBRIAN TO ORDOVICIAN TIME

South America has three major geotectonic elements: (1) the western Andean orogenic belt, (2) the South American platform, and (3) the foreland depression between the orogenic belt and the platform (Figure 7).

The South American platform may be subdivided into the Amazonian, Brazilian and Patagonian (Argentinan) platforms, and the platforms may comprise more than one shield (De Almeida and others, 1973).

The Brazilian platform is subdivided into the Atlantic shield, in the coastal area and the Brazilian central shield, in the center of the country; while the Amazonian platform is sub-divided into the Guyana shield, north of the Amazonas river valley and the Guaporé shield, south of the Amazonas river valley.

Similar rock successions, radiometric ages and structural behavior between the Guyana and Guaporé shields support the view that they have long been a single tectonic unit.

The Amazonian and Brazilian platforms show distinct tectonic histories. These two units may have collided during latest Precambrian-Cambrian time during the Brazilian cycle along the Paraguay-Araguaia fold belt or geosuture when the here named Goiás ocean may have closed.

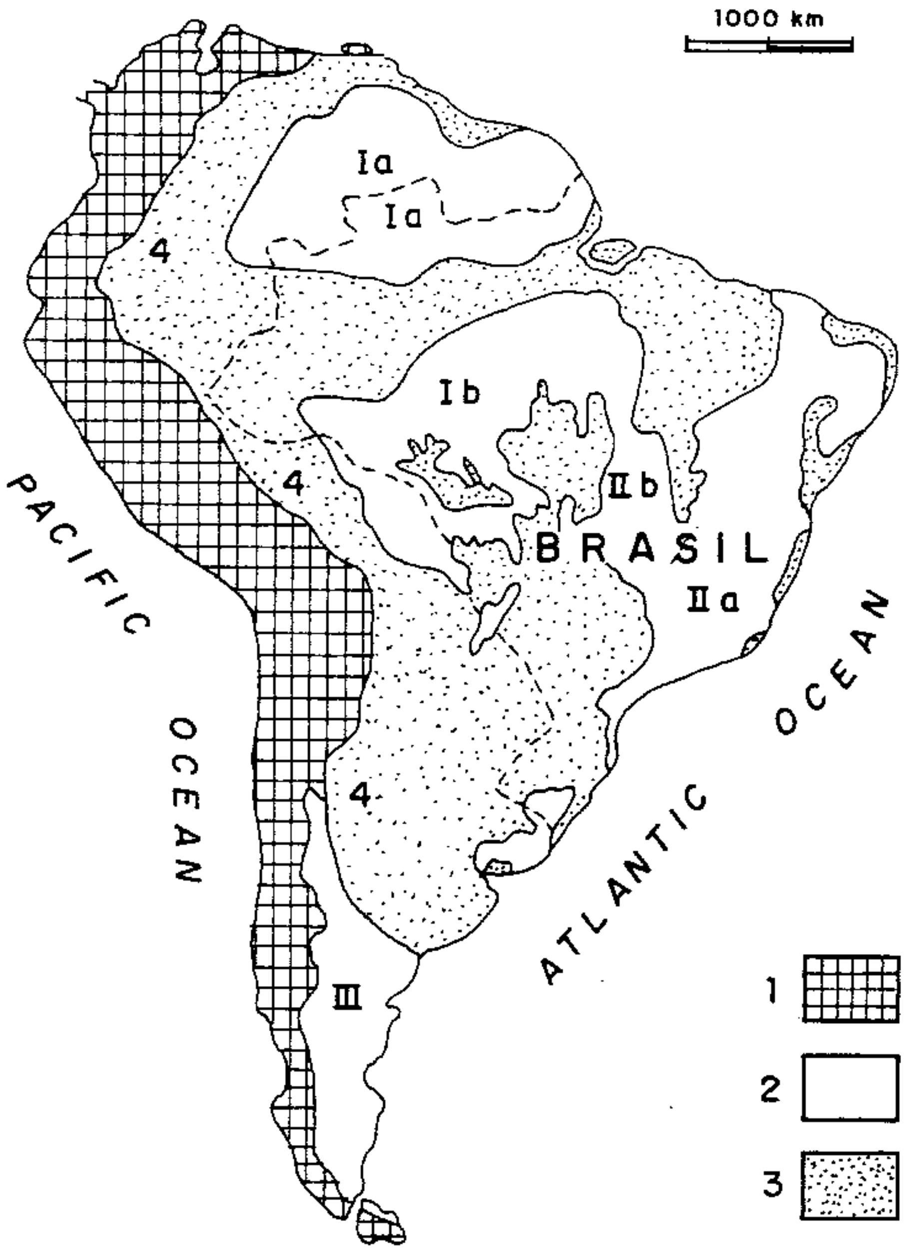


Figure - 7 Main South American geotectonic elements. 1- Andean fold belt; 2- South American platform; I-Amazonian platform (Ia-Guyana shield and Ib-Guapore' shield); Brazilian platform (IIa Atlantic shield and IIb-Central shield); III-Patagonic platform; 3-Phanerozoic cover; 4-Pericratonic trough. (Modified from De Almeida and others, 1981)

The presence of basic and ultrabasic serpentinized rocks (De Almeida, 1967) over a distance of 500 km along the Araguaia fold belt suggests the existence of an ophiolite complex in this geosuture. This ophiolite complex probably represents fragments of an older oceanic floor that was tectonically emplaced (obducted) into the Araguaia metamorphic belt. The ophiolite is about 1,000 m.y. old (De Almeida, 1967); its age represents the time of the generation of oceanic crust, and has little relation to the time of orogeny or subduction.

The Amazonian platform has not undergone a geosyncline-type deformation since 1,100 m.y. ago, whereas the Brazilian platform is probably the result of a series of ocean openings and closings from long ago up to 500-450 m.y. ago, when multiple continental collisions occurred. During this early Paleozoic interval several small oceans closed in eastern South America and Africa. The Brazilian geosynclinal seas and the Brazilian platform were intimately connected to the Pan-African geosynclinal seas; actually the Brazilian and Pan-African tectonic cycles correspond to the same tectonism that resulted in the closing of several oceans (Figure 8), and consolidation of the huge Pangea 1 continent. Between the coastal Ribeira and the Araguaia fold belts a small continent existed, made up of the Goiás block and the São Francisco craton.

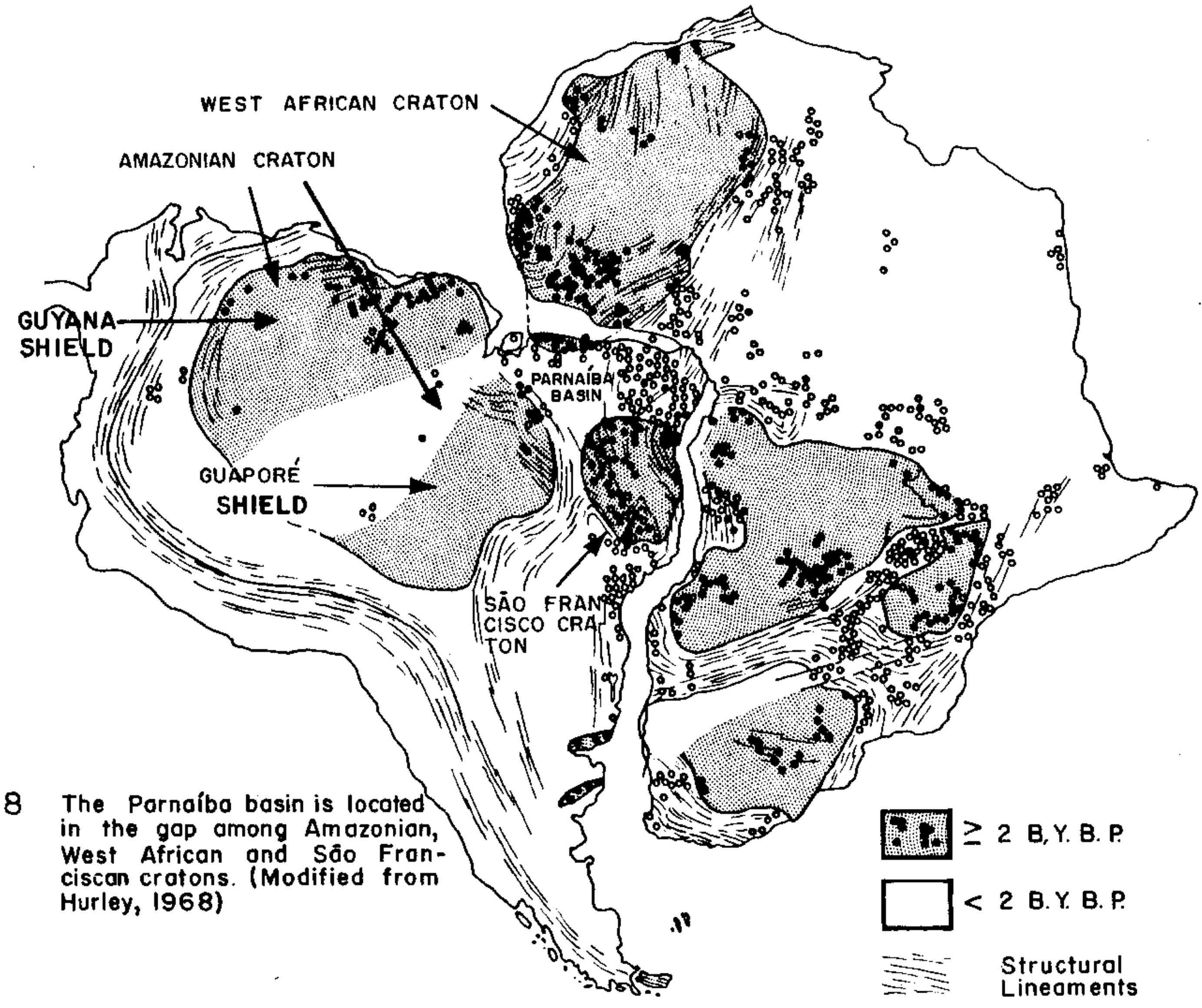


Figure- 8 The Parnaíba basin is located in the gap among Amazonian, West African and São Franciscan cratons. (Modified from Hurley, 1968)

Brazil is located almost entirely on the Amazonian and Brazilian platforms, except close to the Peruvian and Bolivian borders where it is situated in the Andean foreland depression (Acre Basin). The nation is underlain by four flat-lying, large Paleozoic intra-cratonic basins named: Paraná, Parnaíba, Amazonas and Solimões (Figure 2).

AMAZONIAN PLATFORM

The Amazonian platform or craton is made up of the Guyana and Guaporé shields (Figures 8 and 9). The Guyana shield is surrounded on the west by the Subandean-Colombian basins; on the north by the Orinoco River (Venezuela) and Atlantic Ocean; on the south by the Solimões and Amazonas basins and on the east by the Amapá Atlantic coastal plain.

In the north Andean areas, in Venezuela, isolated old Precambrian rocks are believed by Cordani and others (1979) to belong to the Guyana shield. This shield embraces the Guyana Republic, Suriname, French Guyana, and parts of Venezuela, Colombia and Brazil. It is drained by a dense network of Amazon and Orinoco River tributaries and it is covered by a thick tropi-cal forest and large savannah patches. Many of the rocks of the Guyana shield have only been studied on a reconnaissance level and only a limited number of radiometric age determinations exist

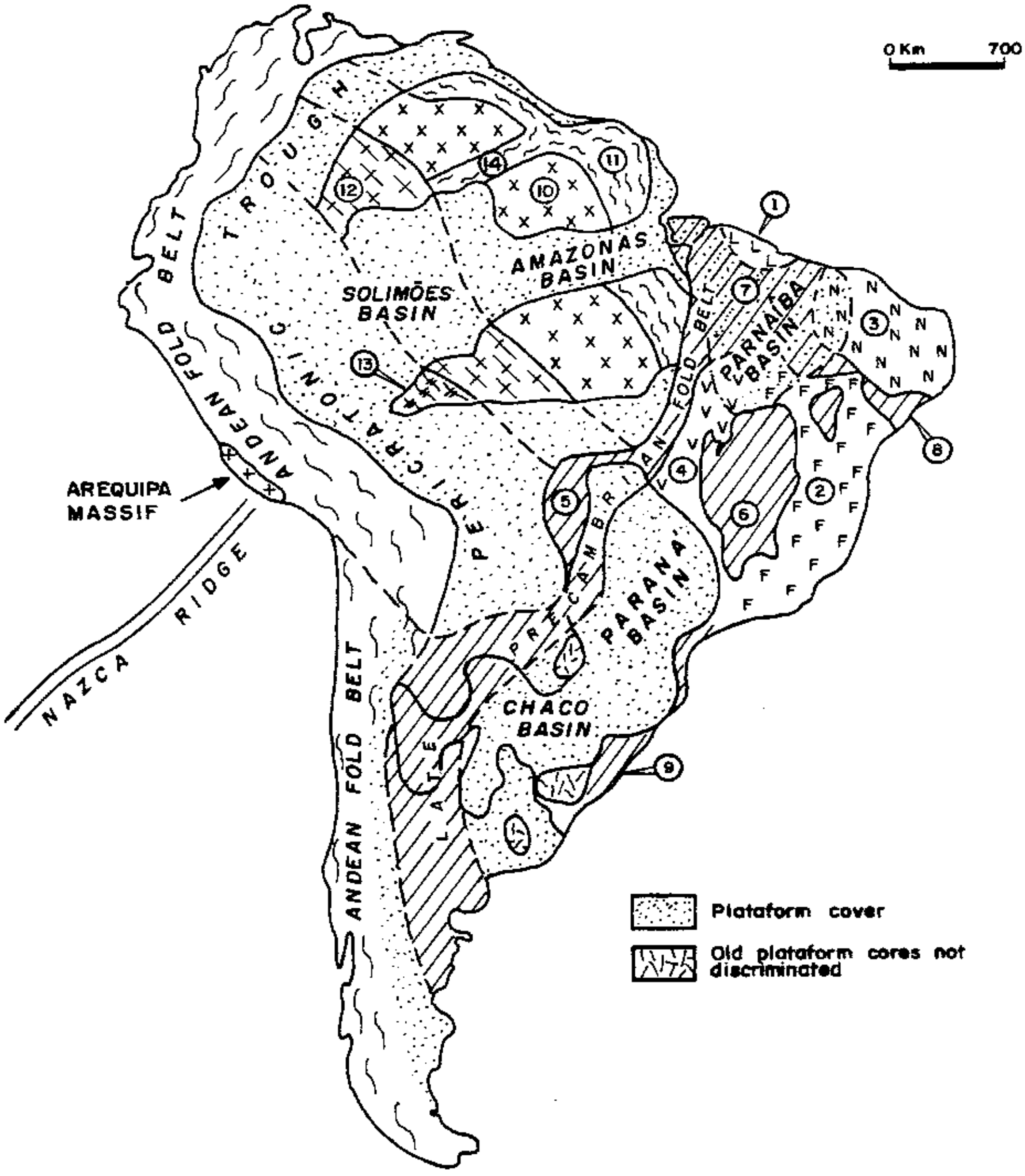


Figure- 9 Main South American tectonic units: 1-São Luis craton. 2- São Francisco craton. 3- Northeast shield. 4- Goiás Massif. 5- Araguaia fold belt. 6- Brasilia fold belt. 7- Araioses fold belt. 8- Proprió fold belt. 9- Ribeira fold belt. 10-Central Amazonian province. 11- Maroni-Itacaiunas fold belt. 12- Rio Negro-Juruena fold belt 13- Rondonian fold belt. 14-Central Guyana fold belt. Based on Cordani and others (1979) and De Almeida and others (1981)

Two tectonic events are well known in the most part of Brazil. The Transamazonian Cycle (2,100-1,800 m.y. old) and the Brazilian Cycle (750-450 m.y. old); other tectonic events are less known (De Almeida and others, 1981).

Rb-Sr whole-rock isochron work (Torquato and Cordani, 1981) suggests that at least three major tectonic cycles occurred in the Guyana shield, at about 3,600-3,000 m.y., 2,800-2,700 m.y. and 2100-1800 m.y. ago. The areas corresponding to each of these tectonic cycles are not fully mapped so events older than 2,100-1,800 m.y. (Transamazonian Cycle) are poorly recognized in Brazil, but are better identified in the neighboring countries. It is believed that the central area of the Brazilian part of the Guyana Shield is made up of very old rocks, some of them reworked in the Transamazonian Cycle.

The Guaporé Shield has an area over 1,600,000 km^2 and is covered by a tropical forest and savannah patches with an altitude averaging about 600 m. It is bounded on the north by the Solimões and Amazonas basins, on the south, southeast and east by the Araguaia fold belt (geosuture), over 3,300 km long that is located in the Brazilian platform.

The Araguaia belt (Figure 9), oriented roughly N-S in the northern part is, on a Gondwana reconstruction, in direct continuity with the Rokelide fold belt (Figure 10) of Sierra Leone,

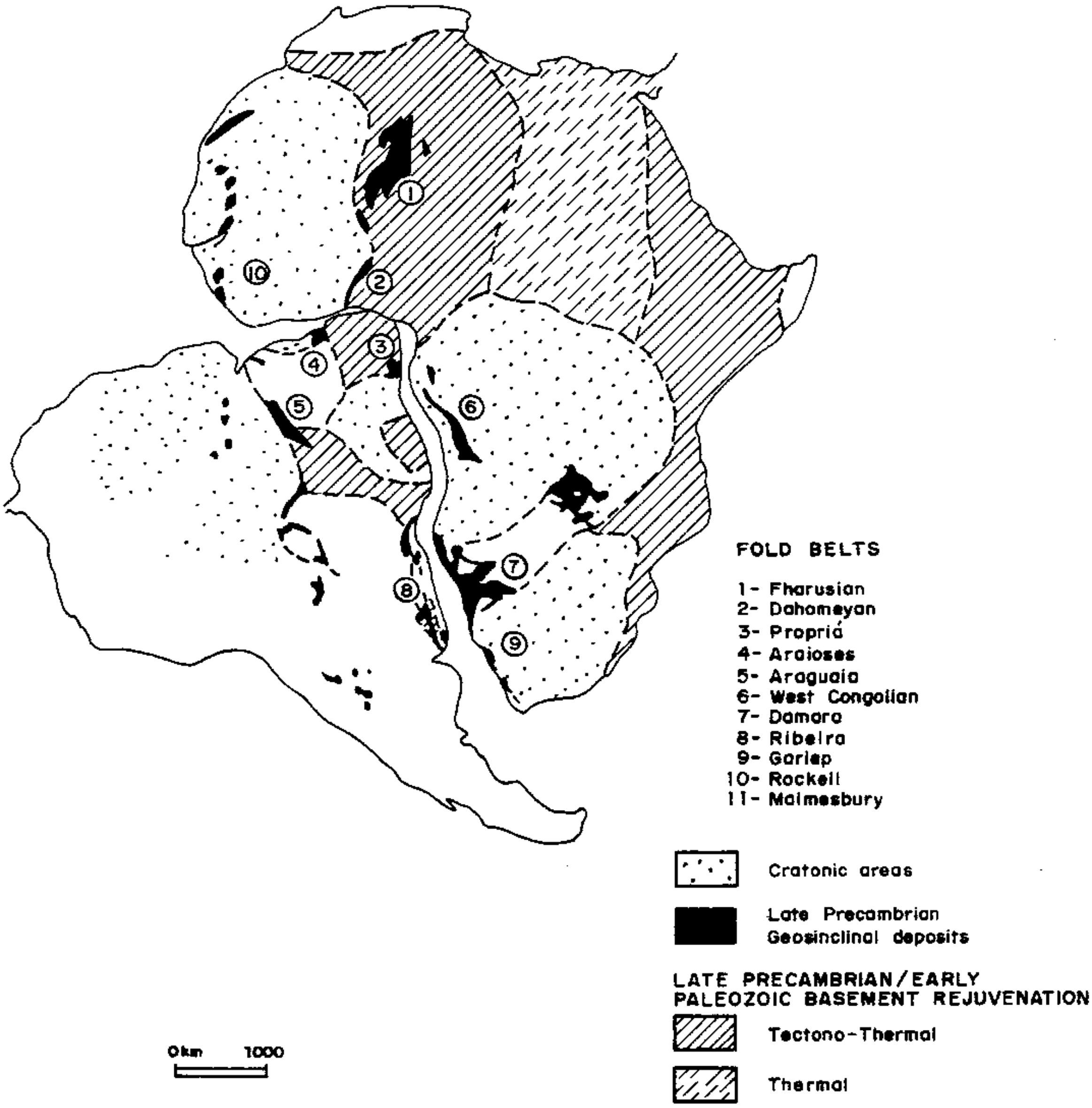


Figure- 10 Modified from Smith and Hallam's (1970) reassembly of Africa and South America showing distribution of Late Precambrian-Early Paleozoic fold belts and areas of basement rejuvenation / of the Pan-African-Brazilian cycle.

Africa (Torquato and Cordani, 1981; Hurley, 1973; Smith and Hallam, 1970). The south extremity of the Araguaia fold belt reaches La Plata (Argentina), at the Atlantic Ocean (Aceñolaza and Miller, 1982). On the west, the Guaporé shield is bounded by the Pericratonic Acre (Brazil), Beni (Bolivia) and Chaco (Paraguay) basins. However, in the substrate of the Andean area, Precambrian rocks are present and may belong to the Guaporé shield. The 2,000 m.y. old Arequipa massif of the coastal Cordillera of southern Peru may also have been part of the Guapore shield since the Rondonian orogeny (1,400 to 1,100 m.y. ago).

Shackleton and others (1979) reported an Rb-Sr whole rock age of 1,980 m.y. for the metamorphism of granulite gneisses from the Arequipa massif. These dates suggest that a contact may exist between the Arequipa massif, when the Brazilian thermotectonic overprint is removed, and the Brazilian platform which carries the Brazilian overprint, toward the south of the massif. The Arequipa Massif underwent several Paleozoic thermotectonic events (Shackleton and others, 1979) related to subduction of the Paleopacific ocean floor and is overlain by Paleozoic sediments with a typical South American biota. The Amazonian shield may have constituted ancestral South America which collided with present eastern South America and northwestern Africa (Figure 9) during the Brazilian cycle, as dealt with further below.

CENTRAL AMAZONIAN PROVINCE

The oldest formations found in the Guyana shield (Figure 9), composed of igneous and metamorphic rocks of amphibolite and granulite facies, yield Rb-Sr reference isochrones of about 2,700-2,800 m.y. as well as isolated age values within the 3,000-3,600 m.y. interval (Torquato and Cordani, 1981) in the Imataca complex of Venezuela on the north side of the Guri fault zone.

Old rocks that comprise the Guyana complex in Brazil are widely distributed and make up the basement upon which younger rocks lie on the Guyana craton (Issler and others, 1974).

The complex displays amphibolitic to granulitic metamorphic facies consisting of granulite, a great variety of gneisses, amphibolites, migmatites, anatextic and metasomatic granites, diorites, gabbros and ultramafic rocks (De Montalvão, 1976). These rocks are intensively folded with axes oriented in NW-SE and WNW-ESE directions. Strong granitization and migmatization is common everywhere in the complex.

The presence of older granulitic enclaves suggests that most of the Guyana complex is older than the Transamazonian cycle rocks, but was isotopically rejuvenated or remobilized during this latter orogenic cycle (Issler and others, 1974).

The oldest rocks occurring in the Guaporé craton comprise the Xingu complex (Silva and others, 1974) which is widely distributed and makes up the basement upon which younger rocks lie. The complex is composed of granulite, gneiss, migmatite, granite, granodiorite, diorite, trondhjemite, syenite, amphibolite, metavolcanic rocks, schist and quartzite (Issler, 1977).

The complex exhibits an amphibolitic to granulitic metamorphic facies and is older than 2,500 m.y. Amphibolite relicts from the Tapiragé hill have furnished an age as old as 3,280 m.y. and in northeast Goiás some crystalline rocks were dated as 3,000 m.y. old (Suszczynski, 1981). Several tectonothermal episodes have caused remobilization and isotopic rejuvenation of these rocks. The structural trends of this basement terrane are approximately parallel to that of the Guyana complex (NW-SE).

MARONI ITACAIUNAS FOLD BELT

Along the eastern side of the Amazonian platform, metamorphic-volcanic sedimentary sequences overlie the Guyana and Xingu complexes with regional angular unconformity (Figure 9). This rock assemblage makes up part of an extensive metamorphic belt named Maroni-Itacaiunas (Cordani and others, 1979). This rock assemblage (Vila Nova Group) is composed of schist, quartzite, itabirite (ironstone), amphibolite and serpentinite displaying greenschist to amphibolite metamorphic facies (De Almeida and others, 1981). Important manganese deposits are included in the group. Ectinites belonging to the group have a radiometric age of 2,090 m.y. (Cordani and others, 1973). In many places older basement remains occur in the area of exposure of the Vila Nova Group.

In the Guaporé shield the Xingu complex is overlain by the Grão-Pará Group which is composed of quartzite, phyllite, mica schist, jaspilite, itabirite and basic metavolcanic rocks (Issler, 1977). This group is equivalent to the Vila Nova Group and it contains economic iron ore formations in the middle part of the section. The fold axes trend NNW and the metamorphic grade ranges in age of greenschist to amphibolite facies. Mafic rocks yield an age of about 2000 m.y., corresponding to the Transamazonian cycle. Molasse marking the end of the cycle is represented by the Rio Fresco Formation that was intruded by granitic plutons about 1,850 m.y. ago (De Almeida and others, 1981).

The Vila Nova Group is a metasedimentary, metasomatic and magmatic product of the evolution of a geosynclinal complex developed during the Transamazonian cycle.

Choudhuri (1980) pointed out that the combination of basic to acid volcanism with submarine extrusion and sedimentation strongly suggests a back-arc-basin environment with subduction zones at the greenstone belt of the northern Guyana shield. Verhofstad (1970) also suggested that the main effects of orogeny and metamorphism in the greenschist facies, and the marginal acid volcanism resulted from plate descent off the pré-Transamazonian craton. These data are consistent with plate collision during the Transamazonian cycle.

In the northern portion of the Guyana shield, at the end of the Transamazonian event, about 1,850 m.y. ago, widespread volcanic-plutonic activity produced rocks of acid and intermediate composition represented by the Surumu and Iricoume formations. In the Guaporé shield important volcanism also took place.

After the magmatic activity, a continental molasse named Roraima Group was deposited in the area of the common boundary between the Republic of Guyana, Venezuela, Brazil and in Suriname. The rocks of the Roraima Group make up a flat-lying plateau where the sequence starts with polymictic conglomerate, sandstone and arkose, overlain by sandstone, shale and siltstone, and finally a thick sequence of sandstone and arkose is present at the top. It appears to have covered, originally, an area more than 1,200,000 km^2, and the main source region was towards the NW, where the Atlantic Ocean is today (Singh, 1974).

After deposition of the Roraima Group a very important phenomenon occurred that may be related to continental distension in the region close to the Roraima Basin. Widespread basic magmatism took place with the development of thick diabase sills and dikes, dated as 1,700-1,600 m.y. old (Prien and others, 1973). The Pedra Preta Diabase may be related to continental break up similar to that that occurred in Mesozoic. Before the use of radiometric age determinations these basic rocks were confused with Mesozoic ones related to Gondwana rupture.

CENTRAL GUYANA FOLD BELT

Lima and others (1982 in Caputo and others, 1983) redefined the Guyana central fold belt which may be an arm of the Maroni-Itacaiunas fold belt or may be an independent fold belt reworked during the Transamazonian cycle. This geotectonic unit extends from the Uraricoera River to the city of Paramaribo, Suriname.

RIO NEGRO JURUENA FOLD BELT

In the western side of the Guyana shield and western-central part of the Guaporé shield a fold belt named Rio Negro-Juruena is present (Cordani and others, 1979). This belt may contain many geosuture zones and may have developed between 1,750 and 1,400 m.y ago.

This area underwent extensive volcanic-plutonic magmatism producing rhyolite, rhyodacite, andesite, dacite, tuff, ignimbrite rocks and lahar deposits. Associated sediments and granodioritic, granitic and syenitic subvolcanic, circular intrusive rocks also occur. This rock assemblage may suggest that accretion continental was due to ocean-continent collision (Teixeira, 1978).

RONDONIAN FOLD BELT

In the southwestern part of the Amazonian platform, Cordani and others (1979) mentioned the presence of a younger fold belt named Rondonian. Its tectono-magmatism lasted from 1,400 to 1,100 m.y. ago and may be related to continent-continent collision with intense compressive deformation. This tectonism was responsible for faulting, granite and granodiorite intrusions and rhyolite, andesite, trachyte and basalt extrusions in the Rio Negro-Juruena fold belt area. The Arequipa massif may have been part of the continent that collided with Western Amazonian platform about 1,200 m.y. ± 100 m.y. ago.

About 1,200 m.y. ± 100 m.y. ago rocks of the whole Amazonian platform underwent a tectono-magmatic activation producing intense regional shearing and cataclasis with the formation of mylonite, pseudotachylite, tilted blocks, and dynamic metamorphism. Magmatic processes were active along the faults with intrusion of granitic rocks. This disturbance is known in the Brazilian part of the Guyana shield as the Jari-Falsino episode. In the Guaporé shield this same event is named the Madeira episode.

All of these processes, observed in the Amazonian platform, seem to be related to continental collision with intense compressive deformation in the Rondonian fold belt area. Following the Jari-Falsino episode, tholeiitic magmatism developed outside and on the southern limit of the Guyana craton where stocks and dikes yielded ages between 1,050 and 850 m.y. ago (Teixeira, 1978). This magmatismis perhaps related to new continental break up elsewhere in the craton.

Stewart (1972) claims that the Pacific Ocean opened about 850 m.y. ago in the western United States, and Matsumoto (1977) states that the Pacific Ocean opened at about the same time in Asia. Shackleton and others (1979), in studying Peruvian rocks, agreed with the opening of the Pacific Ocean at about that time. This period (around 850 m.y.) seems to be one of worldwide fragmentation. Possibly, many new oceans formed, synchronously with tholeiitic intrusions and extrusions.

During the final decline of the volcano-plutonism of the Amazonian platform, several sedimentary formations associated with minor basic volcanism, were laid down in distinct isolated basins. Presently, insufficient geological data, makes correlation between the formations in these basins difficult.

According to Issler (1977) the sedimentary cover overlies wide areas and is generally made up of non-mature lithic sandstone, lithofeldspathic sandstone, arkose, shale, siltstone and polymictic conglomerate. Such rocks are rich in unstable minerals, suggesting that they resulted from a first or second depositional cycle and originated in rugged source areas. The different formations are non-fossiliferous and occur in large isolated areas, making correlation difficult.

The Beneficente Group, overlain by the Caiabis and Guajará-Mirim groups and the Gorotire and Prosperança formations, (still in almost subhorizontal position) appear to be the oldest sedimentary Precambrian cover, slightly folded without signs of regional metamorphism.

The Beneficente Group, exposed in the north-central part of the craton is composed of sandstone, arkose, siltstone, shale, stromatolitic limestone, glauconitic sandstone with quartzite and quartz-mica schist restricted to fault zones (De Montalvão and others, 1979). It was deposited in continental and marine environments and the sea possibly invaded the area from the south. The general trend of the Beneficente folding is E-W with N-S inflections. De Montalvão and others (1979) dated the group as 1,400 m.y. old, based on Rb-Sr determinations on sediments.

Silva and others (1974) attributed the origin of the folding observed in the group to orogenic movements, but it seems that the depositional and structural milieu was epicontinental instead of geosynclinal, and the disturbance may have been caused by the Madeira shearing event related to stresses originated in the Rondonian fold belt which in turn are interpreted as due to continent-continent collision.

The Caiabis Group was laid down in a graben located at the southwest of the shield and is composed of conglomerate and sand-stone with intercalated basalt. The upper clastic beds are designated Dardanelos Formation. The lower and upper basalts yielded ages of 1,416 ± 14 m.y. and 1,225 ± 20 m.y. respectively (De Montalvão and others, 1979).

The Guajará-Mirin Group is situated in the western area of the shield and is located in two grabens named Pacaás Novas and Uopione, separated by a basement high. The sections are composed of arkose and sandstone underlain by basalt 1,038-967 m.y. old (Leal and others, 1978).

The Gorotire Formation occurs in the eastern part of the Guaporé craton and comprises arkose, sandstone, chert and siltstone with stromatolitic structures and conglomerate beds.

The Gorotire fold axes are oriented approximately NW-SE. Its age Is unknown, but it underlies unconformably the Prosperança Formation. The paleogeographic situation at the time of deposition of this unit is unknown but the sea communication probably was located at the south (Suszczynski, 1981).

The Acari Formation made up of limestone, chert and red beds overlies the Prosperança Formation. The Prosperança Formation is composed of red, pink, cream and white sandstone, siltstone, conglomerate and arkose beds. In some regions, both units are the basement of the Amazonas Basin Paleozoic sequence, and they also crop out in the Guaporé and Guyana shield areas. De Montalvão and others (1979) questioned the correlation of the Prosperança Formation with the Prainha Formation made by Caputo and others (1971) and considered the Prainha Formation older and equivalent to the Dardanelos Formation of the Caiabis Group. Diabase dikes in the Prosperança and Dardanelos formations are 1,400 and 1,100 m.y. old (Montalvão and others, 1980) indicating that the Prosperança Formation may also not be correlated with the Dardanelos Formation.

BRAZILIAN PLATFORM

The Brazilian platform is composed of several structural units trending roughly N-S in Brazil (Figure 9). The westernmost structural unit is a long fold belt, named Araguaia, that stretches from the Amapá shelf in the Atlantic Ocean close the French Guyana up to La Plata (Argentina), again in the Atlantic Ocean. Radiometric dates of granulite facies metamorphism of the Arequipa massif (Peru) correspond to those from the Transamazonian cycle (Shackleton and others, 1979). Shackleton and others (1979) did not report Brazilian orogenic cycle activity in the 2,000 m.y. old Arequipa massif. This observation suggests that the massif is connected with the Amazonian shield. To the south, in the Sierras Pampeanas of northwestern Argentina, radiometric ages indicate that the imprint of the Brazilian orogenic cycle may have extended as far west as Salta Province (McBride and others, 1976). The northern limit of the effects of the Brazilian cycle is not known but it probably lies southeast of the Arequipa massif. Important tectonic units observed in the Brazilian platform comprise the Goiás massif, São Francisco and São Luis cratons, Araguaia, Araiosis, Brasília, Propriá and Ribeira fold belts.

ARAGUAIA FOLD BELT

The Araguaia fold belt represents a geosuture that separates the Amazonian domain from terrane affected by the Brazilian orogenic cycle and is here interpreted as recording a continental collision in Late Precambrian-Cambrian times between na Amazonian passive continental margin and a Brazilian active continental margin, along a probable subduction zone dipping towards the east (Figure 11). The Araguaia fold belt is viewed as an area of crustal impact and collision between two crustal plates. This very impressive geosuture can be deduced in the world gravity map (Figure 12).

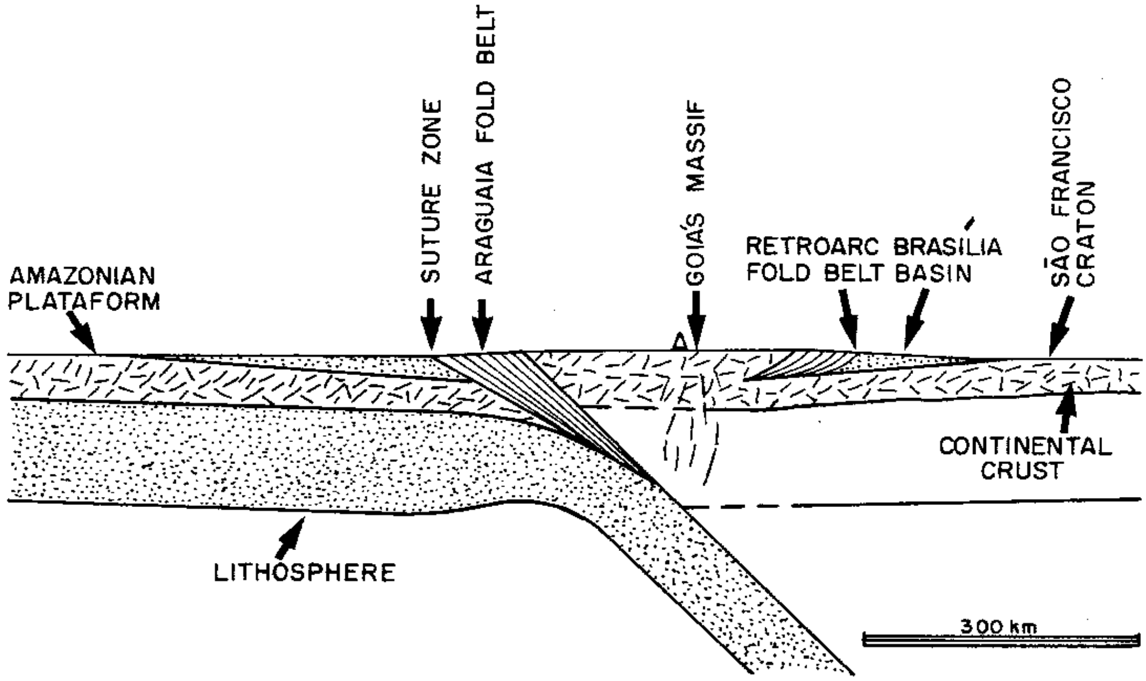
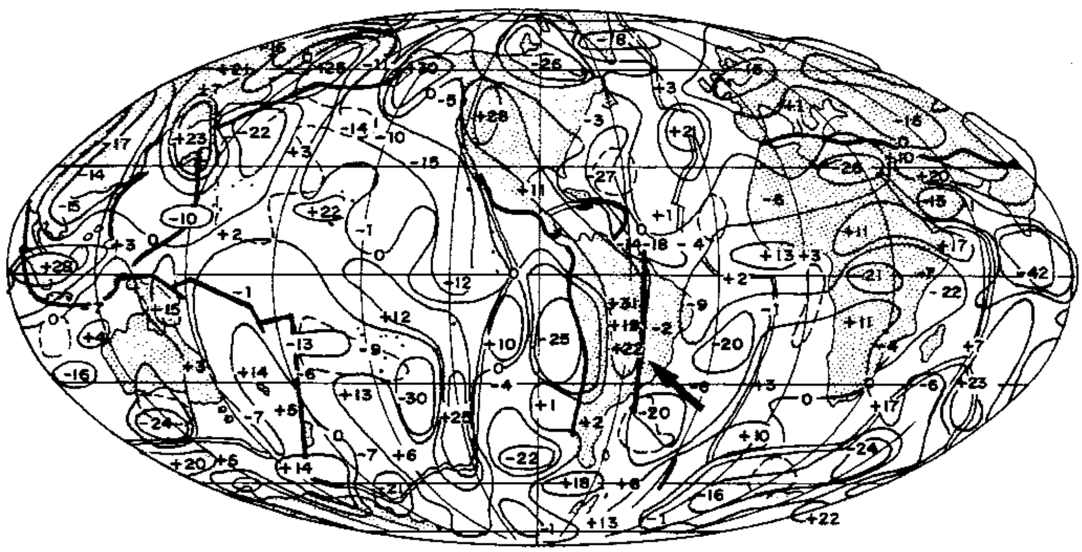


Figure- 11 The Goiás massif is regarded as an unroofed magmatic arc above the subduction zone of the Araguaia geosyncline according to Dickinson's model (1977)

In this map (Figure 12) gravity boundary separates rocks of the Guaporé shield (+19 to +31 milligals) from rocks of the São Francisco craton (-2 milligals). Since this crustal impact, the Brazilian fold belt areas have remained stable, while tectonic activity continued at the western edge of South America in the Andean area, and at the southern tip of South America (Cobbing and others, 1977). The subduction zones of the Brazilian fold belt may have changed from the edge of the Goiás block to the western edge of the South American continent, along the present Pacific margin at the end of the Brazilian cycle.

Peru has a history of successively rising bodies of calc-alkaline magma, parallel to the present Pacific margin which began about 450 m.y. ago; this indicates a change in the structural behavior from a passive Atlantic-type margin to an active Pacific-type margin at about that time (Shackleton and others, 1979). South of the Arequipa massif or south of the Andean bend, tectonic episodes have occurred almost uninterrupted from Late Precambrian to Phanerozoic times. In southern South America, from the Sierras Pampeanas massif, successively younger tectonic events occur towards the west and south from Late Precambrian up to present time (McBride and others, 1976). The Araguaia mobile belt verges towards the Amazonian craton and displays greenschist metamorphic facies at the Brazilian-Bolivian border. It contains Precambrian



Geossuture across South America
Compression
Tension
Approx. 3000 Fathom

Figure- 12 Free air anomalies in milligals referred to a fifth-degree figure. Calculated from the spherical harmonic coefficients / of the gravitational field of degrees 6 through 16 of Gaposchkin and Lambeck (1971).

World gravity map shows a sharp contrast between Amonian (+19to+31 milligals) and Brazilian (-2 milligals) platforms along the Araguaia fold belt indicating a large discontinuity between both geotectonic units.

trench deposits represented by flysch sediments, unconfonnably overlain by conglomerate and glacial marine deposits (De Almeida and others, 1981). Above an angular unconformity, stromatolitic carbonates followed by pelites were deposited (De Almeida and others, 1981).

In the north part of the Araguaia fold belt in the Tocantins arch (which separates the Parnaíba from the Matajó Basin), the basement is made up of phyllite, schist and quartzite. In the western Parnaíba Basin region, gneiss, amphibolite, schist and quartzite occur at the base of the section, followed by diamictite and phyllite with quartzite intercalations, itabirite and metavolcanic rocks of the Tocantins Group. The fold axes are oriented N-S and before the break up of Gondwana this rock assemblage had direct connection with the Rokel Series (north west Africa) which comprises tillite, phyllite with quartzite and limestone intercalations, itabirite and metavolcanic rocks (Torquato and Cordani, 1981).

ARAIOSES FOLD BELT

Beneath the Parnaíba cover, one arm of the Araguaia fold belt, named here Araioses fold belt, may continue northeast in the direction of the Dahomeydes fold belt that is located on the eastern side of the West African Craton.

In wells RP-1, RP-2 (Rio do Peixe one and two), MO-1 (Mocambo) and

JU-1 (Jerusalém) drilled by Petrobras, close to the Ferrer Arch, in the São Luís Basin (Figure 2), more than 2,000 m of metamudstone and phyllite occur with diamictite of possible glacial origin at the base of the section preserved in the Araioses fold belt. The glacial nature of the diamictite was here interpreted on the basis of its texture: a gray clay matrix supports angular and rounded clasts of great petrographic variety up to 15 cm across as observed in cores. Metamorphism of the diamictite occurred about 483 ±38 m.y. ago, based on a Rb-Sr age determination (Thomaz Filho, 1981, written communication), therefore at the end of the Brazilian cycle. This indicates that these diamictites have no relation to the Late Ordovician glacial rocks of northern Brazil and northwestern Africa, but may be as old as Late Precambrian.

Late Precambrian glacigenic rocks have been recorded over much of Africa, particularly in the Rokelides, Dahomeydes, and Pharusian fold belts and in the Taoudeni and Volta basins of north West Africa (Culver and others, 1980). The presence of Late Precambrian glacial sediments in the same stratigraphic position strengthens the validity of lithological correlations between north west Africa and north east South America. The uplift which preceded the opening of the Pan-African, Brazilian and other oceans at low latitudes may have caused formation of highlands where ice developed in times of general refrigeration.

For example, Late Precambrian tilllites in western North America are found at the bottom of a thick section deposited after rifting that gave origin to a new ocean (Stewart, 1972). At that time, low latitude rift shoulders could have been glaciated due to strong upwarping and the glacial products could have been deposited in the foothill basins. Another important conjecture is that in the area of the Parnaíba Basin a triple junction may have existed between the Amazonian, West African and São Franciscan convergent plates when Brazilian and Pan-African oceans were being closed.

SÃO LUÍS CRATON

Beyond the northern-northwestern limit of the Parnaíba Basin, in the São Luís area, gneiss, schist, amphibolite and metavolcanics rocks occur intruded mainly by granite and diorite. These rocks have been folded, metamorphosed and granitized during the Transamazonian tectono-thermal event, 2,000 ± 200 m.y. ago (Hurley and others, 1967; Kovach and others, 1976). In a pre-drift Gondwana reconstruction the São Luís cratonic area appears to have direct connection with the West African craton where sediments deposited in old geosynclinal basins have been folded, metamorphosed and invaded by granites. This constitutes evidence of a major event 2,000 ± 200 m.y. ago, referred as the Eburnean event in Africa (Bonhomme, 1962).

GOIAS MASSIF

East of the Araguaia fold belt, a long massif made up of migmatite, granulite and ultrabasic rocks occurs with ages as old as 3,700 m.y., 2,600 m.y. and 2,000 m.y. (Marini and others, 1977). This massif is part of the São Francisco craton. The Brazilian tectono-thermal imprint is shown by a K/Ar age determination on biotite from a granite, which yielded 530 ± 16 m.y. (Hasui and De Almeida, 1970). These rocks may be interpreted as diapiric granites formed during the Brazilian cycle. Cordani and others (1973) think that some of the granitic rocks may actually have been formed during the Brazilian cycle, but the same granite also yields whole rock ages as old as 1,330 m.y. and 2,000 m.y. (Marini and others, 1977) so more work is needed to determine the correct time of formation of these granites. Field relationships seem to indicate that the Bambui Group metamorphosed at 600 m.y. ago (Thomaz Filho, 1981), has been disturbed and metamorphosed by N-S elongated granite bodies in this area. The rocks of the Goiás Massif show evidence of polymetamorphism, isotopic rejuvenation and intense fracturing (De Almeida and others, 1981). The extremities of the Goiás Massif are covered by the Parnaíba and Paraná Paleozoic basin sediments.

Satellite low gravity values found in the Goiás Massif and São Francisco craton suggest thickening of the continental crust, perhaps due to subduction of part of the Guaporé craton under the Goiás massif (Figure 11).

The central part of the Goiás Massif and the Araguaia mobile belt were probably pushed against the Guaporé shield during the Brazilian Orogeny, since the tectonic transport is towards the west. The Goiás Massif may have been a leading edge of the São Francisco plate (Figure 9). The massif is here envisaged as an unroofed magmatic arc above the subduction zone of the Araguaia geosyncline, possibly analogous to the 2,000 m.y. old Arequipa Massif in central Andes today. The 800 km long and 100 km wide Arequipa Massif, is located within the present day magmatic arc-trench gap, and the arc volcanoes are situated beyond its NE side, but in Paleozoic, Mesozoic and Tertiary times calc-alkaline plutonism and volcanism occurred within the massif, indicating the former position of the magmatic arc (Shackleton and others, 1979). Batholiths of various Phanerozoic ages intruded the Arequipa Massif indicating an active subduction zone beneath it.

BRASILIA FOLD BELT

The Brasflia fold belt (De Almeida, 1967) occurs east of the Goiás magmatic arc; it trends roughly N-S and is 1,100 km long. with vergence towards the east.

From bottom to top the rock section comprises shale, sandstone and shale with limestone and dolomite intercalations, followed by red clastic sedimentation at the end of the tectonic episode. Andesitic lavas in the north and subvolcanic granitic intrusions in the south constitute the main magmatic activity (De Almeida and others, 1981). This belt is here seen as a backarc foldthrust belt according to Dickinson´s model (1977) shown in the Figure 11. On the east margin of the Brasília belt a retroarc foreland basin developed, that covers part of the so-called São Francisco craton.

The area between the Araguaia and Brasília fold belts is named the Tocantins province by De Almeida and others, (1981).

SÃO FRANCISCO CRATON

The São Francisco craton is an old structural block that lies In the eastern part of Brazil, close to the coast (De Almeida, 1977). According to De Almeida and others (1981) the craton is made up of granitic-gneiss complexes migmatized and metamorphosed to high-grade amphibolite and granulite facies. Granitoid, mafic and ultramafic rocks occur throughout the craton. Radiometric determinations have yielded ages of about 3,100 m.y., 2,600 m.y., 2,000 m.y., 1,200 m.y. and also reveal an imprint of the Brazilian cycle on its edges.

The northwest part of the craton is overlain by sediments of the Parnaíba Basin. Several Precambrian sedimentary covers overlie the craton, including the Macaúbas Group which records continental glaciation in Late Precambrian time (Rocha-Campos and Hasui, 1981). These cratonic tillites could be equivalent to glacial sediments found in Brazilian and north Pan-African fold belts. Widespread carbonate sedimentation represented by the Bambui Group followed the tillites. Metamorphism of the Bambui Group was dated as old as 600 m.y. (Thomaz Filho, 1981).

In its central area, the Espinhaço mobile belt was deformed and metamorphosed in greenschist metamorphic facies about 1,200 ± 100 m.y. ago (De Almeida and others, 1981). Sedimentation, producing deposits 4,000 m thick, lasted from 1,700 m.y. to about 1,100 m.y. ago in the Espinhaço belt, and may have begun after a phase of continental break up. The São Francisco craton makes up the São Francisco Province of De Almeida and others (1981).

RIBEIRA AND PROPRIÁ FOLD BELTS

Northeast, east and southeast of the São Francisco craton near the present coast is another mobile belt named Propriá in the north and Ribeira in the south (Cordani and others, 1968). This area has been so deeply eroded that alternated bands of the old rejuvenated basement and Brazilian cycle rocks are present. Because of such intense erosion, old reactivated basement crops out in faulted or folded structures, while Brazilian rocks are preserved in synclines and structural lows. The basement, remobilized during the Brazilian cycle, is composed of gneissic-migmatic complexes of Early to Middle Precambrian age (De Almeida and others, 1981).

The Brazilian cycle rocks comprise phyllite, schist, quartzite and carbonate rock a thousand meters thick (De Almeida, 1968). This sequence may have constituted an old carbonate shelf around the São Francisco craton. The metamorphism ranges from greenschist to amphibolite facies; magmatism lasted from 650 m.y. to 540 m.y. and is composed of stocks and batholiths of granitic rock. Molasse deposits were laid down between 510 and 470 m.y. ago (De Almeida and others, 1981).

I believe that the Propriá fold belt may have been connected to the Dahomeydes and Araguaia fold belts and then separated by another triple junction between the West African, São Franciscan and Nigerian plates. These mobile belts record another line of continental collision between the eastern sides of the West African and São Franciscan cratons and the northeast and central African cratons (Figure 10).

The Ribeira belt comprises a basal sequence of sandstone, siltstone, shale with carbonate lenses near the top and an upper sequence composed of shale, mudstone and turbidites. The rocks show greenschist to amphibolite facies metamorphism, and granitic plutons intruded the section from 650 to 540 m.y. ago. Post orogenic deposits, with andesitic and rhyolitic lavas extruded between 500 and 450 m.y. ago, filled small deep basins in which thick sedimentary sequences over 5,000 m thick were laid down. The final cooling of the belt took place about 450 m.y. ago (Torquato and Cordani, 1981).

In northeast Brazil, at the end of Brazilian collision, brittle deformation characterized mainly by dextral strike-slip faulting took place. The fault system has severely disrupted the fold belt; and erosion was so active that at present it is difficult to determine the original position of the different blocks.

The E-W trending Pernambuco and Patos faults are the best developed ones onshore, while N-S trending conjugate faults are less well developed. Several faults can be traced across the margins of South America and Africa in a pre-split reconstruction. The E-W faults are parallel to the Saint Paul, Romanche, Chain and other oceanic fracture zones. For example, the Romanche fracture zone links the São Luís area to Ghana in a pre-drift reconstruction. The Pernambuco fault zone runs into the 27°5 fracture zone and into the Ngaoundire fault and the Patos fault runs into

Cameroon (Allard, 1969). This suggests that Brazilian and Pan-African tectonisms were very important in controlling the break up of Gondwana during Mesozoic times. The NE trending mylonite zones, probably corresponding to thrust fault planes, developed perpendicular to major stresses and are cut by the E-W strike-slip faults. Some investigators consider that this fault system is very old, but it was probably established at the end of the Brazilian cycle as a direct result of continental collision between the eastern part of South America and central and western part of eastern Africa.

In Africa the same brittle deformation system cuts 586 m.y. old granites and 530 m.y. old ignimbrites (Ball, 1980), so the Pernambuco strike-slip fault event may be younger than 530 m.y. The equivalent of the Pernambuco faulting event was interpreted by Ball (1980) in Africa as similar to that which took place when Tibet and China were indented by the Indian block during the Himalayan orogeny (Tapponier and Molnar, 1976). These faults extend over 1,000 km in Africa (Ball, 1980), but if the Brazilian continuation is considered they extend to about 2,000 km in length. In Asia, the Altyn Tagh strike-slip fault, more than 3,000 km long is attributed to continental collision between India and Asia.

It is commonly said that the break up of Gondwana did not follow basement trends in northeast Brazil. The basement in that area has actually split apart along the E-W and N-S Brazilian-Pan-African strike slip fault system directions, although in the south, the Ribeira metamorphic belt has split apart mainly along directions of the Brazilian cycle thrust fault system.

From Ordovician time on, the thrust and strike-slip faults acted as normal faults with vertical displacements. These faults were of considerable importance in controlling the shape, position and structural behavior of the sedimentary intracratonic and rifted coastal basins. In Cambro-Ordovician time, a large land mass named Pangea 1, made up of parts of present configuration South America, North America, Africa, India, Antarctica and Australia formed, although some parts were accreted in Paleozoic and Mesozoic times by late orogenic events (Aceñolaza, 1982).

PLATE TECTONICS VS. ENSIALIC TECTONICS

It is appropriate here to discuss a relevant subject which is under much debate: The question is to what extent the concept of plate tectonics may be applied to the development of the Precambrian-Early Paleozoic Brazilian and Pan-African orogenies.

In South Africa and Brazil, the concept of an "in situ" ensialic or intracratonic orogeny without significant sea floor spreading has been developed as an alternative to plate tectonics.

Hurley and Rand (1969, 1973) and Hurley (1973) presented geological evidence for ensialic mobile belts which may have rifted apart slightly, accumulated sediments and were later intruded by granitic rocks, as if a craton were split by rising hot material without being separated for great distances. Hurley and others (1967), Hurley (1968, 1973) and Hurley and Rand (1973) correlated the Amazonian cratonic rocks and tectonic events with those of the West African craton and suggested that the Rokelide fold belt between both cratons was formed "in situ" with little or no oceanic seaways. I regard the tectonic and rock correlations correct, but do not agree with the concept that during the folding both cratons reunited with the same original position and orientation as they had before rifting apart.

The Transamazonian and older rock fold axes are oriented NW in the eastern Guyana shield whereas the Eburnian (2,000 ± 200 m.y. old) and pre-Eburnian rock fold axes are oriented NE in the West African craton. The almost orthogonal structural orientation of the cratons suggests that they reassembled with a position and orientation different from the original one, during the Brazilian-Pan-African orogeny. Only the Brazilian-Pan-African trends, corresponding to the Rokelides and Araguaia fold belts line up. The boundary drawn by Hurley (1973) between Eburnean and pre-Eburnean rocks in the West African craton and between the

Transamazonian and pre-Transamazonian rocks in the Guyana craton is no longer correct because pre-Transamazonian granulites occur in eastern Amapá (De Almeida and others, 1981).

Hurley (1973) also pointed out that there was no paleomagnetic evidence showing that the cratons had been split apart by oceanic distances, but Morel and Irving (1978) based on paleomagnetic data from North America and South Africa, proposed a reconstruction where a large ocean existed between South Africa and North America 800 to 675 m.y. ago and, that at about 600 m.y. ago, these oceanic distances were consistent with North America welded to the West African craton. An ocean may therefore have surrounded the West African craton. In the Rokelide and Dahomeyde fold belts, the existence of ultrabasic and basic rocks of ophiolitic character, normally interpreted as pieces of oceanic crust, also suggests the presence of a paleoocean in the region (Burke and Dewey, 1973; Black and others, 1979; and Burke and others, 1977).

In Brazil, the Brazilian mobile belts have previously been considered as having formed in an ensialic milieu, without development of oceanic crust during its evolution. De Almeida and others (1973) explained that the substrate of the southeastern Brazilian Ribeira mobile belt was formed by rocks cratonized in Transamazonian or earlier cycles and then regenerated during the Brazilian cycle. They considered the abundance of acid plutonism and vulcanism and the scarcity of basic and ultrabasic vulcanism to indicate that the fold belt had a continental sialic nature. They pointed out that there was no evidence of an oceanic basin in the area occupied by the South Atlantic Ocean, generalizing this view to all South American Late Precambrian-Cambrian mobil belts.

In my opinion a model based on plate tectonics explains better the structural behavior, the distribution of igneous and metamorphic rocks, and the paleomagnetic and gravimetric data. The opening and closing of an ocean, named the Wilson cycle (Dewey and Burke, 1974), is the most powerful means of interpreting the evolution of an orogenic belt. Wilson (1968) analyzed geotectonic cycles in terms of oceanic evolution from youth (continental break up) to old age (continental collision) when the ocean floor disappears by subduction. Therefore, only fragments of oceanic crust may be preserved. The subduction zone is converted into a suture belt where the fragments of older oceanic crust may be tectonically emplaced and may rest on the continental orogenic belt. The obducted fragments are named ophiolites and are interpreted as outcrops of the ocean floor (Figure 13).

Ophiolites comprise the following rock types in order of descending stratigraphic position (Moores and Vine, 1971):

a. Carbonate or siliceous abyssal sedimentary rocks. Low-K tholeiite, commonly pillowed.

b. Low-K tholeiitic gabbros and diabase commonly exhibiting cumulus texture (stratification), interpreted as formed in a magmatic chamber, at sea-floor spreading centers.

c. Ultramafic rocks usually exhibiting a metamorphic fabric

This rock sequence may be complete or incomplete. The gabbro-ultramafic contact in ophiolite complexes is usually interpreted as the fossil Mohorovicic (Moho) discontinuity.

Ophiolitic complexes have been described in the Pan-African fold belts around the African Taoudeni Basin (Rokelide, Dahome and Pharusian mobile belts) by Burke and Dewey (1973) and in the Brazilian fold belts (Araguaia and Ribeira mobile belts) by De Almeida and others (1973, 1981) which are the South American continuation of the Pan-African fold belts of Africa.

The coastal South African Damara orogenic belt, corresponding in a pre-drift reconstruction to the eastern part of the Brazilian Ribeira orogenic belt, has been regarded in the past, as an ensialic orogene, but was recently interpreted by Porada (1979) in terms of a closing ocean and continental collision. He suggests the presence of a former ocean in the place now occupied by the South Atlantic Ocean. Most of the Damara fold belt is located on

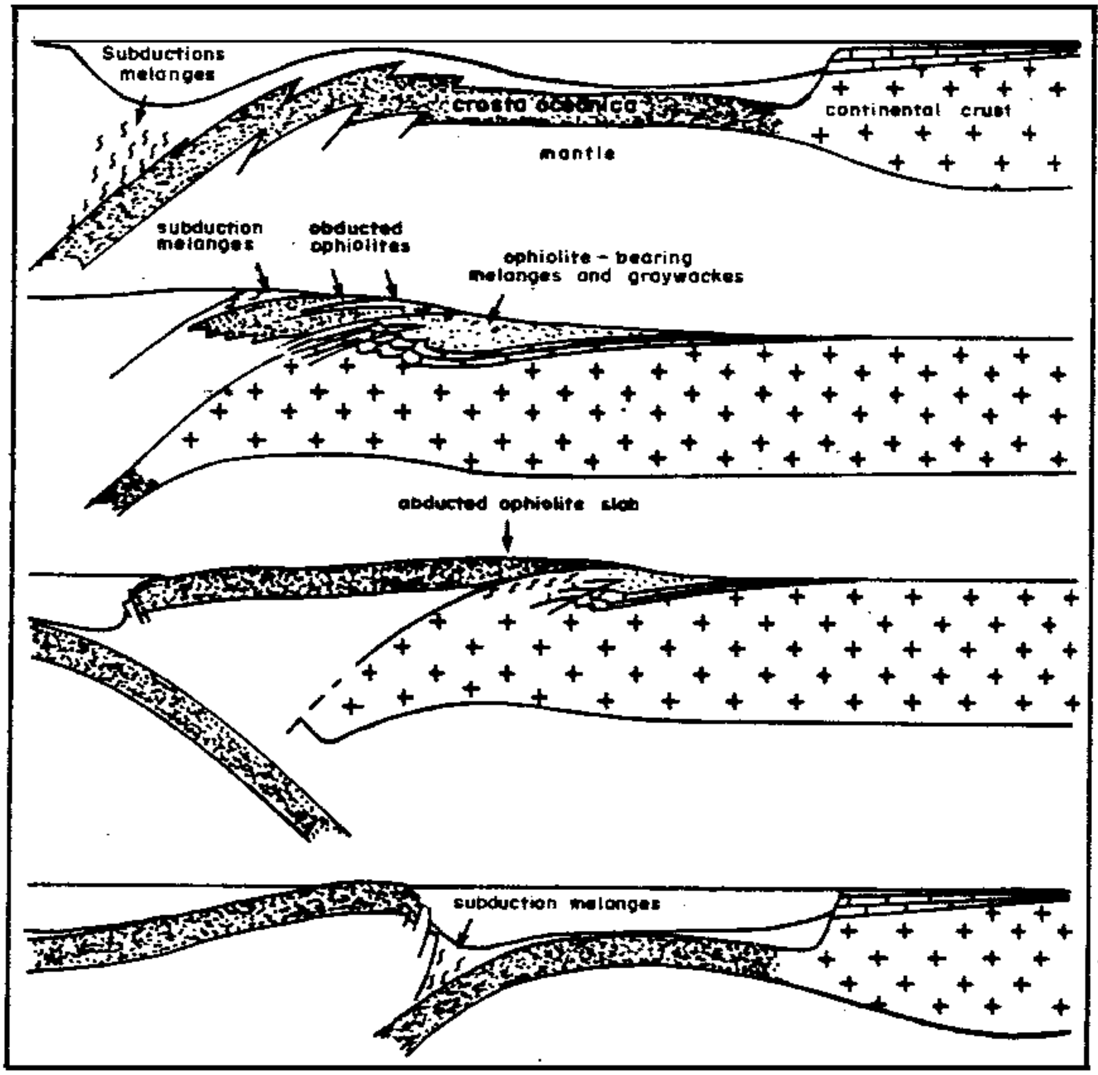


Figure 13. Possible obduction mechanisms for emplacement of ophiolites (after Dewey and Bird, 1971)

the continental shelves of the South Atlantic Ocean. Another arm of the same Damara mobile belt, occurring across South Africa, was similarly described as formed in an intracratonic environment (Martin and Porada, 1978) and is now envisaged by Barnes and Sawyer (1980) as resulting from ocean crust subduction and continental convergence, on the basis of geochemical, structural and petrographic studies. These interpretations suggest that plate tectonic activity existed in Late Precambrian time. Thus, the Brazilian and Pan-African cycles are here interpreted in terms of Plate interaction.

Beloussov and others (1979) discussed the data which contradict past mobilistic reconstructions of the position of continents (plate tectonics). They stressed that north Africa and north South America have had a completely different geologic history in their development, and they do not recognize that the Guinean, Ghanaian (and Saharan) basins were the northern continuation of the Amazonas and Parnaíba basins. However, there is a very good correlation between north African and north South American coeval sediments as it can be concluded from lithologic, biologic, tectonic and paleoclimatic similarities shown in this study. These similarities are considered by many investigators and by me as evidence of a close connection between northwestern Africa and northeastern South America from Late Precambrian to Early Cretaceous times.

CHAPTER 5. ORIGIN AND DEVELOPMENT OF NORTHERN BRAZILIAN BASINS

The origin of the northern Brazilian intracratonic basins is poorly understood. This qualitative discussion analyzes the genesis and epeirogenesis of these basins based on the geologic record, as well as on the concepts of isostasy and plate tectonics

BASIC CONCEPTS

According to Menard (1980, in Bird, 1980), the causes of epeirogenesis may be (1) external loading and unloading, (2) bending of plates plunging into subduction zones, (3) internal density changes, and (4) dynamic effects of mantle motion.

External loading depresses the crust due to the earth's elasticity and plasticity. The elastic deformation is immediate and quickly recovered when the force causing the deformation is removed. Under an extensive load over a prolonged period, the mantle flows and the crust depresses plastically (Daly, 1934). Beyond the loading an uplift should develop. The load can be sediments, thrust fault plates, volcanic rocks, water or ice. Unloading can be caused by erosion, evaporation (dissection of a basin), ice melting and removal of water (regression). Unloading has the opposite effect on earth's crust. In ice-covered regions, because the loading process is fast in geological terms, a peripheral depression forms and beyond it a well developed forebulge. The forebulge is concentrated in front of loading because the mantle viscosity is too great to allow unifonn spreading of asthenospheric material widely over the surface of the Earth. The forebulge hypothesis proposed by Daly (1934) demands pure broad bending of the crust without localized zones of vertical fracturing.

The total change in sea level from a stage free of ice on land to one corresponding to maximum widespread Pleistocene glaciation is more than 150 m. The glacio-isostatic rebound of some Pleistocene ice-covered areas is on the order of 300 m and the forebulge lowering is approximately 80-100 m with an assumed width of 3,000 km (Newman and others, 1971, p. 796). When an ice sheet expands, the proglacial depression and forebulge migrate outward. When the ice sheet retreats, the forebulge migrates inward and collapses while the peripheral depression migrates inward and becomes shallower as a result of isostatic rebound. A stable ice cap generally does not occupy the peripheral depression, as water does, due to ice viscosity. At the terminus of glaciers, evaporation and melting take place, causing deposition in lakes and depressions. The transfer of water from sea to land and vice-versa causes worldwide regressions and transgressions, with corresponding facies changes in the geologic record. Sedimentary loading also produces a forebulge but it is imperceptible because its growth may be in balance with erosion.

ORIGIN OF THE BASINS

The origin of the Parnaíba Basin as discussed here is speculative. The Parnaíba Basin lies in the gap between the São Luís craton (continuation of West African craton) in the north and the Goiás massif and São Francisco craton in the south (Figure 8). Due to its position protected between both cratons, the area of the basin may not have been subject to strong deformation during the general convergence of the Amazonian, north West African and São Francisco cratons with the Eastern African craton, so that folded mountains did not form. Probably, after the multiple collision, general cooling in the low folded area contributed to subsidence of the basin.

The origin of the Solimões Basin is not yet known and may be a foredeep basin related to Andean tectonics since the Late Ordovician time. The Amazonas Basin presents physical features that may indicate its origin. Gravity surveys made by Petrobras show that the depositional axis of the basin is characterized by an E-W belt of gravity highs that are more than 1,300 km long. The belt is interrupted by a broad gravity low in the western extremity of the basin at the Purus high. Each gravity anomaly has a broad circular or elliptical shape varying from +50 to +95 milli-

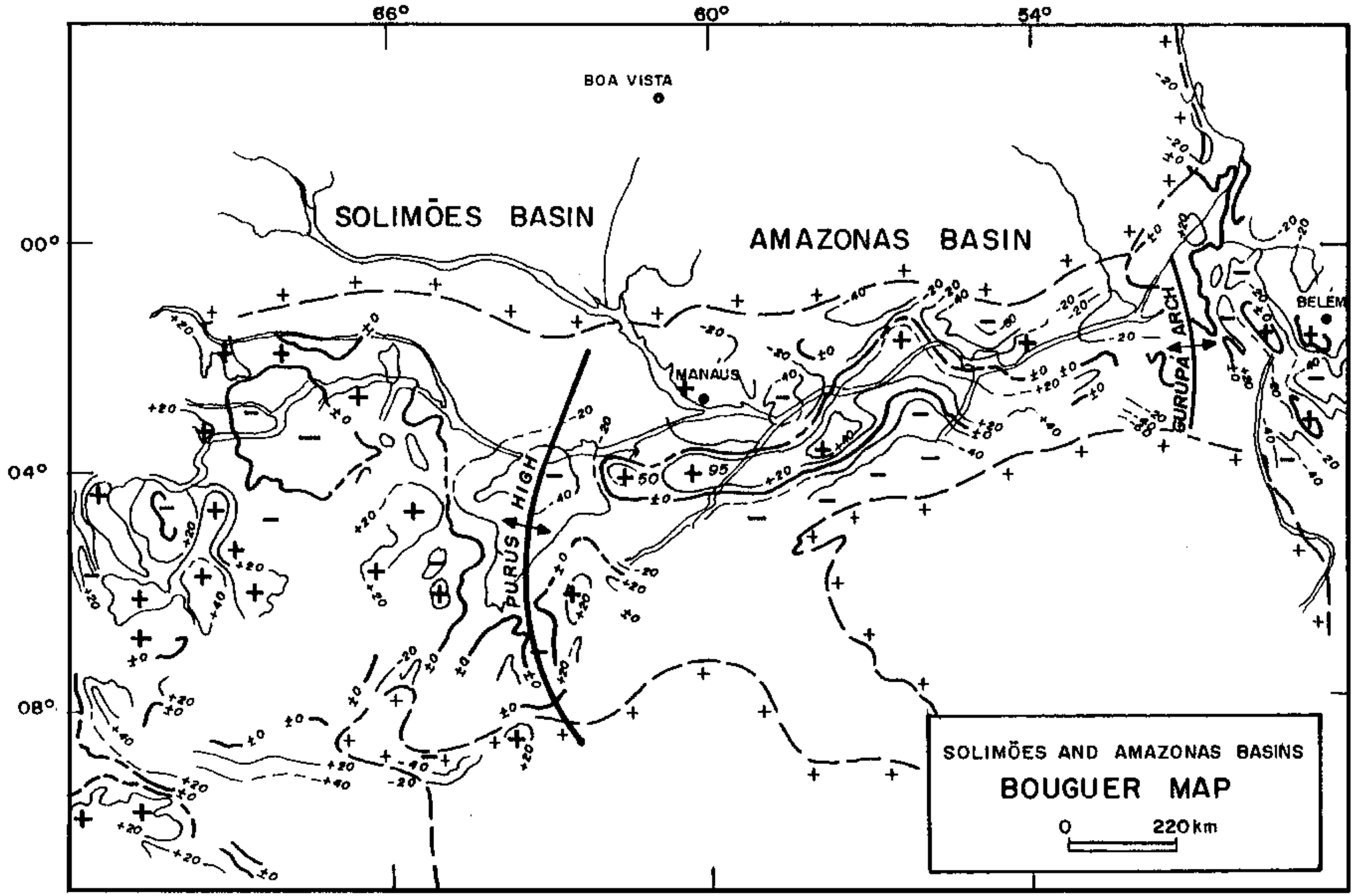


Figure- 14 Bouguer map of Amazonas and Solimões basins. A belt of gravity highs is observed at the Amazonas basin axis related to heavy ultrabasic rocks connected to the initial rifting which caused the basin.
Bouguer map prepared by Petrobras.

gals. On each basin flank there is a belt of gravity lows with some scattered small gravity highs (Figure 14). One of these marginal highs was drilled (well 1-CM-1-PA) by Petrobrás and a pyroxenite was cored. The coincidence of gravity highs with the basin axis suggests an intimate connection between the heavy masses and the formation of the basin. In the eastern Amazonas Basin, the chain of gravity highs bifurcates.

Linsser (1958), based on gravity surveys and presence of ultrabasic rocks, developed the following hypothesis for the origin of the Amazonas Basin:

1. Tencion forces in the crust opened it
2. The cracks were filled with magma up to a hydrostatic level.
3. The cracks were filled with sediments from the top of the magmatic rock to the surface. This load caused a disturbance in mass equilibrium, and further sinking.
4. And finally, the earth's crust was rifted or downwarped forming a shallow basin that was filled with more sediments.

On the other hand, Porto (1972) proposed the same explanation used by McGinnis (1970) for the origin of Illinois and Michigan basins. In those basins, it is thought that initial deposition was caused by collapse of incipient rift systems due to the load of high density crustal intrusives.

I propose here, in addition to drawing on earlier works, that a combination of two mechanisms may be responsible for the origin of the Amazonas Basin: (1) A thermal event may have produced initial uplift of the lithosphere and volcanism followed by erosion, rifting, and cooling, causing subsidence, and (2) subsequent bending of the basin floor in response to sedimentary loading.

The Amazonas Basin is regarded as a rift valley, that is, "an elongate depression overlying places where the entire thickness of the lithosphere has ruptured in extension" (Burke, 1980). It was a rift that failed to develop into a new ocean, and is inferred that the Amazonas rift was connected to the opening or to the closing of the Goiás Ocean. During closing of the Goiás Ocean, the rift may have been formed or reactivated, resulting in accumulation of a thick Paleozoic sedimentary sequence. The connection (triple junction) between the Amazonas rift and Araguaia-Rokelide fold belt is presently located to the east of the Amazonas River mouth.

Geophysical surveys in Amazonas River Mouth Basin (offshore) has revealed a gravimetric anomaly associated with a fracture zone. Rezende and Ferradaes (1972) considered this anomaly as caused by intrabasement extrusion of tholeiitic basalt related to Early Cretaceous tectonism and magmatic activity. Ponte and Asmus

(1978) pointed out that the anomaly may be considered as an offshore extension of those old gravimetric highs that are present in the Amazonas Basin or the anomaly could have been caused by 180-220 m.y. old basic intrusions related to widespread volcanic activity in the interior Amazonas Basin.

Another possibility exists. The gravity anomaly in front of the Amazonas River mouth may have been caused by the presence of ophiolites in the Araguaia-Rokelide fold belt which is beneath the continental shelf in the Amazonas River Mouth Basin. The time of Amazonas Basin rifting is not known because the pyroxenite gave dates as old as 622 ± 30% on the basis of Rb-Sr method (Thomaz Filho, 1982, written communication). If the age of the pyroxenite is about 800 m.y., it may correspond to the time of opening of he Goiás Ocean, but if the age is about 450 m.y., it may correspond to the time of closing of the Goiás Ocean. In this case, the rift would be classified as an impactogen (Burke, 1980). An Early Paleozoic age is favored because first sedimentary rocks were laid down in Ordovician time.

Before rifting, uplift may have occurred, along with simultaneous erosion of volcanic, sedimentary and basement rocks in the domed area and deposition in peripheral depressions. When the area started to subside, the lighter acidic rocks around the ultrabasic stocks may have subsided together with the denser ones, as suggested by the gravity lows on the basin flanks and gravity lows (lighter rocks) among the gravity highs corresponding to the ultrabasic stocks.

BASIN AMPLIFICATION

During downwarping, interconnection and widening of these depressions may have taken place as a result of stream concentration and activity. The depression probably drained to the Theic Ocean (ocean between Gondwana and ancestral North America before the Acadian orogeny (McKerrow and Ziegler, 1972). Erosion may have removed local base levels and temporary sediment storage sites so that the Amazonas Basin was invaded by the sea in Early or Middle Ordovician time when a rise in sea level may have triggered a new stage of basin subsidence. Although the age of the lowermost strata is not known in the region, due to a lack of fossils in the section, they correlate with Ordovician rocks of northern Africa (Taoudeni Basin). After thermal subsidence, the underlying lithosphere may have responded to the sediment and water load by bending. The actual deformation may have extended beyond the original rift as a result of the lithosphere's rigidity, therefore causing basin broadening (Beaumont and Sweeney, 1978).

The sedimentary record shows that the basin grew by continuous onlap; younger beds covered older ones on the margins as the basin broadened. At times of moderate sea level lowering, sedimentary coarse fill was deposited in the central parts of the depression. For example, a load of a 50 m column of water over a few thousand years causes subsidence of approximately 17 m (Burke, 1979). Such initial subsidence would continue as long as the sea occupies the area. This mechanism allows for deposition of a thick sequence of shallow marine sediments in the Amazonas Basin. Outside the basin, a small forebulge may have formed due to outflow of asthenospheric material from beneath the basin floor. However, the elevation of the ground may have been imperceptible because of the balance between uplift and erosion.

The first sedimentary cycle in the Amazonas Basin may have begun in the Early or Mid Ordovician and stopped in late Early or early Mid-Silurian times when sea water removal from the basin brought an end to sedimentation and subsidence.

The second sedimentary cycle began in late Early Devonian (Emsian Stage) and ended in the Mid Early Carboniferous (Visean Stage).

Devonian sediments overlap Silurian ones, indicating greater transgression in Devonian than Silurian times and basin amplification.

The third sedimentary cycle began in Late Carboniferous time (Westphalian D Stage) and ended in the Late Permian time.

The fourth sedimentary cycle began in, the Cenomanian Stage and ended in the Senonian Stage, and the fifth sedimentary cycle began in Paleocene time and ended in the Recent time.

CHAPTER 6. GENERAL REMARKS ON STRATIGRAPHY

There is still some confusion in the literature about the designation of northern Brazilian intracratonic basins. In the past, almost any geological reference about the Paleozoic Amazonas Basin dealt with the area between the Purus Arch and the coast, which encompasses half the eastern part of the Amazonas region.

Since the fifties the Brazilian Amazonas Basin name has been extended toward the western border of the country and subdivided into upper (western part), middle and lower Amazonas basins. Since then, each investigator has considered different areas as Lower and Middle Amazonas basins. For some people, the Lower Amazonas Basin is considered as only the area close to the Amazonas River mouth where there is a Mesozoic rift (Marajó Graben or Basin). For others the region between the Purus Arch and Gurupá Arch has had its western side called the Middle Amazonas Basin and its eastern side called Lower Amazonas Basin, although in my view there is no evidence to support the recognition of two distinct basins between the Purus and Gurupá arches. Other workers have considered only the upper and middle basins without a lower basin, or, they see the upper and lower basins as only one basin.

An additional source of confusion is that many Subandean basins are also called upper Amazonas basins, although there is a tendency to be given for them different designations. Loczy (1963) distinguished the following upper Amazonas basins:

(1) Upper Amazonas Basin in western Brazil (states of Acre and Amazonas)

{2) Upper Amazonas Basin in northeastern Bolivia (Rio Beni and Caupolican regions)

(3) Upper Amazonas Basin in eastern Peru (Montaña region)

(4) Upper Amazonas Basin in eastern Equador (El Oriente)

(5) Upper Amazonas Basin in southeastern Colombia (Potumayo and Caquetá regions)

Due to the overuse of the designation of “Upper Amazonas Basin”, it is proposed here to designate the western (upper) basin as Solimões Basin and the eastern basin simply as Amazonas Basin, dropping the words upper, middle and lower in Brazil (figure 15).

The graben occurring at the Amazonas River mouth has long been known as Marajó Basin. This graben should not be called lower Amazonas Basin (Figure 2).

The Parnaíba Basin is also known as Maranhão, Piauí-Maranhão, and Meio Norte (Middle North). The most used designation is Parnaíba Basin, and other names should not be encouraged. The Solimões Basin (700,000 km^2) has its designation derived from the Solimões River (name of the Amazonas River upstream of Manaus) and comprises the area from the Peruvian-Colombian border to the west of the city of Manaus. It has a section as thick as 3500 m.

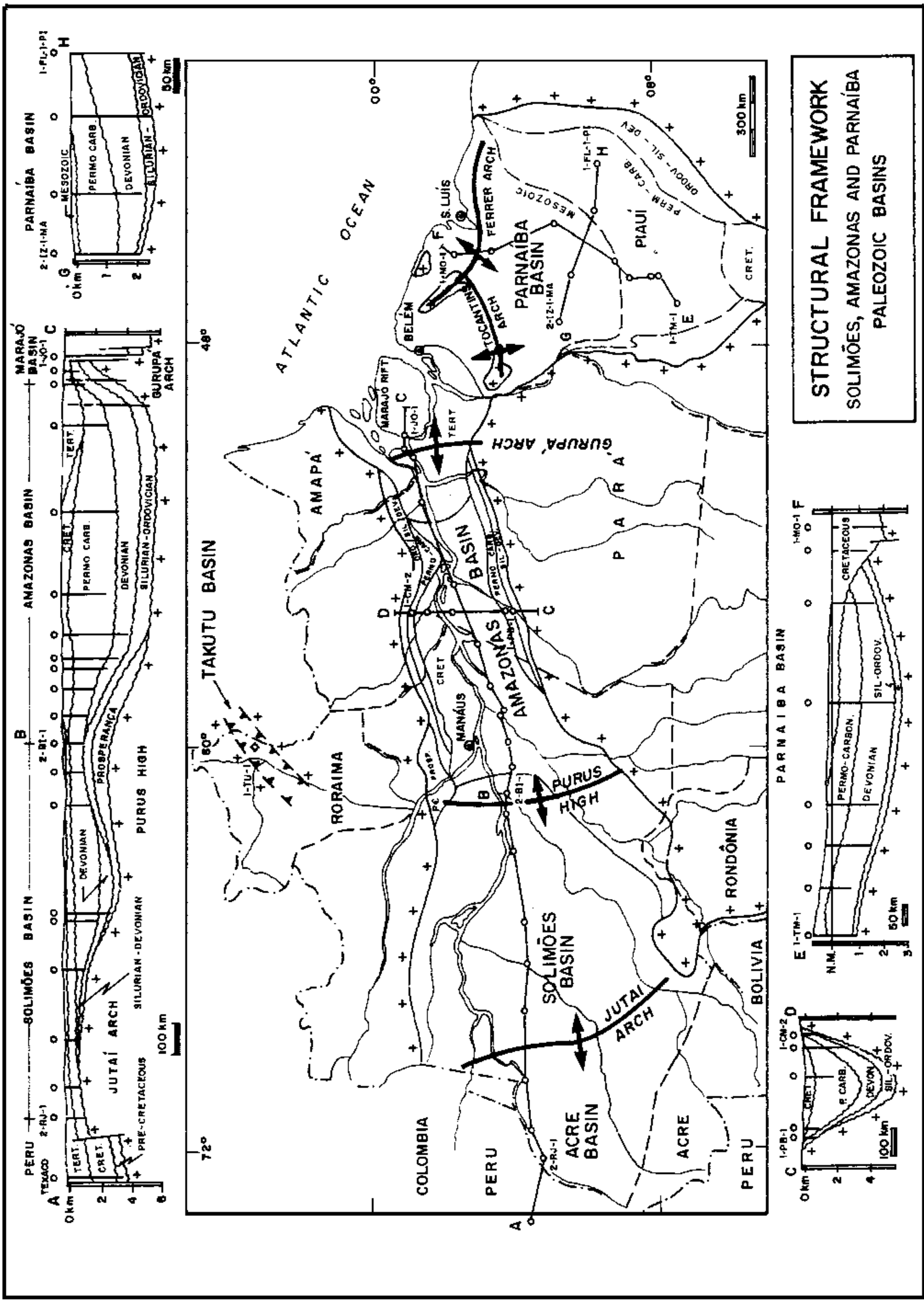


Figure 15 Structural framework of Solimões, Amazonas, and Parnaíba basins

This basin is situated in the most remote region of the South American tropical forest. It is located between 2 and 8 degrees south latitude and 63 and 69 degrees west longitude and its shape is fan-like with the major width developed at the west and its depositional axis oriented NE-SW.

The Solimões Basin is bounded on the west by the Jutaí arch, on the east by the Purus high, on the north by the Precambrian Guyana shield and on the south by the Precambrian Guaporé shield. The Jutaí arch separates the Solimões Basin from the Subandean basins of Pastaza, Acre, and Madre de Dios.

In the past, Morales (1957, 1959, 1960) postulated that the Iquitos arch separated the Solimões Basin (upper Amazonas Basin) from the Subandean foreland basins. In the Iquitos area, Peru, however, there is a small high instead of an arch. The separation or division is formed near the town of Jutaí far away from the Iquitos region. Therefore in this study the Jutaí "Arch" is used instead of "Iquitos arch" for the divide between the Solimões and Andean pericratonic basins. The Purus high, in the eastern end of the Solimões Basin, separates it from the Amazonas Basin.

The geology of the Solimões Basin is poorly known and its stratigraphic column is not well established, thus the same column of the Amazonas Basin is used for it up to the present time.

However, there is no agreement concerning the formation and facies within both basins. It is of significance to revise the sedimentary succession in the Solimões Basin in order to erect an independent stratigraphic column for it, which can best fit the available geological data. Because the Paleozoic Solimões sediments are covered by Cenozoic strata, the proposed changes in the stratigraphic nomenclature are based only on subsurface data obtained from new wells drilled by Petrobras. Because geological studies are under way in order to locate more hydrocarbon, the geological framework of this remote basin will be better understood in the near future.

It is not possible to state whether or not all names used here will be those finally adopted. The first phase of oil exploration began in the late 1950´s, when gravity terrestrial surveys and stratigraphic drilling were carried out. A second exploration phase began after the late 1970's, when seismic reflection and exploratory wells succeeded in finding economic hydrocarbon accumulations.

In summary, the Siluro-Lower Carboniferous section in some parts of the Solimões Basin is different from and less complete than the section of the Amazonas Basin. The Permo-Carboniferous section is similar to that of the latter. The maximum sediment thickness, including Paleozoic, Mesozoic and Cenozoic rocks is about 3500 m. All transgressions came from the Andean area.

The Amazonas Basin (500,000 km^2) has its name derived from the Amazonas River and extends for 1400 km from the west of Manaus to near the town of Gurupá. The basin is elongate in shape, trending approximately in an east-northeast direction. It is about 300-500 km wide with the narrow part situated at the eastern end. It is located between 1 and 8 degrees south latitude and between 51º and 63º west longitude. The basin is bounded on the west by the Purus arch, on the east by the Gurupá arch, on the north by the Precambrian Guyana shield and on the south by the Precambrian Guaporé shield. The Gurupá arch separates this basin from the Marajó rift which originated possibly in Triassic-Jurassic times. Farther to the east Early and Mid-Paleozoic rocks are preserved in some down-faulted blocks inside the Marajó rift, indicating that the old Paleozoic Amazonas Basin originally extended beyond the Marajó Basin and the coast.

Two Paleozoic outcrop belts are present within the basin. The northern belt is 50-60 km wide and reaches about 1,000 km in length. The southern belt is 40-50 km wide and 700 km long. The sedimentary column ranges in thickness from 6,000 m in the central axial parts to 800-1400 m in the outcrop belts. The Paleozoic section is covered by Mesozoic and Cenozoic continental red beds on the flanks and central areas.

The geology of the Amazonas Basin is better known than that of the Solimões Basin and almost all available geological information from the entire Amazonas area was obtained from this region. The humid tropical climate of the area has generated a very thick soil and a dense forest, so that most of the field work has been done along river banks and river floors, where the outcrops show little alteration.

The Amazonas Basin has been studied since the early 1860´s, but more intensively since the early 1950´1s by the Conselho Nacional do Petróleo (National Petroleum Council) and Petrobras when systematic field geology and geophysical surveys were carried out and many boreholes were drilled. Considerable knowledge has been accumulating since then concerning the surface and subsurface geology.

From Late Ordovician to early Middle Silurian times, seas transgressed the region from the east. The Early Silurian benthonic and pelagic fauna has affinities with the North American and West-Saharan (Taoudeni Basin) faunas (Caster, 1952). The second great transgression took place in the late Early Devonian (Emsian Stage) and simultaneously brought to the Amazonas and southwest Taoudeni basins a fauna rich in North American elements (Hollard, 1967; Boucot, 1975). From the Late Carboniferous to Early Permian time the sea transgressed the area from the west, bringing a rich fauna with Andean affinities.

The current Amazonas Basin stratigraphic .column was erected by Caputo and others (1971, 1972), but some modifications introduced since then are incorporated here in this study.

The Parnaíba Basin (600,000 km^2), has its name derived from the principal river of the region, and is situated southeast of the mouth of the Amazonas River. At the present it is more or less circular or saucer-like in outline, and covers the greater parts of the states of the Maranhão, Piauí and smaller parts of the states of Pará, Goiás and Ceará. Many Paleozoic sedimentary remnants occur in grabens and in down-faulted areas between its eastern outcrop belt and the eastern tip of the South American continent near the Atlantic Ocean (Ghignone, 1972; Mabesoone, 1977, 1978). These remnants are situated in the states of Ceará, Paraíba, Pernambuco, Alagoas, Sergipe and Bahia. Some wells also detected Paleozoic sedimentary rocks in offshore and onshore rift basins along the north and northeast coast of Brazil. This indicates that the former basin, at least, was twice or three times as large as it is now and that its area probably extended to Africa in a pre-drift reconstruction of the Gondwana supercontinent.

The continuous part of the Parnaíba Basin is located between approximately 3º and 11ºS latitude and 40º and 48ºW longitude. It is separated from the Marajó graben on the northwest by the Tocantins arch, from the coastal São Luís and Barreirinhas grabens on the north by the Ferrer arch, and from the Paraná Basin by the São Francisco arch. The western and eastern margins are located on the Brazilian shield, by some named central and coastal shields respectively.

The Parnaíba Basin has a flat-lying stratigraphic section with a thickness of about 2,500 m of Paleozoic sediments and 500 m of Mesozoic beds in its central parts. The Paleozoic outcrop area is about 100 km wide along the western flank and about 300 km wide along the eastern side. The south, northwest and north edges are overlain by Mesozoic and Cenozoic beds.

The Marajó Basin, also called Marajó Graben, is considered as a single tectonic unit but it is composed of minor horsts and grabens such as the Mexiana, Limoeiro and Badajós grabens. The Marajó Graben lies to the east of the Gurupá arch close to the Amazonas River mouth and it is bounded by normal faults which developed in Late Paleozoic, Mesozoic and Cenozoic times. The Siluro-Devonian stratigraphic section is the same as that of the Amazonas Basin, but incomplete. Permo-Carboniferous rocks were found in small parts of the basin due to previous uplift of the area before rifting. The Mesozoic and Cenozoic sequence is about 4,000 to 5,000 m thick.

The rifted São Luís and Barreirinhas basins are located at the north of the Parnaíba Basin and the Urbano Santos and Ferrer arches separate the Parnaíba Basin from the northern coastal rifts. Paleozoic rocks may exist in the lower secttons of the Barreirinha and São Luís basins, but the basal sediments have not yet been drilled in deeper parts of these basins.

On the northeast coast of Brazil, a rifted basin named Sergipe-Alagoas is located between 9º and 12ºS latitude and 35º and 37ºW longitude. This coastal basin has a onshore area of 20,000 km^2 and a offshore area of about the same size. Its basal sequence is composed of Ordovician-Silurian and Permo-Carboniferous sediments, but most of the section is composed of Mesozoic and Cenozoic sediments deposited during phases of the opening and development of the Atlantic Ocean.

The Jatobá, Tucano and Recôncavo basins are also failed rifts which have not evolved to an oceanic phase. They are located in the continent close to the coast. The Jatobá and Tucano basins contain Paleozoic rocks in its basal section. Preserved sediments of Paleozoic age in the coastal rifted basins strongly suggests that the Parnaíba Basin may have extended beyond the present South American Atlantic shorelines.

At the south of the Solimões Basin, in the Ji-Paraná River headwaters, a graben occurs called Pimenta Bueno, oriented WNW-ESE with about 220 km in length and 40 km in width. This area is located about 800 km southward from the Solimões Basin center, in the State of Rondônia, close to the Bolivian border. Paleozoic rocks, some with glacial imprint and Mesozoic beds are present in the Pimenta Bueno Graben.

In this study the stratigraphy of the Solimões, Pimenta Bueno, Amazonas and Parnaíba basins is discussed in some detail. Some Paleozoic sections of the Sergipe-Alagoas and Jatobá basins are treated briefly in order to build up a paleoclimatic, paleogeographic and paleotectonic picture of northern Brazil.

CHAPTER 7. PIMENTA BUENO BASIN

In the Ji-Paraná River headwaters a graben called Pimenta Bueno trends NNE-SSW. It is about 40 km in width and 230 km in length, is located in a very remote area, in the State of Rondônia, close to the Bolivian border.

The stratigraphic column of sediments within this graben and its vicinity is here described based on the available literature because many of the formations exhibit characteristics of paleo-climatic and paleogeographic importance (Figure 16). The stratigraphic units were identified along a strip of about 230 km by about 35 or 40 km, between latitude 10ºS to 12ºS and longitude 60º 30`W to 62º 45´W (Carvalho and others, 1975).

In the State of Mato Grosso do Norte, south of the Pimenta Bueno graben, a similar rock succession is also present with the same characteristics and age as those of the Pimenta Bueno Basin (Olivatti and Ribeiro, 1973, 1976). The strata were identified in the Jauru valley extending over an area of about 600 km^2. The location is approximately latitude 16º S and longitude 59º W. If both outcrop areas in the states of Rondônia and Mato Grosso do Norte are connected, the distribution of these formations is over 700 km in length. The stratigraphic descriptions are based on the literature, but some tectonic, paleogeographic and paleoclimatic interpretations are made by me. The stratigraphic column of the Pimenta Bueno Basin is shown in the figure 17.

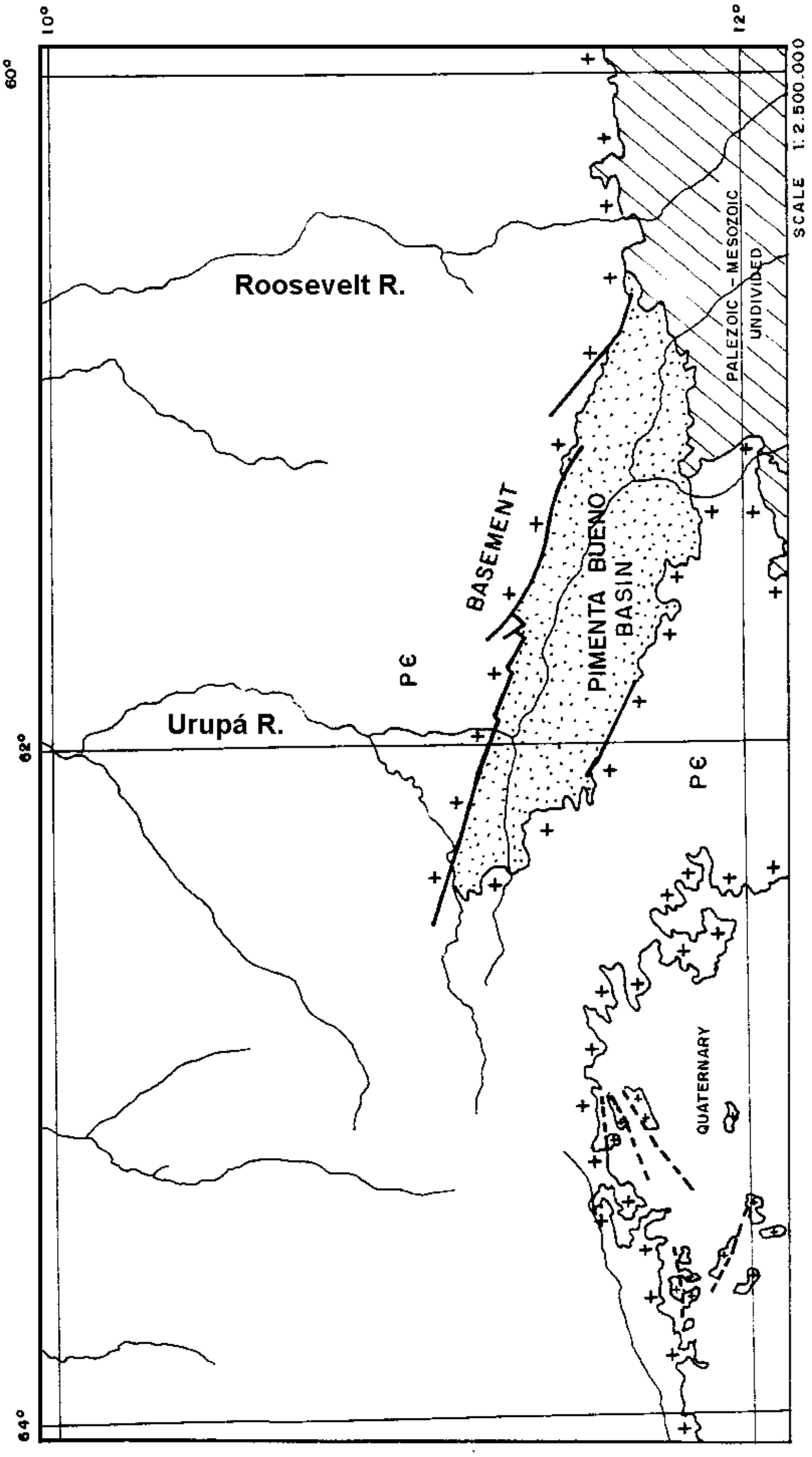


Figure-16 Pimenta Bueno Basin Index Map

PIMENTA BUENO FORMATION

The Pimenta Bueno Formation, up to 150 m thick, was proposed by Leal and others (1978) to describe a section of supposed Late Precambrian-Early Paleozoic age occurring in the Pimenta Bueno Graben, composed of shale, conglomerate, sandstone, diamictite, and coal lenses. However a Late Precambrian-Early Paleozoic age is incompatible with the presence of plant remains and coal, because land plants are not known to exist before Paleozoic times. According to Pinto Filho and others (1977) the Pimenta Bueno Formation consists of chocolate brown, brown red and greenish-gray laminated commonly, fissil, hard, micaceous shale at the base of the section.

In the middle part of the section subrounded, micaceous, feldspathic brown to cream, fine- to medium-grained sandstone beds are present with slump structures and clay balls. Disseminated pyrite and vegetal remains occur throughout the unit. The sandstone beds are relatively rich in pollen, while the shale beds are generally barren. Fossil patterned polygonal ground is widespread on the sandstone beds.

The upper section is composed of mudstone, diamictite, rhyth-mite, conglomeratic arkoses and sandstone beds. The diamictites are nonsorted and structureless, and carry red to brown supporting

CHRONOSTRATIGRAPHIC COLUMN			LITHOSTRATIGRAPHIC COLUMN	GENERAL ROCK DESCRIPTION
ERATHEM	SYSTEM	SERIES	FORMATION	
MESOZOIC	CRETACEOUS	LATE	PARECIS	*Fine-to coarse grained, variegated sandstone and conglomerate beds.*
		EARLY	BOTUCATU	*Fine-to medium grained, feldspathic, subrounded to rounded, well-sorted large-scale cross-bedded sandstone beds.*
	JURASSIC	LATE		
	TRIASSIC		ANARI BASALT	*Black to dark-gray, fine-grained to aphanitic, brecciated, basalt and with fluidal texture.*
PALEOZOIC	PERMIAN	EARLY	FAZENDA DA CASA BRANCA	*Sandstone, feldspathic sandstone, graywacke, siltstone, mudstone, diamictite and conglomerate beds.*
	CARBONIFEROUS	LATE		
			UNNAMED LIMESTONE BEDS	*Yellow to gray fine-grained limestone beds, fossiliferous.*
		EARLY	PIMENTA BUENO	*Redish brown to gray shale, conglomerate, sandstone, diamictite beds with coal streaks.*

Figure-17 Provisional Phanerozoic stratigraphic column of the Pimenta Bueno Basin clasts, inserted in a clay-sandy matrix. Many of the clasts are faceted, polished and striated, and range from cobble- to boulder-size (up to 1.5 m in diameter). Several clasts are surrounded by a calcite veneer and some small ones display a flatiron shape. The clasts are composed of quartzite, acid volcanic rocks, gneiss, metabasite, metasiltstone, rnetasandstone, limestone and granite. Laminated, red to light gray claystone beds, with slump structures, also contain dropstones which disturb the bedding beneath them. Alternated brown claystone and green siltstone with an average lamina thickness of about 2 mm have been interpreted as varves. Some calcite laminae are associated with rhythmite beds.

Despite the presence of Carboniferous palynomorphs and plant remains, Leal and others (1978), De Montalvão and others (1979) and De Montalvão and Bezerra (1980) assigned a Late Precambrian to Early Paleozoic age to the sediments corresponding to the Pimenta Bueno Formation.

Pinto Filho and others (1977) attributed a Permo-Carboniferous age age to the unit, based on palynomorphs identified by Palma (in Pinto Filho and others, 1977). Lima (1982, written communication), from the Petrobras paleolab analyzed the palynological content listed by Palma and concluded that the age of the section is Early Carboniferous.

The palynoforms identified are the following:

Retialetes sp., *Raistrickia* sp. *Azonotriletes* sp. *Cristalisporites* sp *Monoletes* sp., *Convolutispora* sp., *Acantho-Triletes* sp., *Verrucosisporites* sp., *Punctatisporites* sp., *Pilas-Porites* sp., and *Plicatipollenites* sp.

In the Amazonas and Parnafba basins some diamictitic beds in the Visean Stage (mid-Early Carboniferous) are present with a similar palynological content. If these beds do indeed correlate, a Visean age would be assigned to the Pimenta Bueno glacial deposits.

Nahass and others (1975) first raised the hypothesis that the Lower Pimenta Bueno section was deposited under glacial conditions. Pinto Filho and others (1977) also interpreted the section that is informally named Permo-Carboniferous II (upper part of the Pimenta Bueno Formation) as glaciogenic. Leal and others (1978) rejected a glacial origin for the unit and favored the deposition of the Pimenta Bueno Formation during the tectonism that generated the graben in Precambrian time.

Away from the graben, in the State of Mato Grosso do Norte and in the south of the State of Rondônia, a unit, named Jauru Formation, with the same characteristics as those of the Pimenta Bueno Formation was identified at the top of the section. The upper part of the Jauru section, up to 50 m thick, is considered to have been deposited in the Early Carboniferous time by Olivatti and Ribeiro (1973, 1976}. Rocha-Campos (1981a) also supports a glacial nature of the Jauru and Pimenta Bueno formations.

The upper portion of the Pimenta Bueno Formation was undoubtedly laid down in a glacial environment represented by tillites with striated and faceted clasts, and in a proglacial environment represented by claystone with dropstones and varves, periglacial and outwash deposits. The direction of ice flow was determined by Pinto Filho and others (1977) as being from northeast to southwest; however, these authors did not explain how these measurements were made.

The presence of widespread sediments of a glacial nature, in the states of Rondônia and Mato Grosso do Norte, suggests that the Early Carboniferous glaciation covered large areas of northern Argentina and Bolivia (Frakes and Crowell, 1969; Amos, 1972; Ayaviri, 1972; and Reyes, 1972) and the Guaporé shield. There is also some evidence that the Amazonas and Parnaíba basins were glaciated. The idea that the Early Carboniferous glaciation in Bolivia and northern Argentina was exclusively of mountain-type is no longer supported because the glacial sediments were laid down in shield areas and in low-lying intracratonic sedimentary basins.

The Early Carboniferous Brazilian glacial rocks have the same characteristics as those of the Bolivian formations. The red color and smooth veneer of light gray calcite is common to both areas. In Bolivia, the red color observed in tillites was attributed to climate diagenesis (Helwig, 1972). The calcite veneer on some Pimenta Bueno tillite clasts from the basin are similar to the calcite veneer found in Antarctic till clasts (Nicholson, 1964). If the area of Rondônia was glaciated, the Sernambi Formation of the Solimões Basin may be a result of glaciofluvial activity at the periphery of glaciers.

The general idea is that the Pimenta Bueno graben was formed just prior to (Pinto Filho and others, 1977) or simultaneously with (De Montalvão and others, 1979; Leal and others, 1978) the deposition of the Pimenta Bueno Formation. However, I envisage the Pimenta Bueno graben as forming during Mesozoic time when old weak basement zones were reactivated during the rupture of the Pangea continent. The Pimenta Bueno Formation shows many joints in the same direction as those of bounding faults, that is, E-W, WNW-ESE, with dips as high as 90º. A joint system trending N40ºW and N50ºE is also present. The Pimenta Bueno beds are faulted against the Precambrian Uatumã Group; but the faults obviously should be younger than these lower Early Carboniferous beds. The presence of the Early Carboniferous glacial rocks outside the graben indicates that the coarse diamictites were deposited not only in the area of the graben, but also on the Guaporé shield. The graben did not exist when the tillites were deposited.

UNNAMED FORMATION

A fusulinid-bearing limestone unit overlies the Pimenta Bueno Formation, but its thickness and extent are unknown. This unit is tentatively correlated with the Itaituba (Amazonas Basin) or Tarma (Peru) formations of Late Carboniferous age.

FAZENDA DA CASA BRANCA FORMATION

The term Fazenda da Casa Branca Formation was proposed by Leal and others (1978) to designate Permo-Carboniferous sandstone beds which overlie the Pimenta Bueno Formation and the fusulinid-bearing formation. The unit was first described by Padilha and others (1974) in the states of Rondônia and Mato Grosso do Norte. They named it Eopaleozóico Indiviso (Undivided Eopaleozoic). Later, when Permo-Carboniferous plant remains were found in the unit, it was renamed Permo-Carboniferous I unit (Olivatti and Ribeiro Filho, 1976) or Permo-Carboniferous III unit (Pinto Filho and others, 1976). Here the unit is designated Fazenda da Casa Branca Formation following Leal and others (1978).

According to Pinto Filho and others (1976) the unit, up to 210 m thick, consists of sandstone, feldspathic sandstone, graywacke, siltstone, mudstone, diamictite and conglomerate beds. The lower part of the section contains pink, white, yellow, poorly sorted, subangular cross-bedded sandstone beds, with sparse granules and cobbles. Tabular cross-bedding and cut. and fill structures are also common.

Upwards the sandstone is more feldspathic, and in some places, the unit is a brown-red greywacke with angular grains that are poorly sorted and conglomeratic. Brick-red siltstone and mudstone beds also occur, together with red to violet diamictite interbeds. The diamictite has a clayey matrix, supporting granite, quartzite, gneiss, and quartz pebbles.

The top of the section consists of variegated and red mudstone with green and small quartz and chert pebbles.

The Fazenda da Casa Branca Formation was considered by Pinto Filho and others (1977) as lying conformably over the Pimenta Bueno Formation, but Leal and others (1978) stressed that it overlies unconformably the Pimenta Bueno Formation and older rocks. The presence of *Psaronius sp*. Indicates a Latest Carboniferous-Early Permian age for the unit, which suggests the existence of an unconformity between the Early Carboniferous Pimenta Bueno Formation and the Fazenda da Casa Branca Formation.

Pinto Filho and others (1977) correlated the Fazenda da Casa Branca Formation with the Aquidauana Formation of the northwestern part of the Paraná Basin because both formations show similar rock characteristics. The age of the Aquidauana Formation was determined by Daemon and Quadros (1970) as Stephanian (late Late Carboniferous on the basis of palynological studies

(Biostratigraphic interval G of the Paraná Basin). The Aquidauana Formation is synchronous to and interfingers with the basal part of the glacial Itararé Group (Campo do Tenente Formation) according to Schneider and others (1976).

The Fazenda da Casa Branca Formation may have been deposited in glacial and periglacial environments. The presence of Psaronius sp. (plant) in the unit indicates that a climatic amelioration occurred after the deposition of tillite beds.

The paleoclimatology of western Brazil in Latest Carboniferous time can now be interpreted. From the northeastern part of the State of São Paulo southwards, glacial conditions prevailed; from northeastern São Paulo northward to the State of Rondônia glacial, and mainly periglacial conditions predominated; and in the northern Solimões and Amazonas basins, arid dry conditions dominated with significant cyclic evaporitic deposition (Nova Olinda Formation). The ice-cap edges apparently migrated from the Solimões Basin to the northern Paraná Basin from Late Devonian to Late Carboniferous time.

The existence of carbonate rocks of Late Carboniferous-Early Permian age overlying Early Carboniferous glacial rocks in extreme northwest Bolivia and Peru (Helwig, 1972) and the presence of Late Carboniferous-Early Permian glacial rocks in the Paraná Basin reinforces the idea of southeastward intermittent migration.

It is appropriate to introduce here the concept of paraglacial activity, first defined by Ryder (1971a, b), to designate nonglacial processes that are directly conditioned by glaciation. It refers both to proglacial processes, and to those processes occurring around and within the margins of a former glacier (Church and Ryder, 1972). The paraglacial period encompasses the time during which periglacial processes occur. For example, glaciation produces a large amount of detrital material (Corbel, 1964) which is stable in the glacial environment, but is not stable with respect to the succeeding fluvial environment. The removal of the material will continue after the ice melts. The sediment yield in this case has no relation to the production of weathered material.

The great amount of river load under these conditions is able to build alluvial cones and fans, and ultimately, to provide coarse clastics to the depositional basins. If the climatic amelioration in the area is strong, unstable drift material exposed to weathering processes may be easily eroded and oxidized in broad shield areas. If a new glacial episode develops in the area, the debris incorporated in the ice-caps may be red beds. This process may explain the presence of red tillites in Carboniferous times as originating from the alteration of debris left by earlier glaciations in shield areas.

ANARI BASALT

Pinto Filho and others (1977) proposed the name Anari Basalt for basic rocks occurring in the Pimenta Bueno Basin. These rocks were first described by Moritz (1916). The basalt is black to gray, fine-grained to aphanitic with brecciated and fluidal textures and shows fractures trending N50ºW. The age of the basalt is a controversial issue. It may be as old as 208 ± 14 m.y. (Triassic) or, according to another dating, as old as 111 ± 8 m.y. (Early Cretaceous) based on the K/Ar method (Pinto Filho and others, 1977). There is even the possibility of two basalt flows of distinctly different ages occurring in the area. Triassic basalts are present in the Acre Basin and about the north Brazilian coast, and Cretaceous basalts are common in the Paraná Basin.

The Anari basalt overlies unconformably the Fazenda da Casa Branca Formation and the Uatumã Group and it is probably underlain by the Botucatu Formation sandstone. The age of the basalts may mark the time of the Pimenta Bueno graben tectonism.

BOTUCATU FORMATION

In the State of Rondônia, Pinto Filho and others (1977) mapped a sandstone unit with the same aeolian characteristics as those of the Botucatu Formation of the Paraná Basin. Due to this similarity they decided to maintain the name Botucatu Formation for the aeolian sandstones.

The unit is up to 90 m in thickness and consists of fine- to medium-grained, feldspathic, subrounded to rounded, well sorted, large-scale cross-stratified, red to yellow sandstone beds with some frosted grains. Red clayey sandstone beds are common and ventifacts were observed in some sandstone units.

The Botucatu Formation overlies unconformably the Fazenda da Casa Branca Formation, but its stratigraphic relationship with the Anari basalt is not known. The Botucatu Formation underlies unconformably the Parecis Formation. This unit shows very sparse fractures without a defined trend. The presence of poorly oriented joint sets may indicate that the Botucatu Formation was laid down after (?) the Anari basalt extrusions, that is, after a period of tectonic calm.

The age of the unit was inferred on the basis of the age of the Botucatu Formation in the Paraná basin as Late Jurassic/Early Cretaceous.

The environment of deposition of the Botucatu Formation in the Pimenta Bueno Basin is aeolian, as deduced by large-scale cross-bedding (2-4 m high or more) and ventifacts. A fluvial contribution is also identified due to the presence of cross-bedded medium-grained sandstone beds rich in argillaceous matrix. The presence of Aeolian sandstone in the area suggests that it was laid down in the trade wind belt of high evaporation in low latitudes.

PARECIS FORMATION

De Oliveira (1915) proposed the term Parecis Formation to describe a Cretaceous section composed of sandstone and conglomerate beds. The unit consists of fine- to coarse-grained, sandstones with well rounded grains. Their beds are cross- and parallel-stratified, silicified, and are red, white and yellow with conglomerate interbeds occurring primarily at the base of the succession. Some conglomerate beds contain basalt pebbles, probably derived from the Anari Formation (De Oliveira, 1936).

The Parecis Formation overlies unconformably the Anari, Botucatu and Casa Branca formations and Precambrian Uatumã Group and is overlain by Cenozoic lateritic crusts. De Oliveira (1915) assigned a Late Cretaceous age to the unit on the basis of the presence of dicotyledon stems. In the past, many rock formations overlying the Uatumã Group were correlated with the Parecis Formation, but this correlation was proven to be incorrect because most of these rocks were laid down in Precambrian times. The unit represents a fluvial environment of deposition under moist conditions.

Petri and Campanha (1981) pointed out that paleocurrent measurements in the Parecis Formation indicates a northward drainage which may have been connected to Cretaceous drainage in the Solimões Basin. The formation thickness also increases northward in the direction of this basin.

In Cretaceous times, a climatic amelioration, due to the presence of intervening oceans after the break up of Gondwana probably favored rainfall in areas that prior to continental rupture were far away from the sources of moisture. In Cenozoic time, due to the equatorial position of the area, intense laterization took place with the formation of lateritic crusts as thick as 30 m on the Parecis Formation surface. This lateritic deposit is a consequence of extremely moist conditions prevailing in the climate of equatorial Cenozoic forests.

CHAPTER 8. SOLIMÕES BASIN

The Solimões Basin was first studied on a regional scale by Morales (1957). He combined both gravity and seismic information from the area with geological data from the Andean and Amazonas basins available at that time in order to predict its stratigraphy. He proposed three wells in order to establish the sedimentary sequence of the basin and its oil possibilities. Later, after the wells had been drilled, Morales (1959) presented the rock succession of the basin at the 5th World Geological Congress (New York).

Bouman and others (1960) studied the Amazonas and the Solimões basins, and on the basis of stratigraphic information acquired from more test wells they reported on the geology of the area. They retained the same column prepared by Morales (1959), which was based on the Amazonas sedimentary succession. By this time, the geological outline of the basin was established.

The exploratory efforts diminished in the area in the 1960´s due to the discovery of more promising new areas for oil exploration. Renewed exploratory efforts started in the late 1970's, when more wells and geological knowledge provided the means for a better definition of the formations.

Many stratigraphic sections were described by me based on bore holes drilled by Petrobras. Cenozoic and Mesozoic cover does not allow to observe the section ABC on surface. Figure 18 shows cross-

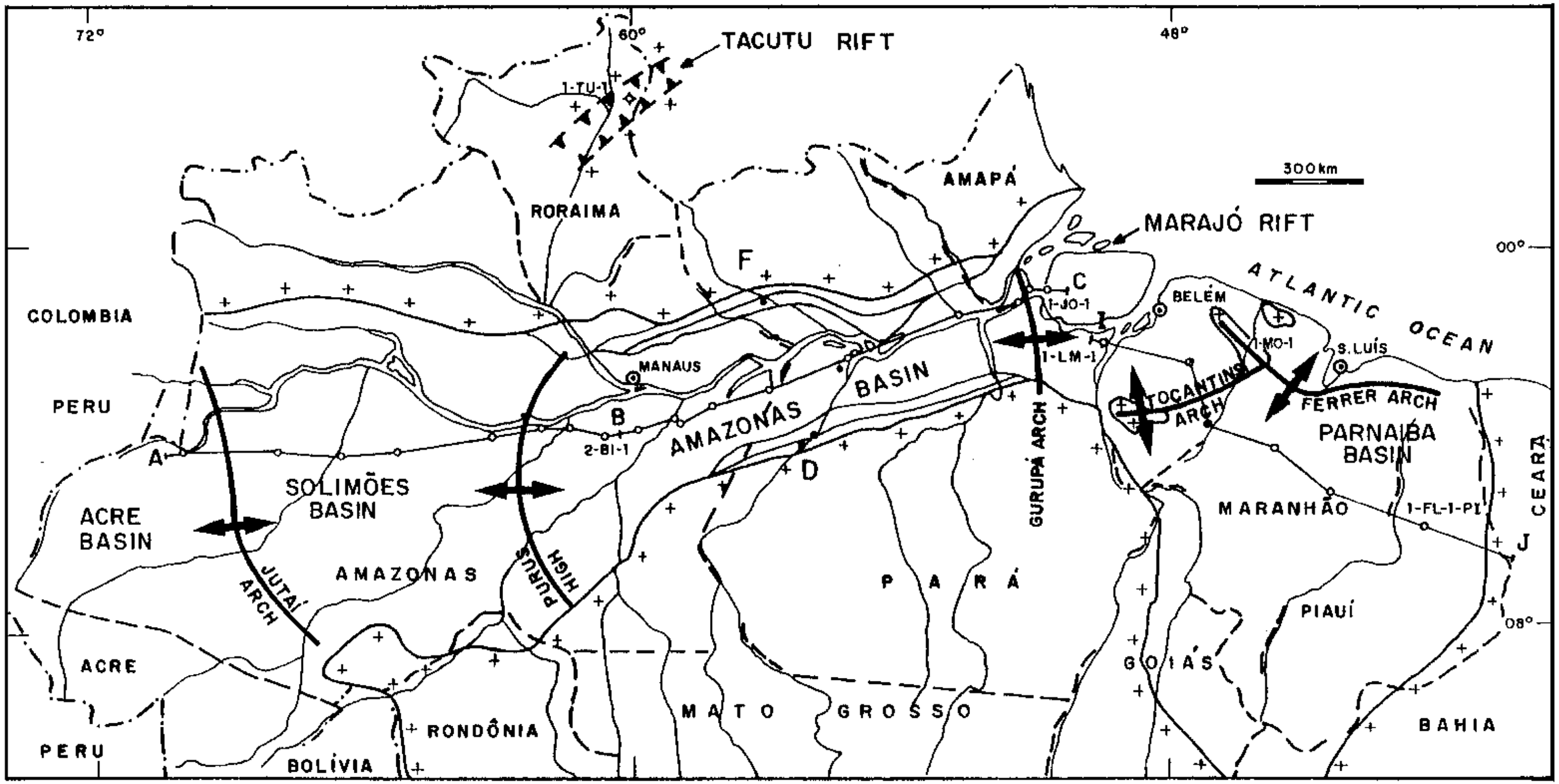


Figure- 18 Major cross section Lines (A– B) throughout Northern Brazilian Paleozoic Basin.

section lines of northern Brazilian basins; figure 19 shows the stratigraphic column proposed for the Solimões Basin; figure 20 shows the contour map of the Precambrian basement and figure 21 shows a cross-section along the Solimões Basin.

The stratigraphy of the sedimentary rocks are here described by me in the following section on the basis of bore hole cores, cuttings and electrical logs.

UATUMÃ GROUP

De Oliveira and Leonardos (1943) gave the name Uatumã Series to an assemblage of basement volcanic rocks, described by Albuquerque (1922), which is exposed along the Uatumã River, 65 km upstream from the Balbina Falls, in the north of the Amazonas Basin. These rocks were cored in Solimões Basin bore holes. For some time, the Uatumã Group was considered as a sequence of sandstone and siltstone beds (Bouman and others, 1960). This sedimentary sequence was later recognized by Caputo and others (1971) as the Prosperança Formation of De Paiva (1929) which overlies unconformably the Uatumã Group considered the basement of the Amazonas Basin. Barbosa (1967) changed the name for the volcanic rocks from Uatumã Series to Uatumã Group. Caputo and others (1971) also used the name Uatumã Group to describe volcanic and plutonic rocks that are found in the Guyana and Guaporé shields as well as in

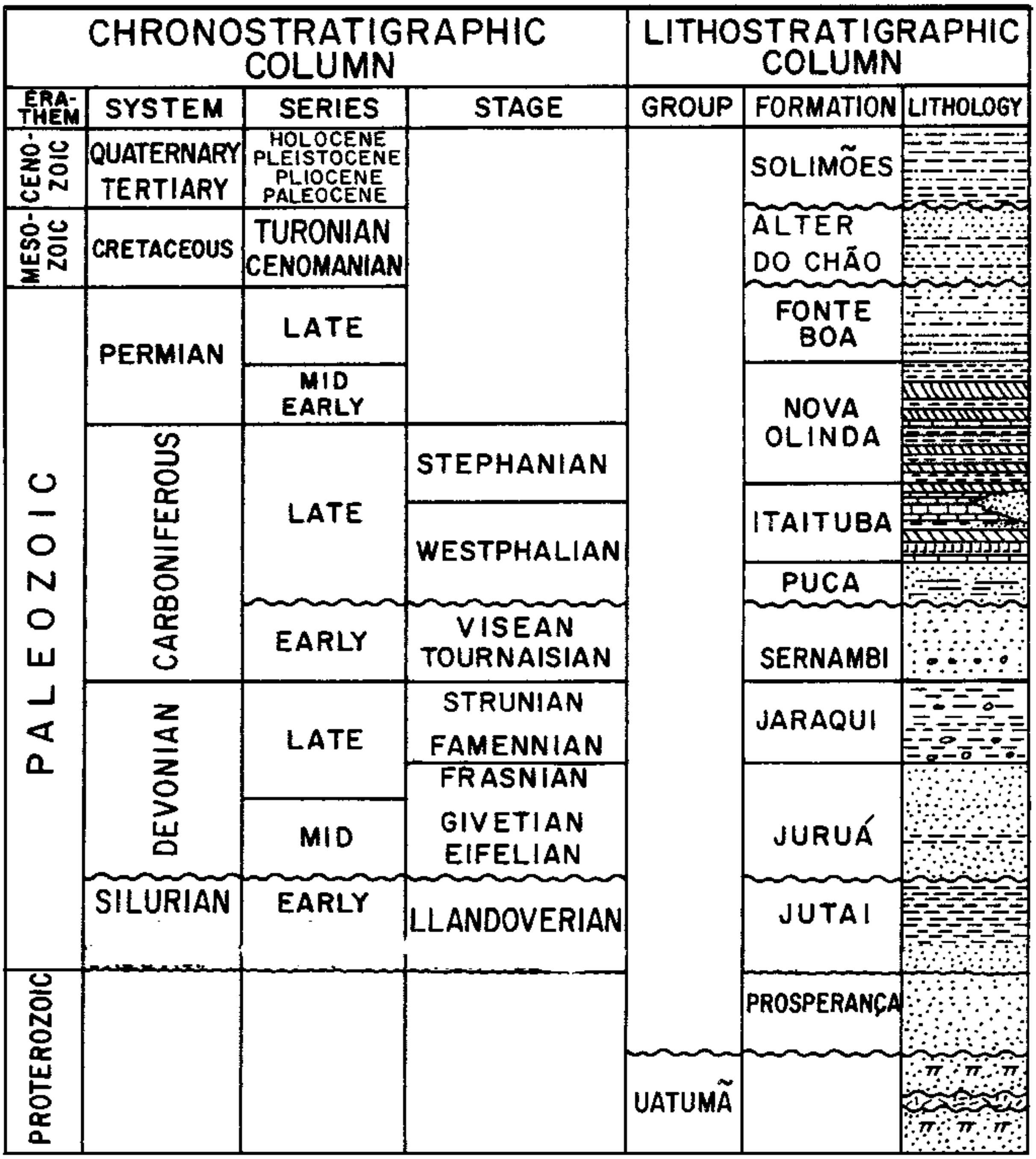


Figure-19 Stratigraphic column of the Solimões Basin some Amazonas and Solimões basement well cores.

The Uatumã Group is defined as a widespread volcano-sedimentary sequence containing andesites, dacites, rhyolites, rhyodacites, ignimbrites, crystal tuffs, lithic tuffs, vitric tuffs and flow breccias, as well as associated granitic, grano-dioritic and syenitic subvolcanic circular cratogenic intrusions. Detailed descriptions of each of these rock types occurring in the Guaporé and Guyana shields are provided by De Montalvão (1976), Issler (1977), and Rangrab and Santos (1976).

The Uatumã Group crops out in topographic and structural lows of the crystalline complex, and its rock succession and thickness change from place to place.

Few of the many bore holes drilled in the basin, have reached the deep Uatumã Group. The cores of the Uatumã Group from the stratigraphic wells 2-FG-1-AM, 2-JA-1-AM, 2-JT-1-AM, 2-RE-1-AC and 2-BT-1-AM have yielded an average age of 1,560 m.y., which correlates with age data obtained on granitic rocks located at the Guyana and Guaporé shields (Kovach and others, 1976).

PROSPERANÇA FORMATION

The Prosperança Formation comprises assemblages of sandstone, siltstone, shale, mudstone and conglomerate. It crops out in the eastern part of the Solimões Basin and in the western

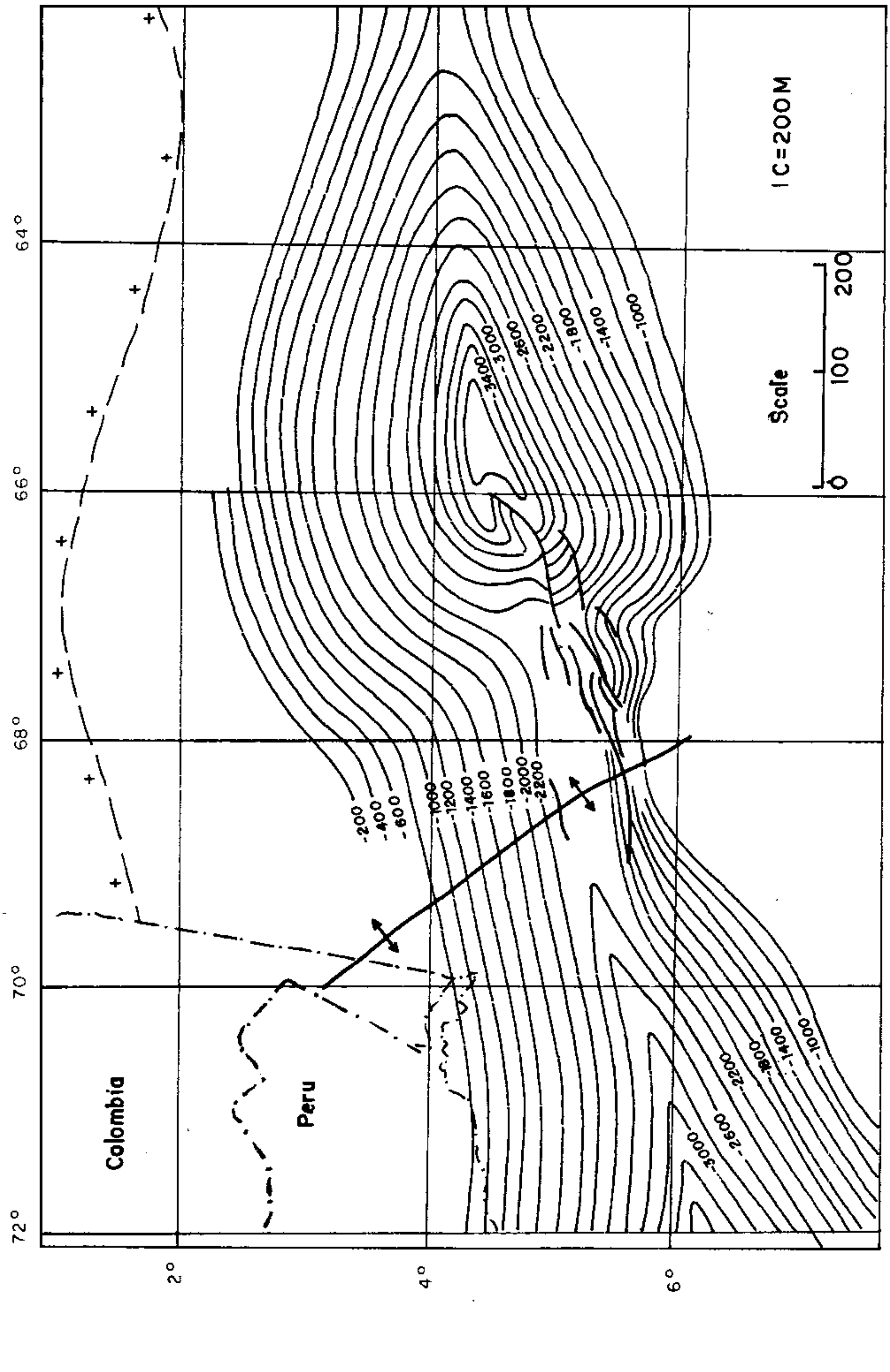


Figure 20 Structural contour of the Precambrian basement - Solimões Basin

part of the Amazonas Basin. De Paiva (1929) named the unit while studying it in the Igarapé Prosperança (Prosperança Creek) and Negro River area. Later, several occurrences of these sandstone and siltstone beds were given different designations. Caputo and others (1971) reviewed the Amazonas Basin stratigraphic column, correlated the different rock exposures, and formally redefined the Prosperança Formation to encompass all occurrences with the same appearance and stratigraphic position. The formation is sub-divided here into three main units.

In the lower sequence, at the base of the section, occur conglomerate beds, composed of quartz, rhyolite and quartzite pebbles. The best exposures are found along the south flank of the Amazonas Basin. The conglomeratic section is followed by cross-bedded, yellowish white, hard, fine- to coarse-grained, subangular to subrounded, partially silicified sandstone. The second unit is comprised of red and cream-colored shale and siltstone beds with sandstone interbeds. The upper unit is composed of cross-bedded white and red brown fine- to coarse-grained sandstone, followed by a red shale and a siltstone sequence, with some limestone interbeds near the top and middle of the unit.

The Prosperança Formation overlies unconformably the Uatumã Group, as observed in the outcrop area. In the Solimões Basin, a major erosional unconformity separates the Prosperança Formation

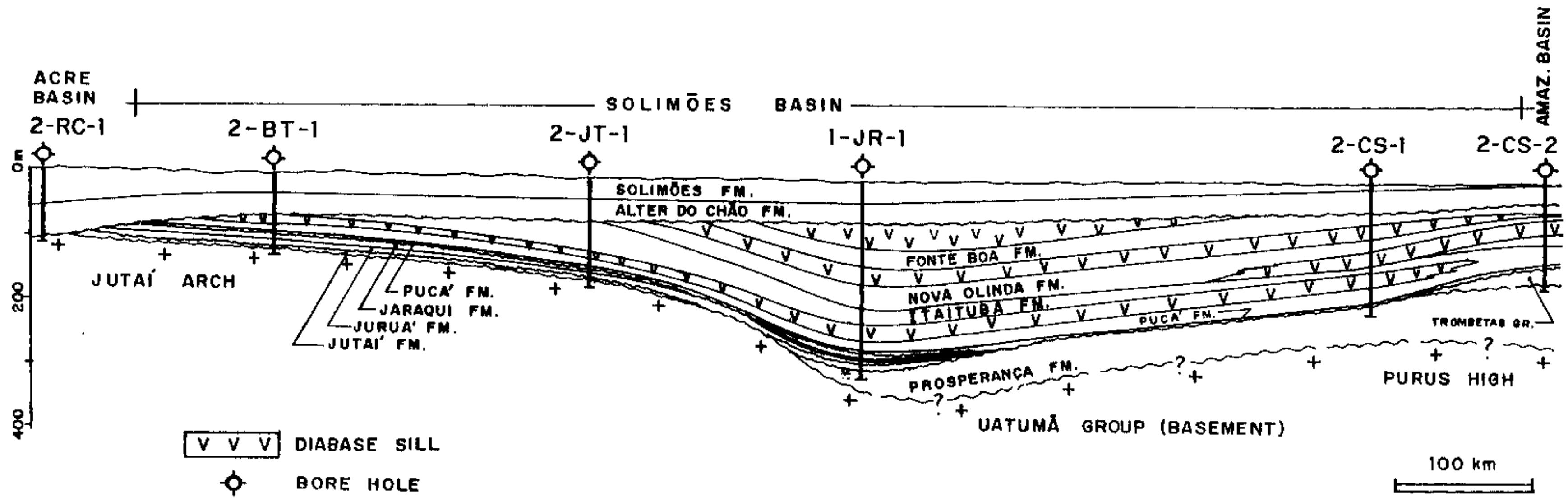


Figure-21 Solimões Basin longitudinal cross-section from the overlying Devonian and Carboniferous formations. Its maximum thickness is not known in the Solimões Basin, but is estimated to be over 1,000 m in the Purus high. The Prosperança Formation reflects a fluvial environment with several fining-upward sequences. Fine-grained sandstone and siltstone with dolomitic limestone interbeds may indicate marine conditions of deposition.

The age of the formation is not known. It is intruded by 1,400 and 1,100 m.y. old diabase dikes (De Montalvão and Bezerra, 1980) and overlies the Uatumã Group. No fossils are recorded in the unit, but structures like trace fossils were recorded by Caputo and Sad (1974) close to the basement, in the Negro River area.

JUTAÍ FORMATION

The name Jutaí Formation is proposed here to describe a rock assemblage composed of light-gray sandstone and dark-gray siltstone beds and shale beds of Silurian age. Its name is derived from the well 2-JT-1-AM (Jutaí stratigraphic well number 1, Amazonas State), and its type-section is located at the 1,483-1,573 m interval with a thickness of 90 m. It was previously called Trombetas Formation (Rodrigues and others, 1971) under the assumption that it was connected with the upper half part of the Trombetas Formation (now Trombetas Group) of the Amazonas Basin.

However, the supposed connection may have not existed.

I describe the section based on bore holes; no outcrops are present.

The base of the section consists of 15 m of thickly bedded white, fine-grained, hard, laminated sandstone, overlain by 27 m of dark gray to black carbonaceous, thinly bedded shaley siltstone with fine-grained sandstone interbeds. Stratigraphically upward one encounters an 18 m thick white to dark, hard pyritic fine-grained quartzite, followed by 15 m of thick dark-gray to black laminated soft, pyritic, carbonaceous shale with arenaceous laminations. The top of the section consists of white to light-gray, fine- to very fine-grained, moderately indurated, laminated sandstone beds. It should be pointed out that there are only three wells serving as control points, and consequently, the environmental interpretation in this unit is very tentative. The age of the Jutaí Formation was determined to be Early Llandoverian (local Biostratigraphic interval III) by palynological studies made by Daemon and Contreiras (1971a,b). No macrofossils were found in the formation. The Jutaí Formation, which is the lowermost of the Paleozoic section, rests unconformably on the Uatumã Group and older rocks. The Prosperança Formation is absent in the area of occurrence of the Jutaí Formation. The Jutaí Formation is unconformably overlain by the Juruá Formation of Devonian age, and the

Puca Formation of Late Carboniferous age.

It is difficult to determine the initial geometry of the Basin, in which the Jutaí Formation was deposited because the unit has only been identified in three wells. Seismic lines made for Petrobras has revealed many reflectors below the Cretaceous cover towards the southwest (Acre Basin) where the unit may occur and become thicker. The formation is absent in wells drilled about 200 km to the east. It is difficult to specify the nature of the paleogeographic setting elsewhere in the Solimões Basin, during Jutaí time because there is a general lack of sedimentary data in the east part of the basin for this time. In addition, this suggests that prior to the deposition of Devonian sediments, the eastern part of the basin was mainly an area of erosion, and therefore, elevated land. I believe that only the western part of the basin was invaded by the sea during the peak of the worldwide Llandoverian (Early Silurian) transgression, in contrast to the Trombetas Group of the Amazonas Basin where the sedimentary cycle started earlier and was more complete.

The graptolite-bearing Pitinga Formation (lower unit III), which records the peak of the Early Silurian transgression in the Amazonas Basin, pinches out at the Purus high eastern flank (Figure 16), reflecting a westward facies-change in the Amazon Basin from a shallow marine environment to littoral and continental environments. At that time, the

Purus high apex, named Coari high, was located at least 100 Km to the west (De Boer, 1964; Rodrigues and others, 1971), so it may have constituted a barrier to the invasion of the sea from the Amazonas Basin into the Solimñes Basin. If this view holds, it can be inferred that the Jutaí Formation was laid down by a sea connected only to the Andean basins (Paleopacific Ocean).

On the other hand, Szatmari and others (1975) argued that sediments of the same Carboniferous age were found in the Solimões and Amazonas basins, which suggests that these basins should have been invaded by the same sea. They also stressed that the deepest part of the basin, where Silurian sediments could be found, had not been tested by drillIng.

Nevertheless, since 1977, several wells have been drilled in the present deepest parts of the basin. Such wells demonstrate the absence of any Silurian sediments in its central and eastern areas. This fact weakened the idea suggested by Szatmari and others (1975). It seems that the marine transgression only reached the western and central parts of the Solimões Basin, and later an uplift in its western side (Jutaí arch) generated a structural deep beyond the eastern edges of the former Silurian depositional area. During Early Silurian time, a wide sea portal opened across Ecuador and northern Peru (Harrington, 1962), close to the Jutaí deposits. Therefore, there are indications of a possible sea connection between the Solimões Basin and Andean troughs.

JURUÁ FORMATION

I propose the term Juruá Formation in this study to describe a sandstone formation occurring at the base of the Devonian section in the Solimões Basin. The figure 22 shows the isopach contour map of the Devonian-Early Carboniferous section (Juruá, Jaraqui and Sernambi formations). The unit consists of white and light gray, hard, well cemented, subangular, cross-bedded and laminated fine-grained sandstone with few thin shale and siltstone interbeds. Its name originates from the stratigraphic well Juruá number 1 (2-JA-1-AM) in the State of Amazonas, and its type-section is found at the 2,788-2,916 m interval with a thickness of 128 m. It was previously named Maecuru Formation (Morales, 1959; Bouman and others, 1960) or Ererê (Rodrigues and others, 1971) because it was considered to be connected with the lower or middle Devonian sediments of the Amazonas Basin which occur 350 km further east.

The Juruá Formation rests unconformably on the Uatumã Group and older rocks in the north and south edges of the basin, and overlies the Prosperança Formation with low angular unconformity in the eastern part of the basin, as determined by dipmeter logs. It also overlies unconformably the Jutaí Formation in the

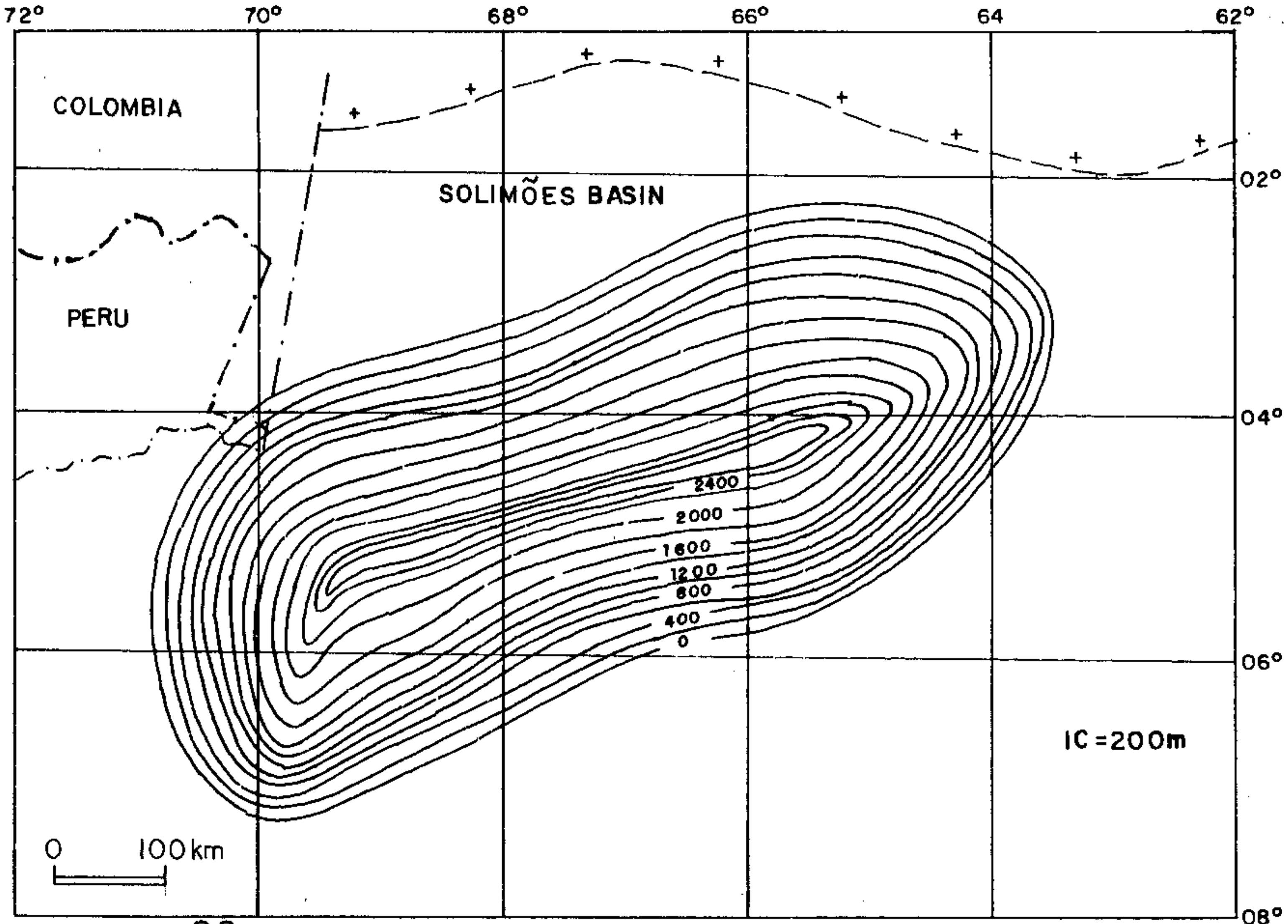


Figure 22 Isopach map of Juruá, Jaraqui and Sernambi Formations Solimões Basin.

western side of the basin. It has been shown that, after sedimentation had commenced, younger horizons of the unit were deposited successively further east, overlapping the Jutaí Formation and older Juruá beds.

The basin in which the Juruá Formation accumulated was larger than the Jutaí Basin. A regional reconstruction of the depositional basin of the Juruá sequence involves wide extrapolation because the control points are widely spaced and frequently inadequately distributed.

From Mid-Silurian to Early Devonian time a worldwide marine regression took place (Hallam, 1977) which caused a break in the sedimentation in the Solimões Basin which lasted for more than 50 m.y. The regression lasted from Late Llandoverian (Early Silurian) to Eifelian (Mid-Devonian) time and its peak occurred at the end of the Mid-Silurian time, when the Andean troughs were also abandoned by the sea (Harrington, 1962).

The general sedimentological picture suggested by these sediments Is of a transgressive situation in which alternation between beach-type, deltaic, and shallow marine sedimentation took place. It seems that the rate of downwarping in the present basin center was low, about 130 m, if deposition took place between the Eifelian and Late Frasnian times.

The sea's transgression in the Amazonas Basin began earlier

(Emsian Stage - late Early Devonian) than it did in the Solimões Basin (Eifelian Stage-early Middle Devonian). The unit started its deposition in Eifelian and terminated in Late Frasnian times (early Mid-Devonian to early Late Devonian).

The Juruá Formation is synchonous with the Ererê and Barreirinha formations of the Amazonas Basin. At the present Purus high, in the Amazonas Basin, the Ererê Formation is made up of littoral and shoreface sediments which thin towards the arch. This marginal environment indicates the proximity of the western limit of the Amazonas Basin during Eifelian and Givetian times (Mid-Devonian). The Coari arch seems to have remained emergent during Eifelian and Givetian times, preventing any connection between the Amazonas and Solimões basins.

Whether the Barreirinha Formation, which records the maximum Devonian transgression in the Amazonas Basin had any connection with the Solimões Basin is a moot point.

The Juruá Formation shows a decreasing sand-grain-size upwards, suggesting a progressive increase in the water depth during deposition. Only in Peru, close to the Brazilian border, there is a deep water radioactive Devonian shale similar to the Barreirinha radioactive shale of the Amazonas Basin preserved (well Texaco 110). It seems that the Solimões Basin was a shallow water platform of the deeper water Subandean basins. In such conditions, the Devonian sea that flooded the Amazonas Basin was linked to the Phoibic Ocean (old ocean between Gondwana-Laurasia before the Hercynian orogeny (McKerrow and Ziegler, 1972), whereas the sea that invaded the Solimões Basin was linked to a western ocean. The Phoibic Ocean was connected to the Paleopacific Ocean through the northern margin of South America. This explains the differences between Parnaíba and Amazonas faunas and those of Bolivia and the Paraná basins which were linked to Paleopacific Ocean under colder climatic conditions during the Mid-Devonian.

JARAQUI FORMATION

I propose the Jaraqui Formation to designate a unit as much as 110 m thick composed of black diamictite, mudstone, siltstone and shale beds. Its name is derived from the Jaraqui stratigraphic well number 1 (1-JI-1-AM) drilled in the State of Amazonas. Its type-section is found in the 2,743-2,768 m interval. It was previously called Curuá Formation due to its similarities with the radioactive shales which occur in the Curuá Group (Barreirinha Formation) of the Amazonas Basin (Rodrigues and others, 1971). The figure 23 shows the isopach contour map of the Jaraqui Formation.

The Jaraqui Formation is a very important stratigraphic marker in the basin, and it has yielded much significant paleoclimatic and paleogeographic information. The unit´s lithology is strikingly uniform across the basin. The formation is mainly composed of diamictite. It is made up of a hard, dark-greyish, structureless massive clay matrix, supporting coarse-grained sand grains, granules and pebbles of quartz, quartzite, shale and crystalline rocks. These clasts vary considerably in size and roundness as observed in bore hole cores (Photo 1).

Dark-gray mudstone and laminated shale also occur in this formation along with minor amounts of siltstone. In the top of the section, hard, coarse-grained to conglomeratic, white, pale gray, cross-bedded sandstone beds with some dispersed quartzite clasts are intercalated in the section. The uppermost section is composed of black shale as thick as 10-20 m.

The Jaraqui Formation rests unconformably on the Juruá Formation, and is conformably overlain by the Sernambi Formation and unconformably by the Pucá Formation along the basin margins. If continental diamictites of the Jaraqui Formation were found resting directly over the Prosperança Formation in the eastern part of the basin, this relationship would definitively prove the absence of a marine connection between the Solimões and Amazonas basins during Devonian and Silurian times. The onlap of the continental Jaraqui Formation over the Prosperança Formation would indicate land towards the east where the Coari arch was located. However,

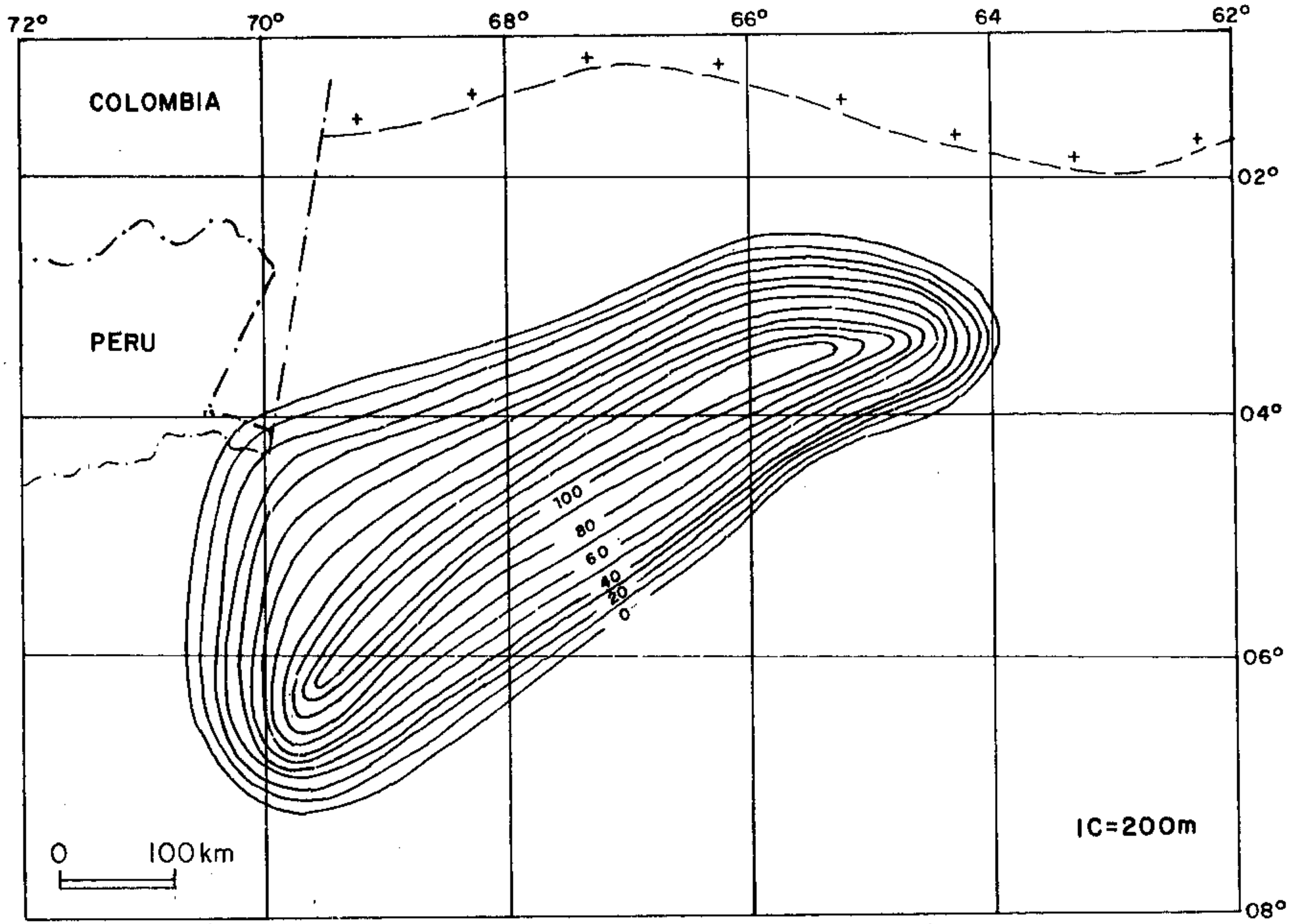


Figure 23 Isopach map of the Jaraqui Formation - Solimões Basin.

this stratigraphic relationship has not yet been found. The easternmost well (2-SR-1-AM) shows that the Juruá Formation pinches out between the Prosperança Formation, which is the basement for Paleozoic rocks and the Jaraqui diamictites, suggesting that the diamictites may overlie the basement towards the northeast.

The age of the unit was determined by Lima (1978) of the Petrobras paleolab, who ascribed a Late Devonian age (Famennian Stage) to this unit, based on palynological data. This corresponds to the local biostratigraphic zone VII established by Daemon and Contreiras (197la,b). Only unidentified brachiopods were found at the top of the unit.

The rock characteristics and intimate age correlation with glacigenic formations of the Amazonas and Parnaíba basins, support a glacial origin for most of the Jaraqui Formation.

It is interesting to note that Famennian marine deposits are not found in South America in the Subandean pericratonic basins. This gap in marine sedimentation is currently interpreted by many investigators as due to orogenic tectonism in Andean areas.

A break in sedimentation revealed by facies changes, or disconformities is seen everywhere in the world in Famennian time. Here this break in the sedimentation is regarded as being caused by a sea level fall due to the Late Devonian glaciation. Uplift on the periphery of the ice cap due to the growth of a forebulge may have contributed to the non-deposition in the Subandean foreland basins.

SERNAMBI FORMATION

I propose the Sernambi Formation to designate a sandstone unit occurring above the Jaraqui Formation. Its name comes from the Sernambi exploratory well number 1 in the State of Amazonas (1-SB-1-AM). Its type-section is found in the 2,761-2,800 m interval. Earlier the unit was named Oriximiná in bore hole final reports, because it occupies the same stratigraphic position of the Oriximiná Formation in the Amazonas Basin. However, it is very different from the Oriximiná section, which consists of an intercalation of shale, siltstone, sandstone and diamictites. The Sernambi Formation comprises sandstone with conglomerate beds and shale interbeds at the top of the section, and encompasses a wide age span. The upper part of the section was confused with the lower part of the overlying Puca Formation, (a Late Carboniferous unit) in some final well reports.

The unit consists of a succession of about 65 m in maximum thickness of white to light gray, poorly-sorted, subangular, silicified, parallel to cross-bedded fine to very coarse and conglomeratic sandstone beds with some dark-gray micaceous shale interbeds. The unit is petrographically immature with scattered quartz granules and feldspar grains. The silty shale and siltstones commonly have abundant organic material. The beds are lenticular and difficult to correlate The Sernambi Formation. apparently rests conformably on the Jaraqui Formation and it is unconformably overlain by the Pucá Formation.

The age of the unit was established by Daemon and Contreiras (1971a,b) and Lima (1982, written communication) from Petrobras paleolab as Latest Devonian (Late Famennian) to mid Early Carboniferous (Visean) on the basis of palynological data. No macro-fossils were found in the formation. Breaks in the sedimentation may occur in the unit, but the data are very poor to detect them. The Sernambi Formation was correlated in time with the Oriximiná and Faro formations of the Amazonas Basin.

After the melting of Late Devonian ice-sheets, the Solimões, Amazonas and Parnafba basins were encroached by the sea. This transgression was followed by a general regression when the Sernambi Formation was laid down. This unit represents the end of the sedimentary cycle that began in the early Middle Devonian (Eifelian) and terminated in the Early Carboniferous time (Visean Stage). The Sernambi Formation was deposited in a continental environment. Some upward fining sequences were identified in electrical logs at the top of the section, suggesting a fluvial environment of deposition.

Part of the unit may have been laid down in a glaciofluvial environment. This last inference was made considering paleogeographic and paleoclimatic conditions in the Andean area and Pimenta Bueno graben as well as in the Amazonas and Parnaíba basins where evidence of Mississipian glaciation also exists.

PUCÁ FORMATION

I propose the name Pucá Formation here to designate a formation composed of an intercalation of predominant sandstone and shale beds of Late Carboniferous age. Its name originated from the exploratory Igarapé (Creek) Pucá well number 1 (1-IP-1-MA) where its type-section is found in the 2,709-2,761 m interval. The formation displays variable thicknesses within the basin. It was previously named Monte Alegre Formation in final well reports, because its stratigraphic position was similar to that of the Monte Alegre Formation in the Amazonas Basin. However, the Monte Alegre Formation consists mainly of sandstone beds while the Pucá Formation consists of sandstone and silty shale interbeds. The figure 24 shows the isopach contour map of the Permo-Carboniferous section (Pucá, Itaituba, Nova Olinda and Fonte Boa formations).

The unit is composed of white, friable, massive or cross-

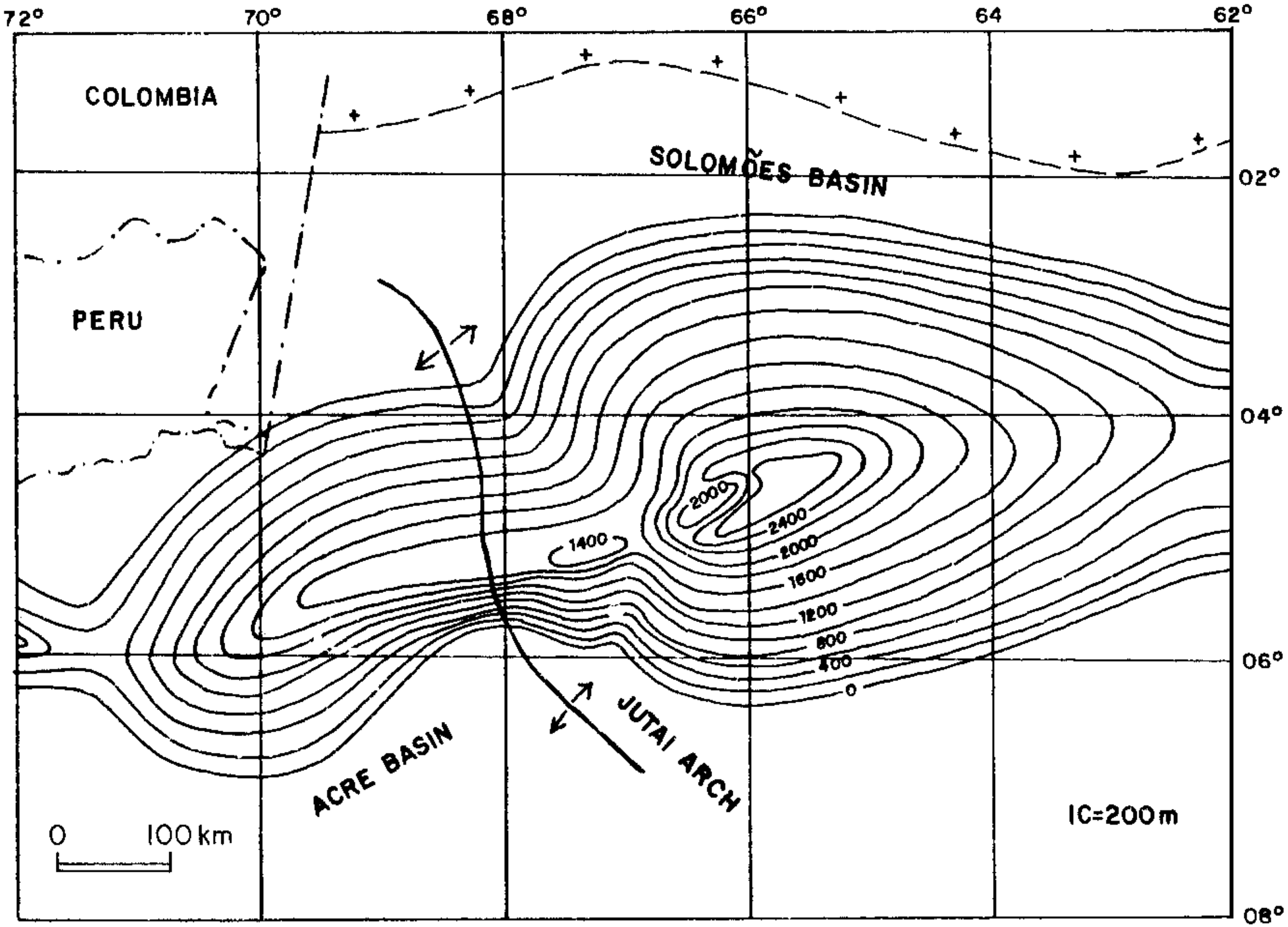


Figure 2.4 Isopach map of Pucá, Itaituba, Nova Olinda and Fonte Boa Formations - Solimões Basin.

bedded, fine- to coarse-grained sandstone beds with basal conglomerates in almost every sandstone body. Finer upward sequences are commonly developed in the sandy section together with an increase in clay content. There are some fine- to medium-grained, large-scale cross-stratified bodies in which an alternation of fine- and medium-grained strata are present. Fine- to very fine-grained argillaceous sandstone beds are also common throughout the section and red-brown to dark-gray shale beds occur intercalated in the sequence with some scattered plant remains.

The lower contact of this unit is disconformable above the Sernambi Formation in the central parts of the basin and above older units at the basin edges. This contact is difficult to locate because the top of the Sernambi Formation and the base of the Pucá Formation have similar electric log characteristics.

The contact is best determined by palynology. Outside of the area of occurrence of Early Carboniferous beds, the Pucá Formation is very thin and discontinuous, disappearing in many places. In this case, the overlying Itaituba Formation rests directly on the Uatumã basement or Prosperança Formation. The uppet contact is conformable and it is placed where clastic sediments are overlain by anhydrite or carbonate beds of the overlying Itaituba Formation. No macrofossils were found in the unit. Its age was determined by Daemon and Contreiras (197la,b) as Mid-Pennsylvanian

(Westphalian) on the basis of palynological studies. According to correlation with the Amazonas basin the deposition of the Pucá Formation began earlier than the Monte Alegre Formation. A Westphalian "C" age Is inferred.

According to Terra and others (1980) the unit here designated Pucá Formation represents continental deposition under desert conditions where fluvial, aeolian and lacustrine, both marginal and central, deposits were cyclicly laid down. The basal sediments of this unit, in the concept of Terra and others (1980), also includes the upper part of the Sernambi Formation of mid Early Carboniferous age (Visean Stage). Therefore, Terra's section presents a gap in the sedimentation corresponding to the Namurian (late Early Carboniferous) and the Early Westphalian (earliest Late Carboniferous) stages. In the Solimões Basin it is noted that from Ordovician to Early Carboniferous time no carbonate or other warm climate sediments were deposited. The high clastic supply associated with a weak tectonic background may suggest that severe climatic conditions controlled the nature of sedimentation during Ordovician to Early Carboniferous time.

ITAITUBA FORMATION

The name Itaituba Series was introduced by Hartt (1874) to designate Carboniferous limestone, dolomite, shale, siltstone and sandstone interbeds. This same rock assemblage occurs continually from the Amazonas to the Solimões Basin. The Itaituba Formation began its deposition earlier in the Solimões Basin than in the Amazonas Basin as observed in bore hole correlation. There is a basal section in the Solimões Basin not found in the Amazonas Basin. The Monte Alegre Formation of the Amazonas Basin and the Pucá Formation are in the same stratigraphic position underlying evaporitic rocks but differ a little in age and may not be continuous. Along the Purus arch, the Monte Alegre Formation is missing in some wells; the Itaituba Formation overlies Devonian to Precambrian formations.

The Itaituba Formation in wells is mainly composed of limestone, dolomite, anhydrite, shale, sandstone and siltstone deposited in numerous well defined sedimentary cycles which are mostly separated by shale beds. Each cycle started with marine carbonates, first bioclastic then stromatolitic, overlain and partially replaced by anhydrite, that in some rare cases is overlain by low-bromine coarse halite, such as that precipitated frequently in wide lagoons today (Szatmari and others, 1975). It is important to remember that any sedimentary rock may constitute the substrate for crystallization of anhydrite crystals below the ground surface in Sabkha deposits. Thus many carbonates are almost entirely replaced and displaced by nodular anhydrite crystals.

Detrital land-derived material contributed to siltstone, shale and fine-grained sandstones in the central regions of the basin and shale and medium-grained sandstones at the basin edges. Salt beds are also developed in the section at the end of some cycles. Brachiopods, pelecypods, crinoids, corals, foraminifera, bryozoans, trilobites and other fossils comprise the rich fauna of the formation. A Mid-Pennsylvanian age was established by Petri (1952), on the basis of fusulinids of the Amazonas Basin. Daemon and Contreiras (1971) pointed out that the unit started its deposition in Westphalian O Stage and terminated in the Stephanian Stage (local biostratigraphic intervals XIII and XIV) in the Amazonas Basin. In the Solimões Basin, Itaituba basal deposits may have been deposited earlier in the Westphalian C Stage.

The deposition of the Itaituba Formation first took place in the western side of the Solimões Basin and the transgression progressed slowly eastward (Caputo and Vasconcelos, 1971), reaching the Coari arch when about 300 m of sediments had been deposited in the central parts of the Solimões Basin (Szatmari and others, 1975). The Coari arch was a barrier to the eastward marine advance to the Amazonas Basin, but a general downwarping in the area, associated with subsidence caused by the increasing sedimentary loading in the Solimões Basin, pushed the divide (Coari arch) between both basins eastward to the position where the Purus arch now exists.

At first no marine water crossed the Coari high. However, after deposition of a considerable thickness (about 300 m) in the Solimões Basin, some marine water crossed the divide, thereby invading the Amazonas Basin flats, forming some minor and local fossiliferous carbonate interbeds in the Monte Alegre sandstone. Thus, that part of the Itaituba Formation in the Solimões Basin is synchronous to the upper Monte Alegre Formation in the Amazonas Basin and to the Middle Piauí Formation in the Parnaíba Basin, as shown by the common fossil content in the three basins. It is also correlated with the Tarma Formation of Peru (Mendes, 1961).

The Permo-Carboniferous Solimões Basin was larger than the Siluro-Devonian Solimões Basin. In the north, east and south edges of the Solimões Basin, the Itaituba Formation covered unconformably the Prosperança Formation and the Uatumã Group.

The Itaituba Formation is a very important paleoclimatic indicator, as it records a shift from a circumpolar belt in Early Carboniferous to a subtropical belt of high evaporation in Late Carboniferous time. It should be noted that during Stephanian (Late Carboniferous) time the Paraná Basin and some portions of the Andean basins were covered by ice. In Late Carboniferous time, the Solimões Basin area was not under the influence of glaciation. The cyclic evaporitic sedimentation may indicate that the sediments were controlled by sea-level changes and climatic conditions.

NOVA OLINDA FORMATION

The Nova Olinda Group was named by Kistler (1954) to designate a sequence composed of Permo-Carboniferous evaporites in the Amazonas Basin. Later, the unit was considered as a formation and was also identified in the Solimões Basin, but with minor development of halite.

The unit is characterized by a wide lithologic heterogeneity and a cyclic pattern in the sedimentation. It consist of anhydrite, limestone, dolomite, shale, siltstone, sandstone, and halite with much bed variation in thickness. In the Solimões Basin, the Nova Olinda Formation has much less halite and sandstone than the Itaituba Formation.

Its lower contact is conformable with the Itaituba Formation and is placed at the base of radioactive shale beds. It is also conformable with the overlying clastic Fonte Boa Formation which is present only in the central parts of the basin. The Nova Olinda Formation is unconformably overlain by the Cretaceous part of the Alter do Chão unit.

The unit is less rich in macrofossils than the Itaituba Formation, which have not been studied in the Solimões Basin. The same fossil groups as those of the Itaituba Formation are found. Daemon and Contreiras (1971a,b) placed the unit in part of the local biostratigraphic interval XIV and in the biostratigraphic intervals XV and XVI (Stephanian to Sakmarian stages) ranging from Late Pennsylvanian to Mid-Permian age. Dwarf forms and *Lioestheridae* specimens as well as salt deposits are indicative of large changes in the water salinity during deposition of the Nova Olinda Formation. At the time of its deposition, the Solimões Basin behaved as a platform with more communication with the sea than that of the Amazonas Basin which was environmentally more restricted.

Morales (1959) stressed that the Iquitos arch had its origin in Devonian time and culminated its development in the Permian. Since Devonian, it was supposed that oscillations of the Jutaí arch strongly influenced the cyclic evaporitic sedimentation of the Amazonas and Solimões basins and at the end of the Permian Period its final uplift was responsible for the isolation and closure of the evaporitic basin. However, according to De Boer (1964), the upper Carboniferous evaporites (Itaituba Formation) thicken over the Jutaí arch, indicating a subsidence rather than arching, and isopach maps from each evaporite cycle do not show any differential tectonism in the Jutaí area (Szatmari and others, 1975) during evaporite sedimentation. In light of such evidence, evaporite deposition in the Solimões Basin was controlled by worldwide sea-level changes, climate, and paleogeography rather than by tectonic oscillations. Evaporitic as well as clastic cyclothems are widespread from Devonian (House, 1975) to Permian times. Many of these cycles may be a result of glacial and interglacial stages in the Gondwana Continent.

FONTE BOA FORMATION

The term Fonte Boa Formation is proposed here to designate a Late Permian sequence mainly composed of variegated siltstone beds with minor shale and sandstone interbeds. The name is derived from the Fonte Boa stratigraphic well number ·1 (2-FB-1-AM) and located in the 538 and 663 m interval.

This section was called Andirá Formation by Szatmari and others (1975) on the basis of its similar stratigraphic position to that of the Andirá Formation in the Amazonas Basin, but these sections. are separated by the Purus arch.

The siltstone beds are greenish-gray, red, brown, and cream-colored, calcareous, and are commonly massive, hard, and at places silicified. Some red-brown and brown-gray, calcareous, laminated shale beds occur grading to siltstone. Fine to very fine, light gray to light cream-colored hard sandstone beds rarely occur in the section.

The Fonte Boa Formation overlies conformably the Nova Olinda Formation. Its lower contact is placed at the top of the limestone or anhydrite beds of the Nova Olinda Formation. It is unconformably overlain by the Cretaceous part of the Alter do Chão Formation. In most places, diabase sills often occur in its upper contact. Its age is inferred as Late Permian on the basis of its stratigraphic position and because the strata are similar to those of the Andirá Formation of the Amazonas Basin whose age was determined by Daemon and Contreiras (1971a,b) on the basis of palynological data.

The Fonte Boa Formation represents the end of the sedimentary cycle that lasted from Mid-Pennsylvanian to Late Permian (or Earliest Triassic) in which warm climate prevailed. The composition of the Fonte Boa Formation is unique in the basin, because it consists mainly of siltstone beds which are generally subsidiary lithology rather than the main sediments of any formation in the geological record. Massive calcareous siltstone beds with laminated shale interbeds may indicate aeolian and lacustrine deposition. Their source may have been loess deposits formed under arid conditions.

The Fonte Boa Formation was previously interpreted as deposited in an isolated basin due to the uplift of the Jutaí arch to the west; however, in Peru the Upper Permian Mitu Formation also consists of regressive continental red beds, so that both formations are interpreted as a result of a major worldwide regression without important tectonic activity in the Jutaí arch.

The total marine withdrawal from the Solimões Basin under arid climatic conditions, may have transformed the flat area into widespread deserts. It seems likely that during Paleozoic times, fast transgressions and regressions observed in the Solimões and Amazonas basins are related to global tectonism and sea-level changes and glacial activity rather than to local tectonism.

BASIC MAGMATISM

From Carboniferous to Jurassic times, the Paleozoic sequence was intruded by diabase which formed dikes and sills. Only in the western part of the basin, effusion of basic magma took place in Carboniferous and Triassic times. Outside the basin, in the northern Guyana shield, Amaral (1974) observed the Taiano diabase dike whose age is 360 m.y. (early Late Devonian-Frasnian Stage).

In the foreland basins, basic igneous activity is not manifested. In these areas, acid magmatism and widespread tectonic activity took place. No Carboniferous to Jurassic diabase intrusions are known in the area west of the Jutaí arch at present. It seems that a cratonic tensional regime east of the arch has changed to a compressional one to the west.

In the Solimões Basin, the load caused by 6,604 km^3 of heavy diabase may have induced an outflow of asthenospheric material from beneath the central parts of the basin toward its margins, including the Carauari area which was uplifted after the magmatic activity. The stacking of the first Andean folded chains and thrust fault plates may have contributed to subsidence of the foreland basins and as a consequence to the uplift of marginal regions, including the Jutaí arch, also due to the outflow of asthe-nospheric material from beneath the Andean chain towards the Jutaí arch in Early Cretaceous time.

ALTER DO CHÃO FORMATION

The name Alter do Chão Formation was first used by Kistler (1954) to describe red beds that overlie the Paleozoic section in the Amazonas Basin. This Cretaceous formation is continuously developed from the Amazonas to the Solimões Basin. The name originated from the Alter do Chão stratigraphic well number 1 and the Alter do Chão hill where the type-section is located.

The figure 25 shows the isopach contour map of the Cretaceous and Tertiary section (Alter do Chão and Solimões formations).

The unit consists of poorly consolidated fine- to coarse-grained and conglomeratic sandstone with minor claystone interbeds and local siltstone and shale intercalations.

Cross-stratified red to variegated, clayey, soft, poorly sorted very coarse- to medium-grained sandstone beds with minor brick red massive or laminated claystone beds comprise the prin-

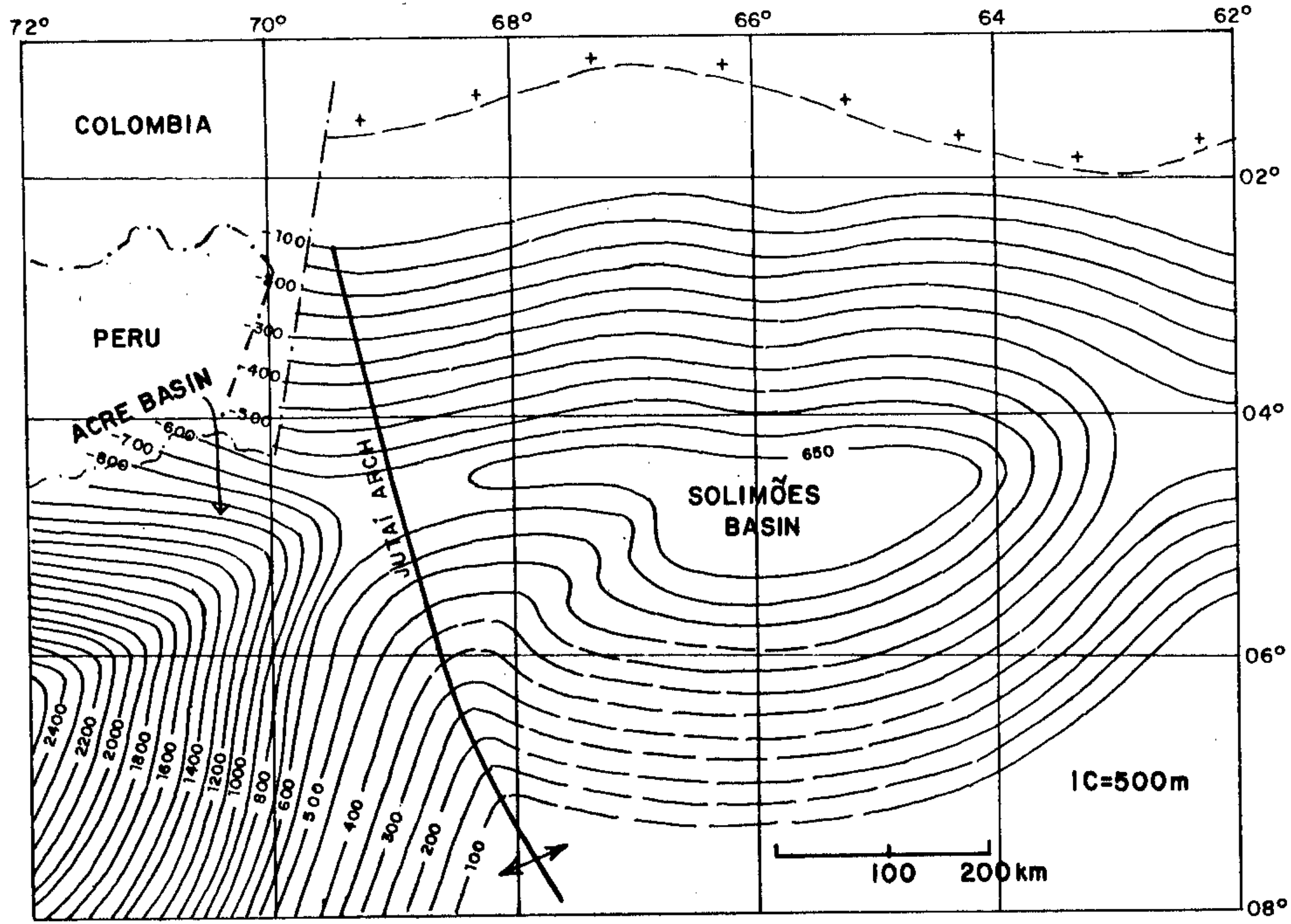


Figure 25 Isopach map of the Alter do Chão and Solimoēs Formations – Solimões Basin

cipal sedimentary bodies. Conglomerate beds have quartz, quartzite, sandstone cobbles and pebbles occurring in many levels in the section. This unit overlies unconformably all older formations of the basin and is unconformably overlain by the Tertiary Solimões Formation.

The Alter do Chão Formation correlates with the Jaquirana Group comprising the Moa, Rio Azul and Divisor formations of the Acre Foreland Basin (Caputo and others, 1979). In the Amazonas Basin, Price (1960) dated this unit as Late Cretaceous on the basis of a Theropoda (dinosaur) teeth. The same sequence was dated by Daemon (1975) as Middle Albian, Cenomanian and Turonian based on palynological studies. In the Acre Basin, the unit increases in thicknes and probably in age range.

Close to the Brazilian-Peruvian border a thin marine glauconitic sandstone intercalation occurs in the Jaquirana Group (Caputo, 1974) which is predominantly continental in origin.

Fluvial and lacustrine related environments of deposition under humid tropical conditions predominated throughout the accumulation of these sediments. The presence of terrestrial vertebrate bones and teeth remains and plant fossils, both wood and leaves, confirm the continental milieu of sedimentation.

The Cretaceous drainage system of the Solimões Basin was directed westward where its overall coarseness gradually decreases and its thickness increases.

It is interesting to notice that the Alter do Chão Formation was deposited and preserved during the time of the major transgression in the foreland Pastaza Basin, Peru.

The transgression might have induced the Solimões Basin rivers to change their regime from erosional to depositional. At the end of Cretaceous and Tertiary times the Andean belt may have continued to rise and its load may have depressed the area between the Andean belt and the Amazonian shield, making room for thick Tertiary sedimentation in the foreland area. This concept of loading of the lithosphere was invoked by Price (1971) in order to explain the subsidence of the Canadian Rocky Mountains foreland. While the Andean belt was being uplifted, the peripheral depression was being deepened and amplified by more sediment loading so that the overlying Solimões Formation may have over-stepped the Jutaí arch, covering an area of about 1,000,000 km^2 of Brazilian territory.

SOLIMÕES FORMATION

The name Solimões Series was first used by Rego (1930) to designate a Cenozoic argillaceous section along the Solimões River. In the Acre and Solimões basins, this unit had many local names, so Caputo and others (1971, 1972) correlated the various sections and gave them the name Solimões Formation.

In the Solimões Basin, the unit is up to 450 m thick and is mostly composed of soft, light gray, gray and greenish gray, massive to laminated clay with lignite interbeds from 2 to 10 mm up to 90 cm in thickness. Subangular to subrounded, fine- to coarse-grained, soft, white sandstone interbeds are present in the upper part of the section.

The Solimões Formation overlies unconformably the Alter do Chão Formation. It seems that the unconformity time span increases from the Acre Basin to the Soltmões Basin and to the Purus arch. In the foreland Andean basins the Cretaceous-Tertiary section is probably continuous. The formation contains abundant terrestrial fauna and flora which consist of nonmarine ostracodes, fish scales, vertebrate teeth and bones, gastropods, bivalves, charophita algae, plant remains, both leaves and stems, and sporomorphs. Detailed fossil collection references are provided by De Oliveira and Leonardos (1943}, and Francisco and Loewenstein (1968).

In the central part of the Solimões Basin, the lower section is probably Eocene or Oligocene in age on the basis of poor paly- nological data (Lima, 1982, oral communication). The age of the lignite beds, which are located about 3,000 km from the source of moisture (Atlantic Ocean), ranges from Miocene to upper Pliocene

(Kokis, 1970) based on ostracodes. The upper part of the section may be Pleistocene or Holocene.

Some geologists who have mapped the Solimões Formation considered it as having been deposited in Pliocene and Pleistocene times. This age assignment corresponds only to its upper part, without considering the age of lower strata drilled for coal and oil.

In Peru, in the Andean foothills, many conglomerate (alluvial fan deposits), sandstone and tuff beds were laid down during Tertiary times, their facies change eastward to sandstone and clay beds. In the Acre Basin, in Brazil, some tuff beds are also present. The Tertiary section is a typical molasse deposit generated by the Andean Orogeny. Paleocurrent directions deduced from cross-bedding indicate that the source area of the Solimões Formation was located in the Andean belt (Mason and Caputo, 1964), but Cretaceous source areas were located in the Amazonian shield.

The Solimões Formation represents a distal molasse sedimentation characterized by alluvial and paludal fine-grained clastics which overstepped the Jutaí arch and Purus high, covering the entire Solimões Basin and the western edge of the Amazonas Basin where the upper part of the unit was probably laid down, also during Holocene time (Santos, 1974).

The depositional environment of the upper part of the unit is also tnterpreted on the basis of radar imageries. Despite the vegetation cover, the surface of the Solimões Formation is crowded with a multitude of meandering paleochannel scars, indicating that the unit was a result of a fluvio-lacustrine aggradation.

In Oligocene time, a marine incursion (Pozo Formation) took place in Peru, in the foreland Pastaza Basin (Kummel, 1948). This may indicate that up to Oligocene time the drainage system was directed towards the Pacific Ocean, but in Miocene time a more intense tectonism closed the western portal, changing the fore-land basins drainage to the Atlantic Ocean.

During the Pleistocene, the glacio-eustatic sea-level falls may have interrupted its sedimentation many times. At present, sedimentation is taking place again in lakes, and along river banks and broad flood plains, indicating that subsidence is still effective. Agassiz, who in 1840 had developed the hypothesis of Alpine and continental glaciation in Europe and in the United States, also visited the Amazonas valley in 1866. In his trip along the Amazonas River and some tributaries, he interpreted the surficial sediments (Solimões Formation) as drift resulted from glacial activity. Agassiz envisaged the ice-sheets encompassing a vast area from the Andes to the northeastern coastal area of Brazil.

Glacial activity in equatorial flat areas caused wide discussion .in scientific centers all over the world, but the presence of warm water mollusk fossils, similar to recent ones in the supposed drift, allowed Orton (1870), Hartt (1870) and Branner (1919) to refute Agassiz' interpretation. The existence of alligator fossils of warm water habitat also militates against glacial activity in the area during Pleistocene times.

CHAPTER 9. AMAZONAS BASIN

Pioneer geological surveys carried out in the Amazonas Basin date from 1862, when Chandless traversed the Tapajós River, and 1863 when Coutinho (in Agassiz, 1866) found Carboniferous fossils in the Cupari River (Tapajós River tributary). Important geological contributions were next provided by Agassiz (1866), Hartt (1870, 1874), Derby (1878), Katzer (1903), Albuquerque (1922), De Carvalho (1926), De Paiva (1929), Rego (1930), De Moura (1932, 1938), De Oliveira and Leonardos (1943), Petri (1952), Oddone (1953) and De Oliveira (1956) who built the basis of the Amazonas Basin stratigraphy.

From 1954 on, many research projects involving the geology of the area were carried out by the geologists of Petrobras. De Oliveira and Leonardos (1943), De Oliveira (1956), Bouman and others (1960), De Boer (1964, 1965, 1966), Lange (1967) and Bigarella (1973) made significant regional studies and provided important reference lists. Ludwig (1964), Rodrigues and others (1971) and Carozzi and others (1972, 1973) studied the sedimentology of the basin.

The historical development of the stratigraphic column was summarized by Caputo and others (1971, 1972) who made the last overall revision of the basin stratigraphy. Some modifications made since then are incorporated in this study. Figure 18 pre-

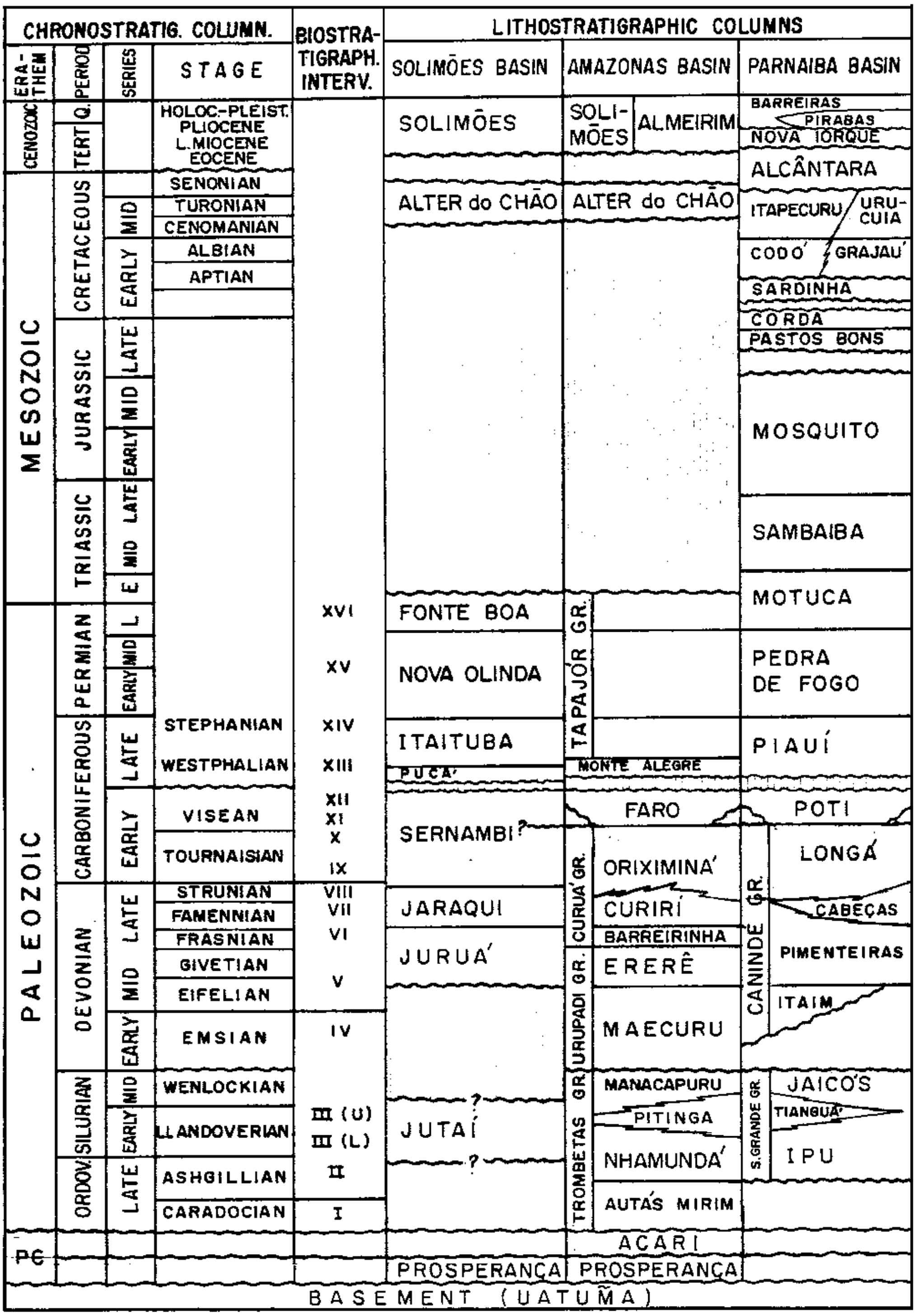


Figure 26 Stratigraphic correlation chart among Solimões, Amazonas and Parnaíba basins.

sents a cross section of the basin.

The stratigraphy of the basin is described by me in the following sections based on surface and subsurface data. The main purpose of this study is to examine the climatic effect on sedimentary rocks, and tectonic and paleogeographic development of the area. Figure 26 shows the stratigraphic chart of northern basins of Brazil; figure 27 shows a cross-section across the Amazonas Basin and figure 28 shows a cross-section along the Amazonas Basin.

PROSPERANÇA FORMATION

The Prosperança Formation, of probable Precambrian age, is common to both Solimões and Amazonas basins and it was described under the Solimões Basin section. It is present in outcrop and subsurface in the western side of the Amazonas Basin and in the subsurface in the eastern and central parts of the Solimões Basin. In the deepest and central parts of the Amazonas Basin, eastward from the boundary between the states of Amazonas and Para there is no bore hole control, and for this reason the eastern extent of the Prosperança Formation is not yet known. Its maximum thickness is estimated to be over 1,000 m in the Purus arch.

The Prosperança Formation separates the pre-Late Carboniferous rocks of the Solimões Basin from those of the Amazonas Basin along the Purus arch, making up a significant geologic divide.

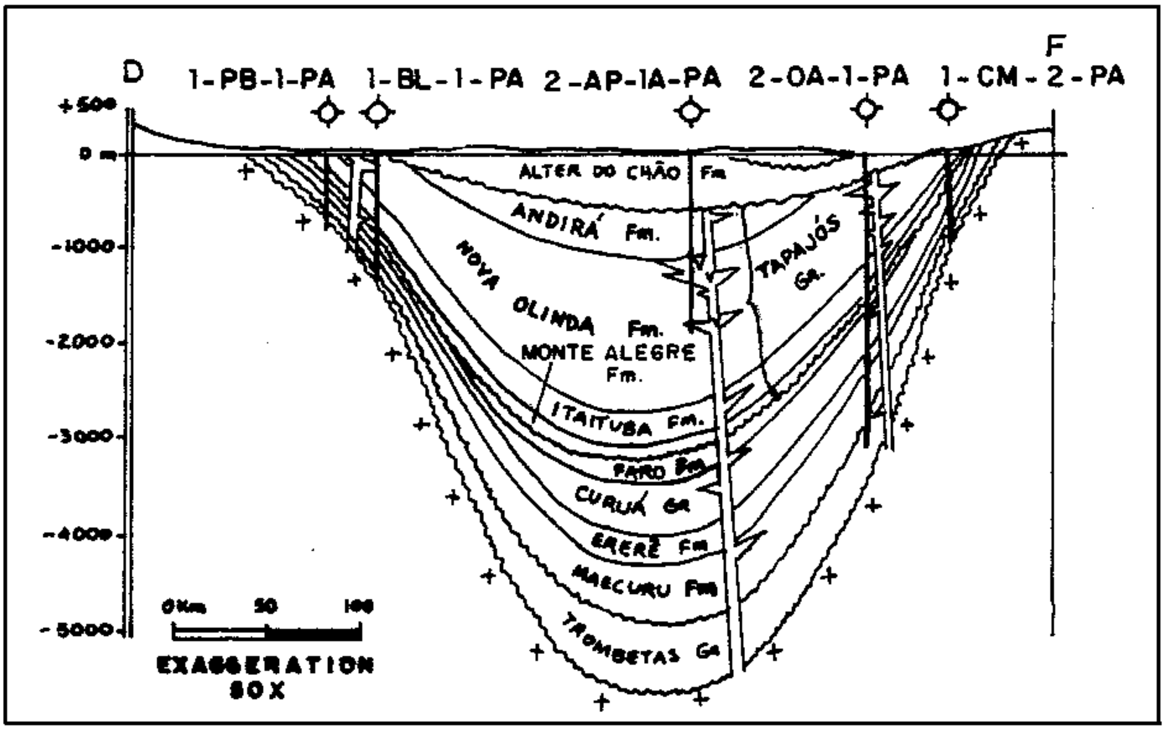


FIGURE 27- Amazonas Basin cross-section

The age of the formation is not known, but it is intruded by 1400 and 1100 m.y. old diabase (De Montalvão and Bezerra, 1980) and overlies the Uatumã Group (basement). This stratigraphic relationship suggests that the unit was deposited around 1500 m.y ago.

The Formation could be correlated on the basis of rock lithology with the Sete Quedas Formation (Geomineraçâo, 1972) and with many other formations overlying the Uatumã Group along the Aripuanã, Sucunduri and Tapajós rivers (Caputo and others, 1971).

ACARI FORMATION

Hoyling (1957) used the designation Acari Formation for a carbonatic sequence occurring in the southwest flank of the basin in the stratigraphic bore hole Acari number 1 (2-CA-1-AM), at the 195-595 m interval. The unit consists of limestone and dolomite beds with mudstone and siltstone intercalations. It partly overlies the Prosperança Formation. The limestone and dolomite beds make up bodies tens of meters thick. These are highly silicified, variegated, hard, stylolitic and as compact as marble. The mudstone and shale beds are normally brick-red and dark brown, hard, compact, and calcareous. Some subordinate cream-colored, pink, and red calcareous, fine- and medium-grained sandstone beds are present in the section.

The Acari Formation overlies unconformably the Prosperança Formation, but the red-color of the Acari Formation could indicate that it is, in part, a continuation of the red beds of the Prosperança Formation because fine red clastics of both units are very similar. The upper contact with the Autas-Mirim, Monte Alegre, Itaituba and Alter do Chão formations is definitely unconformable with a very low angle.

The Acari Formation has not yielded fossils. However, geochronologic dating based on the K/Ar method assigns a Late Pre-Cambrian age (1,360 m. y.) (Thomaz Filho, 1983, oral communication).

The red beds and limestones of the Acari Formation indicate warm depositional conditions. The presence of scattered sand-grains in the mudstone and shale beds suggests strong wind activity. In the Late Precambrian time, the absence of vegetation probably allowed intense aeolian transport in dry as well as in humid climates. Only six bore holes reach the formation, so its complete geographic distribution is not known. These wells are found in the Purus arch and vicinities. In the Solimões Basin the unit has not been well characterized due to a poor sampling.

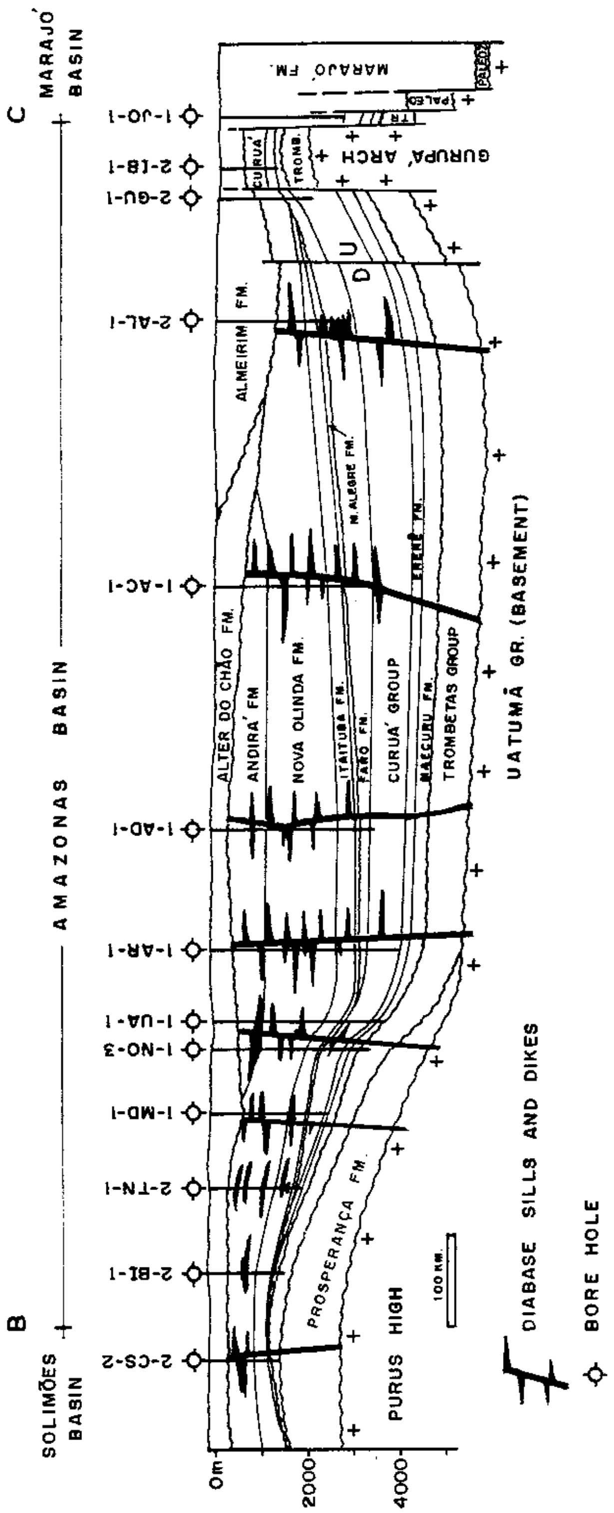


Figure- 2.8 Amazonas Basin longitudinal cross-section

TROMBETAS GROUP

Here it is proposed to raise the Trombetas Series, of Ordovician-Silurian age introduced by Derby (1878), to the rank of group and its subdivisions to the rank of formations. Its maximum thickness is estimated to be over 800 m in the central parts of the basin. The Trombetas Group then consists of the Autás-Mirim, Nhamundá, Pitinga and Manacapuru formations. The Trombetas Group was deposited after the rifting of the Amazonas Basin. This section is considered here as the first sedimentary deposit of the Amazonas Basin. Figure 29 shows the isopach contour map of the Trombetas Group.

AUTÁS-MIRIM FORMATION

The Autás-Mirim Member was defined by Caputo and others (1971) as a section of presumed Ordovician age composed of sandstone beds with subordinate siltstone and shale interbeds which occur at the base of the Trombetas Formation. Here the Autás-Mirim section is raised to the category of a formation and the Trombetas Formation is raised to the rank of group. Figure 30 shows the isopach contour map of the Autás-Mirim Formation. The unit consists of micaceous, feldspathic sandstone beds which are mainly fine-grained, commonly well rounded, red, pink, light green and brown at the base of the section, and predominantly white in the

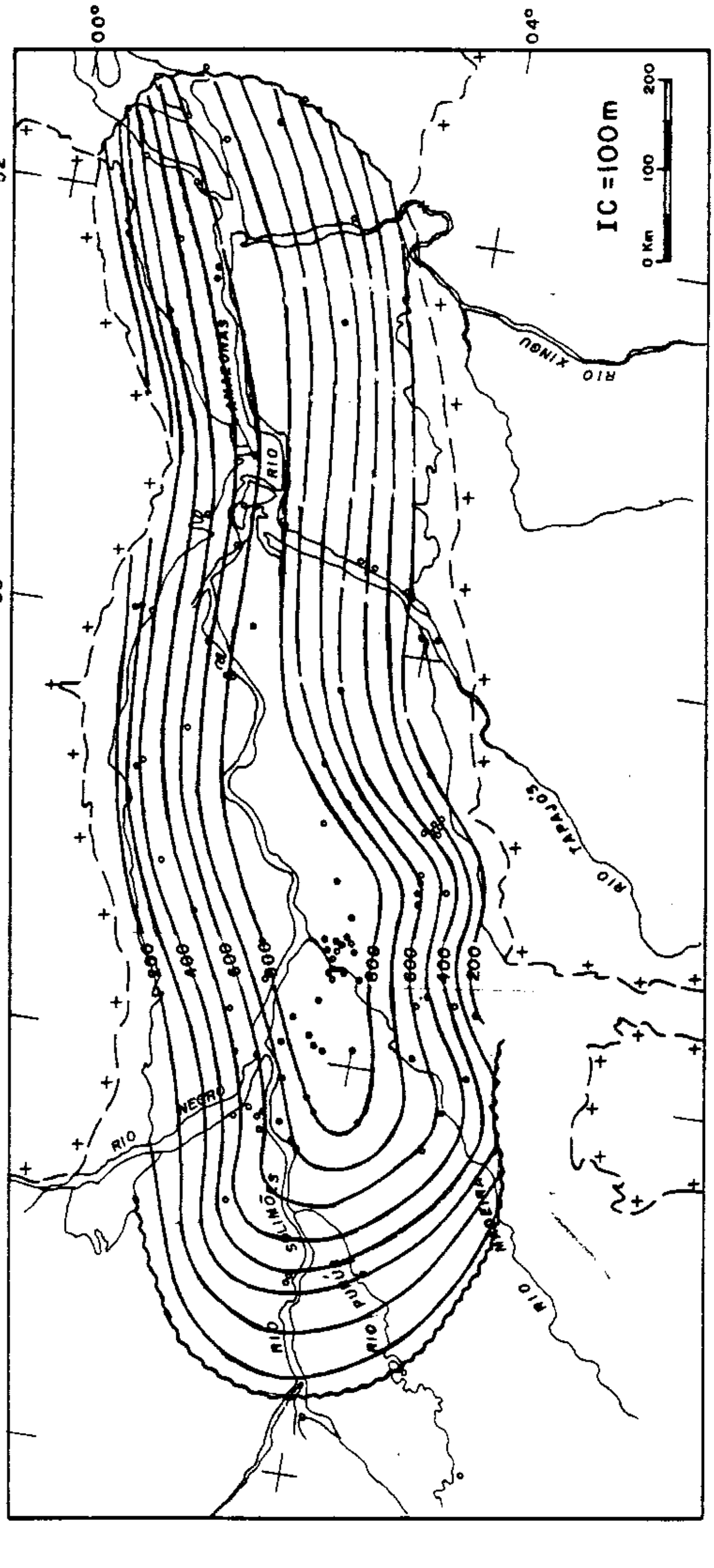


Figure 29 Isopach map of the Trombetas Group - Amazonas Basin

mid, and upper parts. The beds are mainly cross-bedded at the base and top of the section and chiefly parallel laminated at the middle of the section. A kaolinitic matrix predominates and the beds are well indurated. The siltstone beds are light green, greenish gray, highly micaceous and feldspathic, well indurated, and laminated. Minor bioturbations are common. The shale beds are micaceous dark gray, with few biogenic structures, and are well indurated.

The maximum thickness known of the formation is about 350 m. It overlies unconformably the Uatumã Group and the Prosperença and Acari formations at different places. Eastward of the Santarém City, the subsurface distribution of the Autás-Mirim Formation is unknown, but its presence is extrapolated. The unit is present only in the subsurface; it is probably overlain conformably by the Nhamundá Formation.

In some sections, the unit contains ill-preserved quitinozoans and skolithos, trace-fossils, indicating a very shallow marine environment during part of its sedimentation. The Autás-Mirim Formation represents a continental-marine-continental depositional cycle, which slowly onlapped the basin flanks. It was deposited under fluvial-estuarine conditions on the margins of the basin as indicated by cross-bedding, pink sediments, absence of trace fossils, kaolinitic matrix, and fine and medium grain size alternations. The beach deposits are characterized by fine-

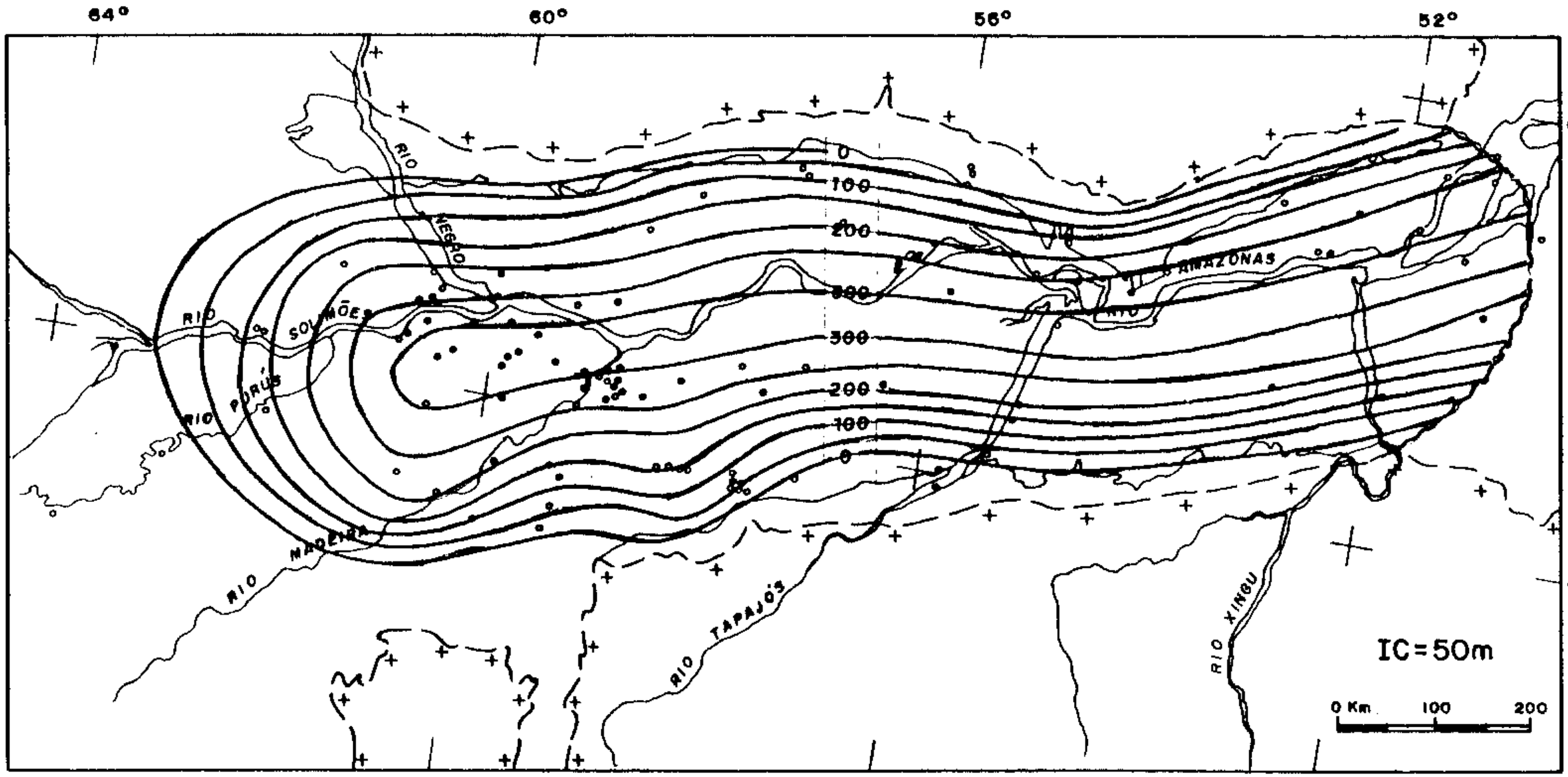


Figure 30 Isopach map of the Autás - Mirim Formation - Amazonas Basin

grained laminated sandstone beds, skolithos, and well rounded fine-sized sand grains. Its grain modal class between O.125 and O.177 mm (Rodrigues and others, 1971) measured in thin-sections and well rounded grains may indicate a strong aeolian contribution. Sea level fluctuations may have allowed intense aeolian reworking of sediments previously laid down, mainly because the loose material was not protected by vegetation in pre-Silurian times.

Along the basin flanks, laminated and very fine-grained sandstones, as well as silty sandstones, are interpreted as shoreface deposits (up to the fair-weather base wave). The horizontal lamination suggests deposition of sand from air suspension during storms (Reineck and Singh, 1980). Offshore deposits are inferred to be present in the central parts of the basin.

The age of the formation is unknown, but as it underlies an Late Ordovician formation and was not probably affected by the Brazilian tectonic event in the Marajó Basin, a Late Ordovician age is inferred (Caradocian Stage?).

The climate may have been cold at the time of the deposition of the Autas-Mirim Formation, as evidenced by the lack of warm climate indicators and the abundance of feldspars and micas. The very fine sandstone and sandy siltstone are highly radioactive as seen in gamma-ray logs, probably due to the high feldspar content.

Previously, Rodrigues and others (1971) interpreted the sediments in the western part of the basin as tidal-originated, but the distance from the Theic ocean deep water (ocean between Gondwana and ancestral North America before the Acadian orogeny-McKerrow and Ziegler, 1972), the probable small oceanic connection, and the shallow nature of the sea, point out for a tideless, or only weakly tidal sea. The environment energy may have come from wind-induced processes, with a negligible tidal influence (Johnson, 1978).

The sea encroached upon the Amazonas Basin from the east, and sediment deposition occurred all the way up to the eastern flanks of the Coari arch.

NHAMUNDÁ FORMATION

The Nhamundá Member was proposed formally by Lange (1967) to designate a sandstone unit which occurs at the base of the Trombetas Formation. Here the Nhamundá Member is raised to the category of formation because it can be mapped throughout the basin. Figure 31 shows the isopach contour map of the Nhamundá Formation.

The Nhamundá Formation, up to 450 m thick, crops out only in the northwestern outcrop belt from the Negro to the Pitinga rivers and the best exposures are found along the Urubu River.

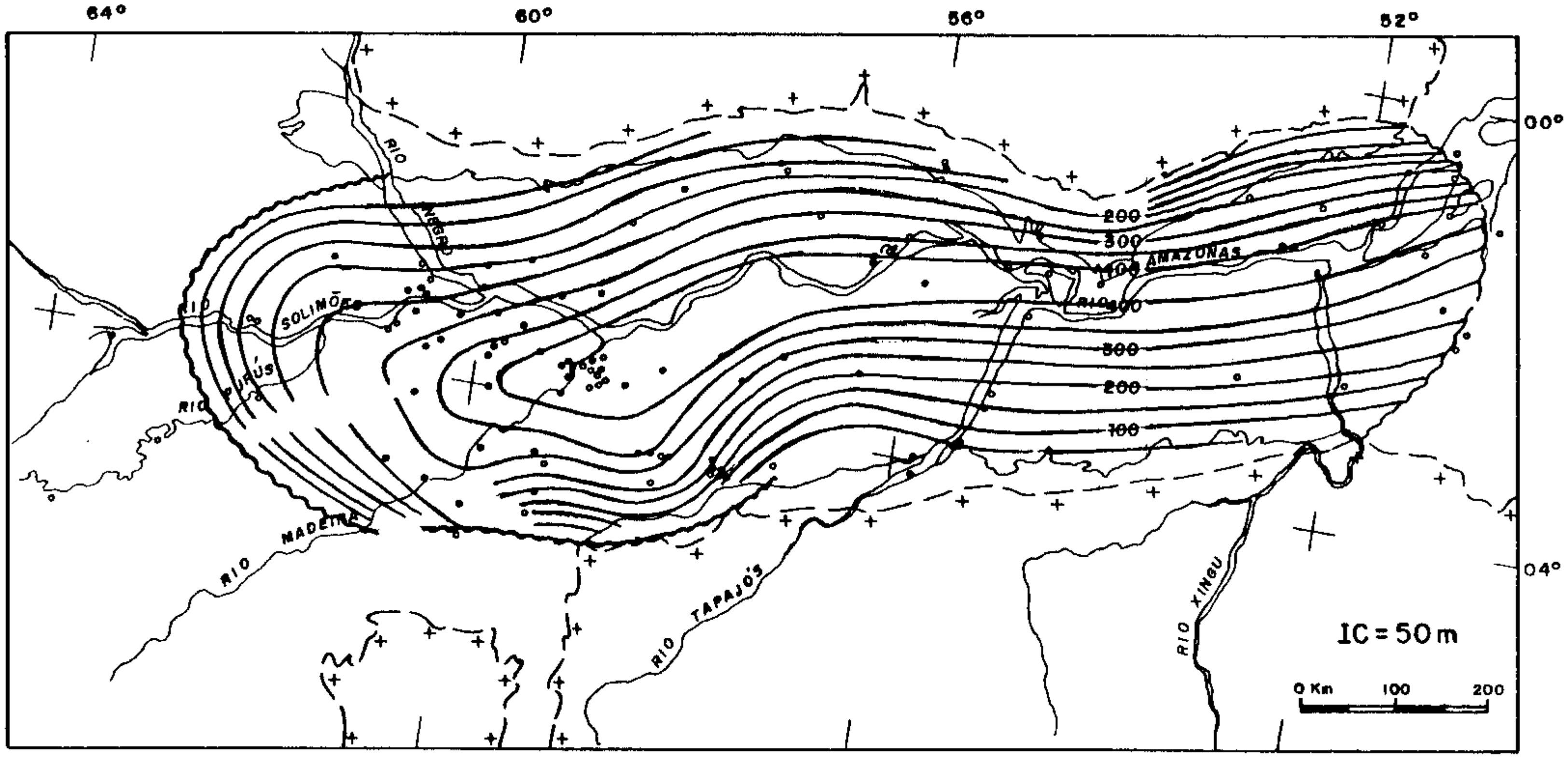


Figure 3 1 Isopach map of the Nhamundá Formation - Amazonas Basin

The lowermost part of the section is not exposed in the outcrop area but it is present in the subsurface throughout the basin.

The outcrops of the Urubu River were subdivided here into three units:

(1) In the lower part of the section a micaceous cream-colored silty sandstone 2 m thick is overlain by sediments, 25 m in thickness, that consist of pink, fine-, and medium-grained, angular, kaolinitic and highly cross-bedded sandstone beds.

(2) The mid-part consists of light gray, hummocky, cross-stratified and ripple cross-laminated sandstone with skolithos and scattered medium and coarse rounded quartz grains and minor gray organic shale interbeds.

(3) In the upper part, the beds are silty and increasingly argillaceous sandstone with marked development of thin, gray, silty shale interbeds (Swan, 1957), stylolites and other trace-fossils. This upper section contains three diamictite horizons (Carozzi and others, 1973). Two of them were observed in wells by Caputo and Vasconcelos (1971) and Rodrigues and others (1971) (Photo 2). The upper diamictite horizons was mapped by Caputo and Sad (1974) in the Carabinani River, a tributary of the Negro River. They display a clayey and silty matrix with dispersed sand-sized grains,granules and cobbles In the upper part of the section, sideritic, hematitic and chamositic strata are common. The uppermost beds

of the Nhamundá Formation consist of thin, greenish-gray, medium- and coarse- grained, subrounded sandstone beds containing pyrite, green sandy clay partings and siliceous bands or beds.

The Nhamundá Formation is probably resting conformably over the Autás-Mirim Formation. Its upper contact may be diastemic at the basin margins and conformable in the central parts. Although periods of non-deposition are suspected to have preceded and post-dated deposition of the diamictite horizons, no macrofossils or any other indication have been found in the unit to document this.

Daemon and Contreiras (1971a,b) and Lange (1967) dated the Trombetas Group (Autás-Mirim, Nhamundá, Pitinga and Manacapuru formations) as Early Silurian based on graptolites found in the Pitinga Formation shale which overlies the Nhamundá Formation.

Here the age of the Nhamundá Formation is infered as ranging from Late Ordovician to Earliest Silurian because the unit is considered as too thick to be deposited entirely in the Early Silurian together with the Pitinga and Manacapuru formations and because it has diamictite horizons correlatable to Ordovician-Silurian glacigenic beds of northwest Africa.

The unit records a general transgression, during which basal sediments, characterized by fine- and medium-grained sizes, cross-stratified sandstone beds with bioturbated sandstone interbeds, may have been laid down on the upper shoreface. Deltaic as well as tidal deposits, were not identified. Storm swells, and fair weather waves at a smaller scale, may have been the dominant wind-induced physical processes shaping the shoreline deposits, which caused reworking of preexisting aeolian dunes and back shore deposits during the transqression. Minor internal structures characteristic of intertidal deposits (foreshore) were not found, perhaps due to a poor connection with open ocean and to a circumpolar position of the area where tides perhaps were weak.

The middle section represents lower shoreface sedimentation where hummocky cross-stratification points out high storm activity (Harms and others, 1975). It is supposed that in circumpolar regions anticyclonic winds acted with great vigor as they do at present in high latitude periglacial areas.

The upper section is interpreted as representing an alternation between lower shoreface and subareal deposits, with fast fall in sea level during glacial expansions resulting in consequent exposure of sediments previously deposited. The diamictite horizons are interpreted by Caputo and Vasconcelos (1971), Caputo and others (1971), Rodrigues and others (1971), and Rocha-Campos (1981d) as a result of glacial activity, on the basis of rock texture, similar stratigraphic position to that of the confirmed glacial rocks of northwest Africa and Sahara region (Beuf and others, 1971, Rognon and others, 1968) and paleogeographic considerations. Basement clasts were found resting on a subhorizontal substrate (dip of one-half degree) in wells 150 km away from the present position of the basement.

In the center, and in the eastern side of the basin, there is no well control, but it is assumed that offshore conditions predominated during interglacial and pre-glacial episodes in these areas.

In the western part of the basin, towards the Coari arch, bioturbated beds are rare and pink and green beds appear, indicating an increasing estuarine and continental influence throughout the section.

The climate during the deposition of the Nhamundá Formation may have 'been very cold to arctic as evidenced by the absence of warm climatic indicators and the presence of intercalated rocks of glacial nature in the upper part of the section, characterized by repeated alternations of marine and continental glacial and glaciofluvial deposits.

The lower diamictites are in part correlated with the Tamadjert Formation tillites of the Algerian Sahara. The base of the Tamadjert Formation consists of tillites and its top comprises glaciofluvial, glaciomarine and marine deposits (Bennacef and others, 1971). It seems that the upper main diamictites were deposited later in the Amazonas Basin (Earliest Llandoverian) than in the Sahara region (Late Ashgill).

It is interesting to note that prior to the Ashgill Stage, in the pre-Cordillera of San Juan, Argentina, Pacific Realm rocks rich in limestone were laid down (Boucot and Gray, 1978) suggesting a warmer gradient from northwest Africa to Argentina.

PITINGA FORMATION

The Pitinga Member was formally proposed by Lange (1967) to designate a section, belonging to the Trombetas Formation, composed of shale, siltstone and sandstone beds with some iron oolites throughout.

Caputo and others (1971) subdivided the section into two members: the Pitinga Member consisting of the basal shale with siltstone and minor sandstone interbeds and the Manacapuru Member consisting of sandstone, with some siltstone interbeds.

Here both members are raised to the rank of formations belonging to the Trombetas Group.

The Pitinga Formation is a good marker with a known maximum thickness of about 280 m, being easily recognizable throughout the basin. Figure 32 shows the isopach contour map of the Pitinga Formation. It consists of green-gray to dark gray, micaceous, laminated, soft, carbonaceous, pyritic, partly sideritic shale with interbedded shaley siltstone. Toward its top, and near basin margins, very fine-to fine-grained thin sandstone beds are present, as well as some chert interbeds which occur on the northern flank. Hematite interbeds are present at the base, and siderite interbeds and nodules are found throughout the unit.

The unit rests conformably on the Nhamundá Formation, and onlaps it at the basin edges. The Pitinga Formation onlaps the underlying units and rests directly over the Prosperança Formation or Uatumã Group on the basin flanks. In the southern basin margin, the Pitinga black shales overlie directly the Uatumã basement with no intervening sandstone or basal conglornerate beds between them at several places. This suggests a very fast transgression under low energy conditions. Towards the Coari high the unit pinches out between the Nhamundá and Manacapuru formations (Figure 33). Its upper contact is conformable with the Manacapuru Formation and at different places along the basin edges is disconformable with the Maecuru (Devonian), Monte Alegre (Carboniferous), Alter do Chão (Cretaceous), and Solimões and Almeirim (Tertiary) formations.

The Pitinga Formation has yielded graptolites, chitinozoans, algae, sponge spicules, brachiopods, mollusks, scolecodonts, foraminifera, crustaceans and other fossil groups. The graptolites were observed at the base of the section (Caputo and Andrade, 1968).

The graptolite *Climacograptus innotatus* Nicholson var. brasiliensis

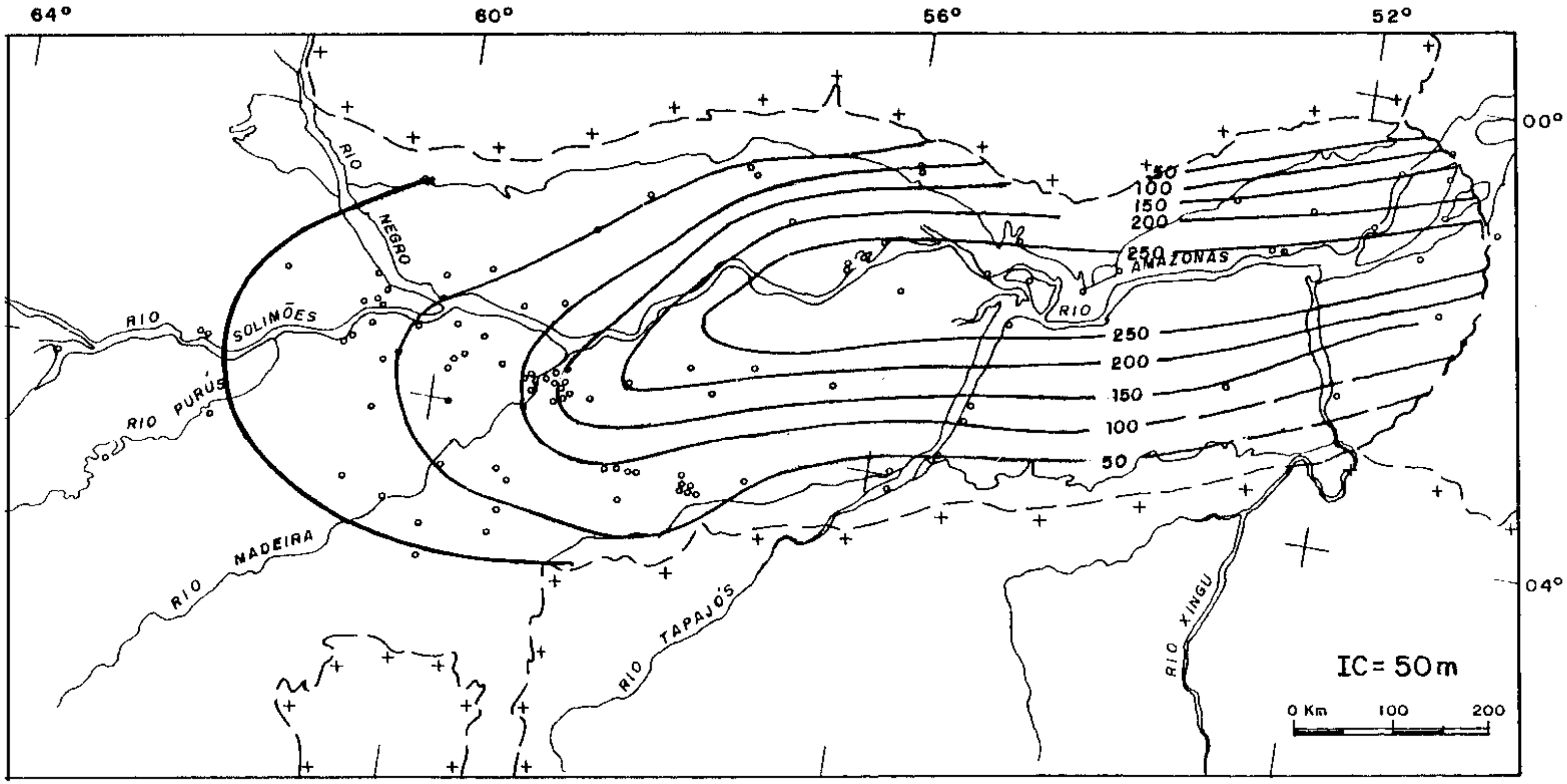


Figure 32 Isopach map of the Pitinga Formation - Amazonas Basin

Ruedemann assigns an Early Silurian age. In Scotland, the *Climacograptus innotatus* is found associated in zone of *Monograptus gregarious* of the Lower Birkhill Shale of the late Early Llandoverian to Middle Llandoverian age (Harrington, In Berry and Boucot, 1973). A late Early Llandoverian to Mid-Llandoverian age is also assigned to the Pitinga Formation. The presence of *Anabaia paraia*, *Tentaculites trombetensis*, *Heterorthella* sp. and *Climacograptus innotatus* Nicholson var. *brasiliensis* in the Trombetas and Caacupe (Paraguay) groups indicates close correlation between both lithologic groups.

The Pitinga Formation is interpreted as a result of predominantly muddy shoreface and offshore deposition during the transgression resulting from melting of widespread ice caps in north Africa and South America. The well-developed lamination and fine-scale bedding are suggestive of relatively quiet water. Some thin sandstone beds in the lower part of the formation contain *Lingula* and *Arthrophycus*, suggesting a shallow nearshore to paralic deposition. Some sporadic shale-pebble horizons up to 5 cm in thickness are present in the section and are interpreted as intraformational conglomerate resulting from glacial activity or episodic storms in a normally quiet environment. The melting of ice sheets may have slowed the anticyclonic wind in the area, reducing the environment energy.

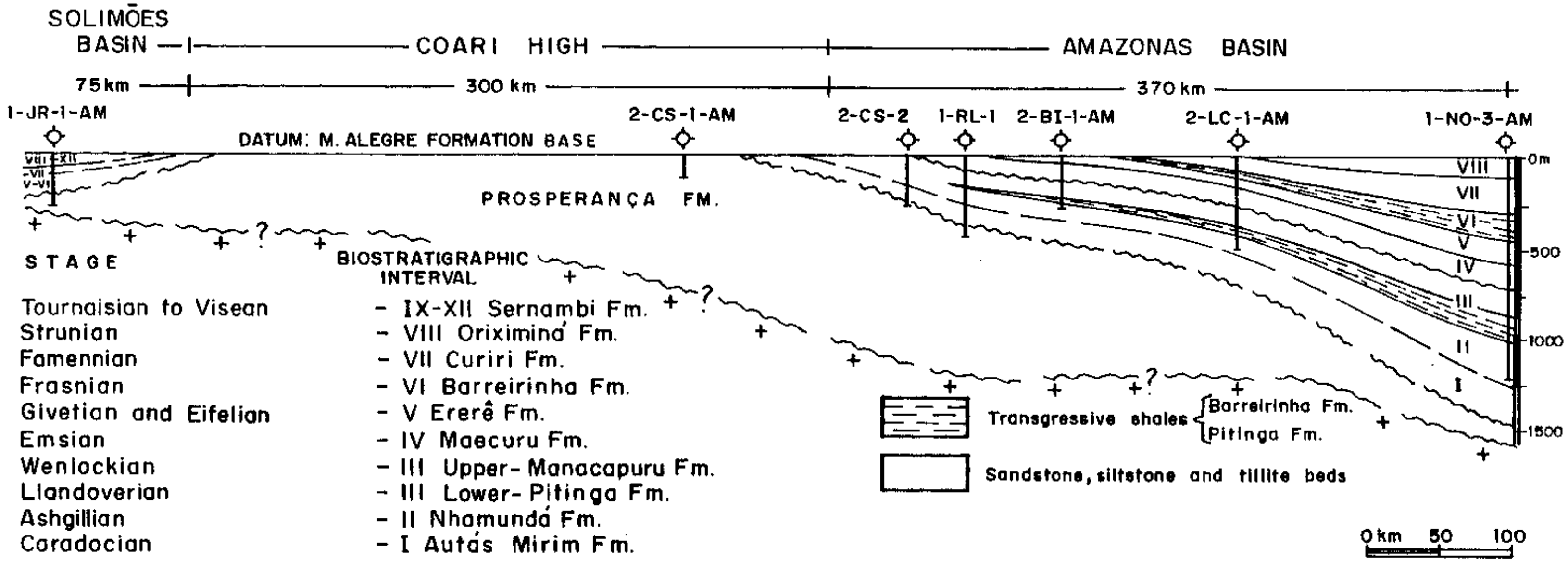


Figure: 33 Ordovician, Silurian and Devonian beds under the pre-Monte Alegre unconformity at the Coari high. All units pinch out against the Coari high indicating the western boundary of the Amazonas basin. There is a paraconformity between Silurian and Devonian Formations.

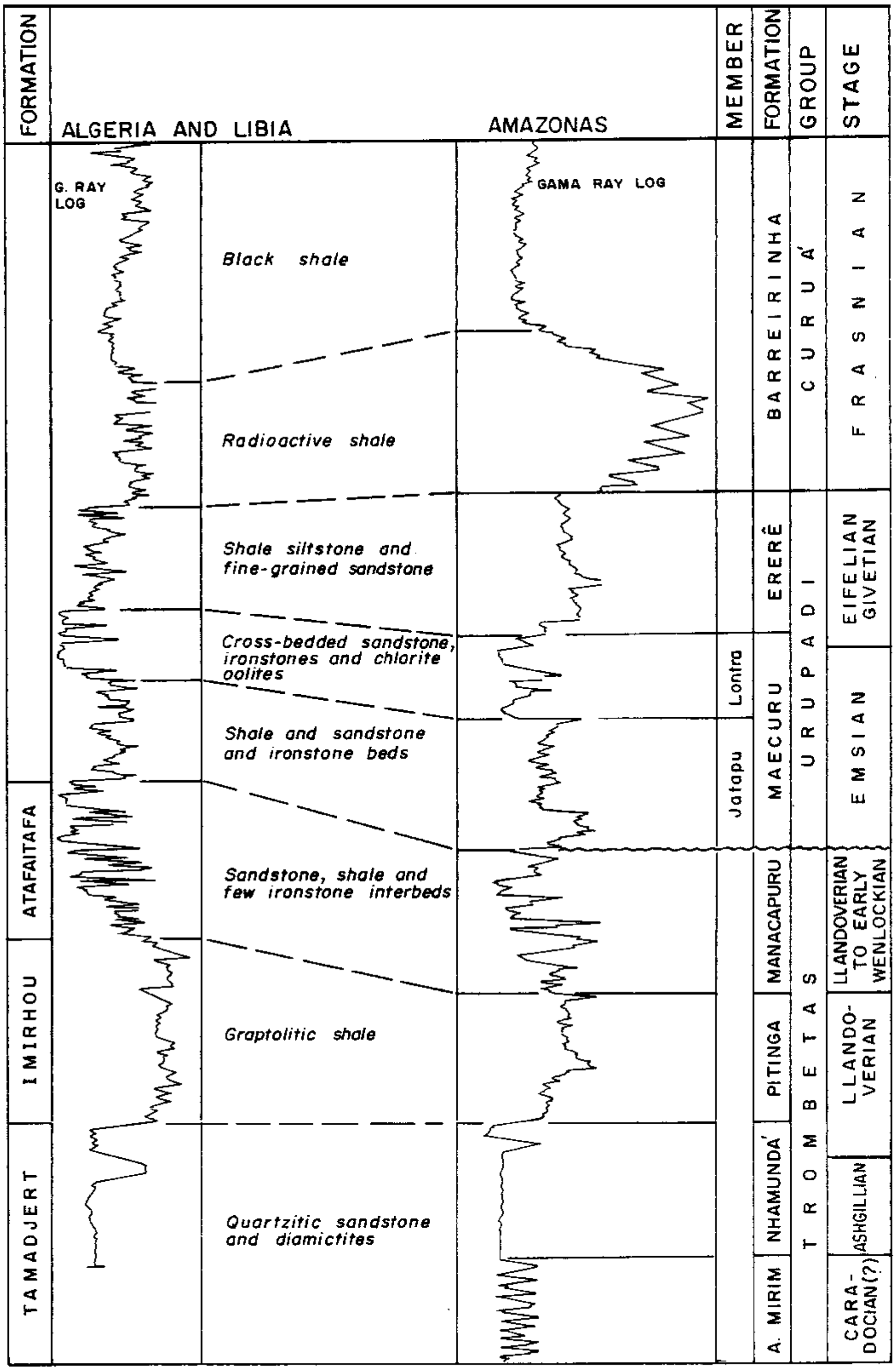


Figure- 34 Well-log correlation between Siluro-Devonian rocks from Central-Saharan and Amazonian Basins

It is interesting to note that in northern Africa the rock equivalent to the Pitinga Formation (graptolitic shales) has its base dated as early Mid-Llandoverian in the Adrar (Trompette, 1973) and Mid-Llandoverian in Zemmon (Sougy, 1964). Figure 34 presents a good borehole correlation between Amazonas and Sahara basins. A Middle Llandoverian age is assigned to transgressive shales in the eastern Coari high flank corresponding to the climax of the Llandoverian transgression in the world. In the Algerian Sahara the Early Silurian "argiles a graptolites" are named Imirhou Formation (Bennacef and others, 1971). Figure 35 presents the distribution of the Early Llandoverian shales in South America and Africa.

The climatic warming may have enhanced chemical weathering which produced a large amount of clay minerals. Transgression may have caused the deposition of coarse clastic sediments in estuaries and river valleys, resulting in important sedimentation of silt and clay in the starved seas during the Pitinga time. In the basal part of the Pitinga Formation illite predominates while in the middle and upper parts kaolinite dominates (Rodrigues and others, 1971) indicating that a change in land weathering from glacial to post-glacial climate may have occurred during its deposition. The clay minerals contain a low boron content (Carozzi and others, 1973) indicating low salinity compatible with great

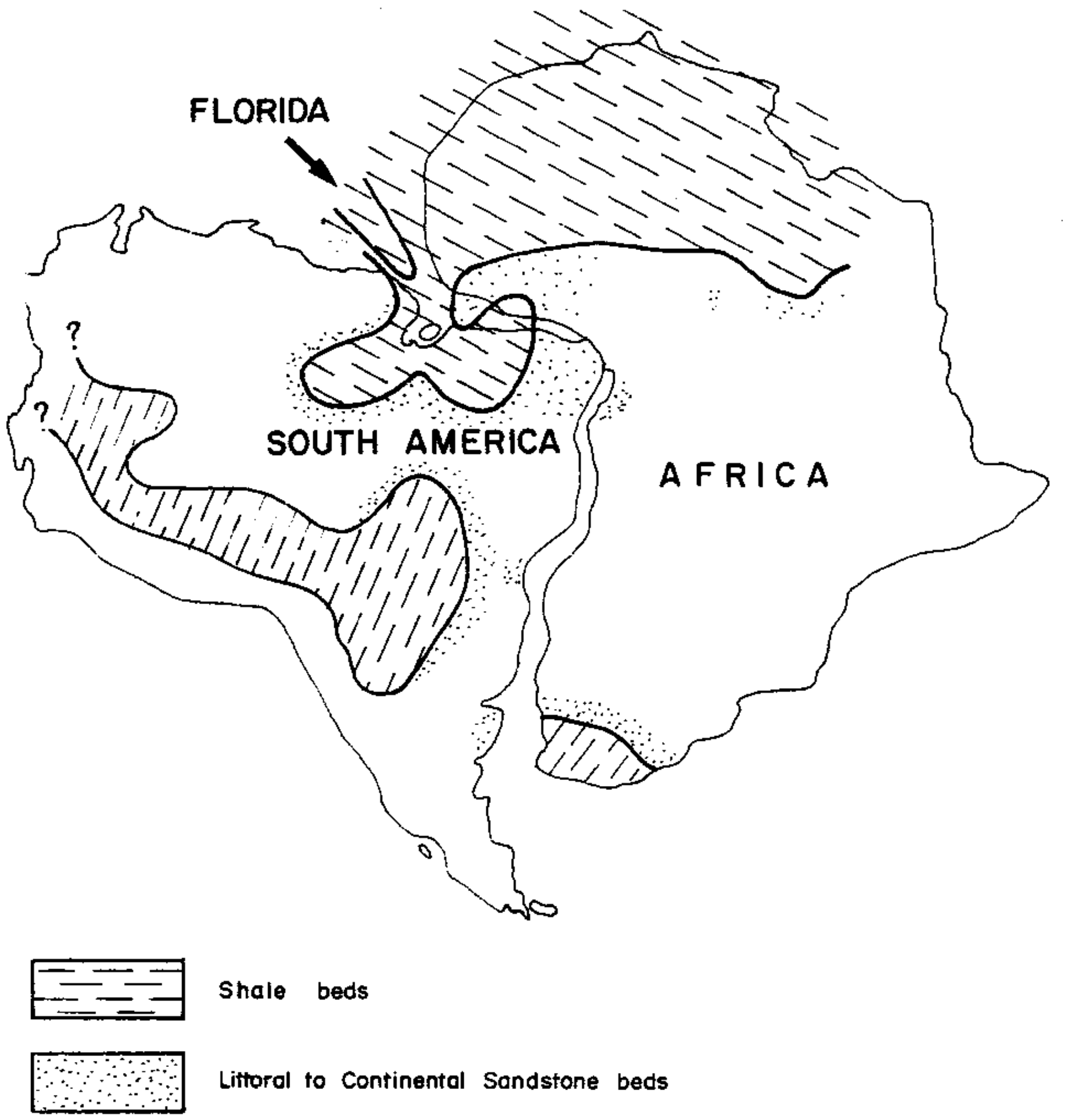


Figure-35 Distribution of Early Silurian shales in western Gondwana, indicating widespreading transgression. North African shale distribution based on Berry and Boucot (1967) and Beuf and others, (1971) and South African shale distribution based on Rust (1973) fresh water ·influx into the basin.

Spjeldnaes (1961) and Berry and Boucot (1967) proposed Ordovician and Silurian paleogeographic reconstructions for Europe and northern Africa based on fauna suggesting very cold climate in the Saharan region and northwest Africa. The Amazonas Basin which may have been linked to northwest Africa may certainly have experienced the same cold climatic conditions. Lange (1970, in Cramer, 1971) also pointed out the similarity between the Silurian chitinozoans of Florida, northern Brazil and northwestern Africa, suggesting close paleogeographic relationship between these areas in Silurian times. At that time, Florida may have been located in Gondwana in the cleft between Africa and South America (Cramer, 1971).

MANACAPURU FORMATION

The Manacapuru Member was proposed by Caputo and others (1971) to designate the upper unit of the Trombetas Formation. Here this member is raised to the category of formation belonging to the Trombetas Group. Its maximum thickness is estimated to be over 200 m in central parts of the basin. The unit is exposed in the northwestern outcrop belt and was uplifted there in Carboniferous time. In the south outcrop belt the Manacapuru Formation is unexposed because is overlapped by Devonian sandstones.

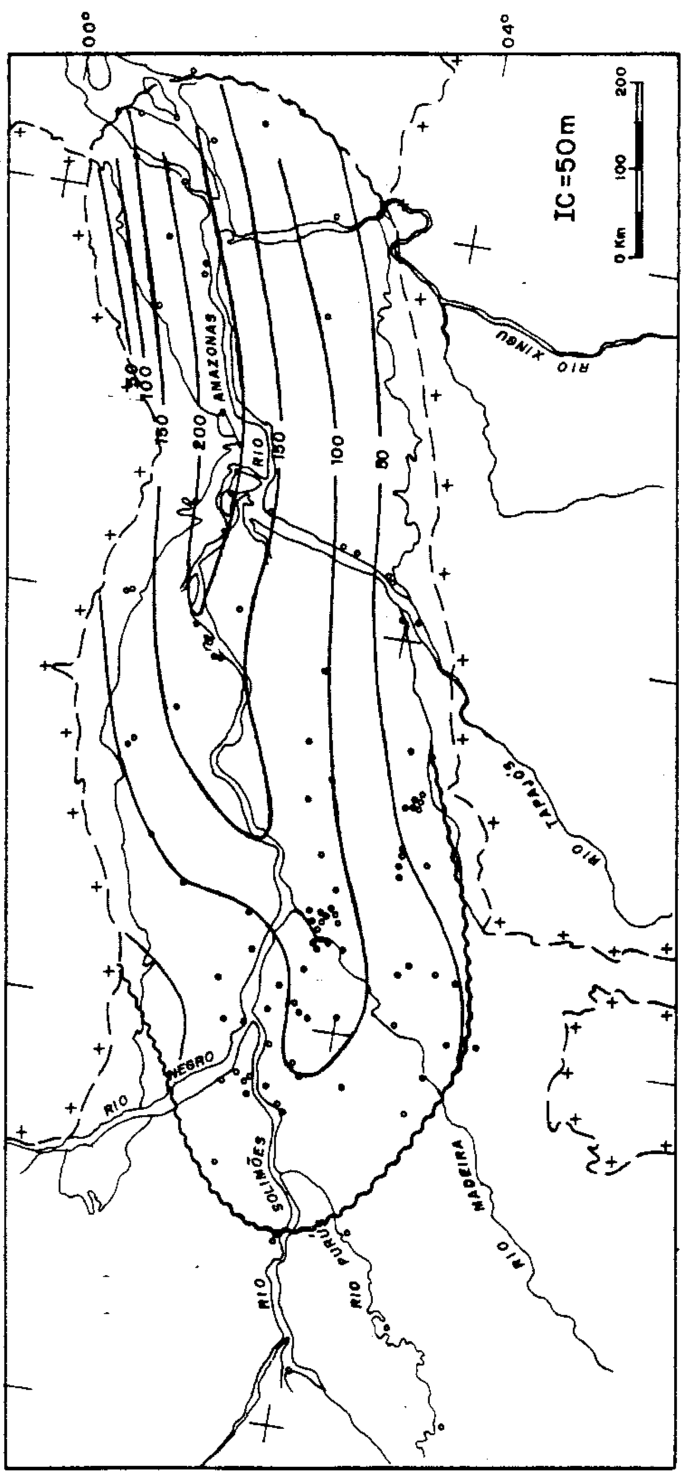


Figure 36 Isopach map of the Manacapuru Formation - Amazonas Basin

Figure 36 shows the isopach contour map of the Manacapuru Formation.

The Manacapuru Formation consists of sandstone beds with shale, siltstone and ironstone interbeds. In the exposures, the formation comprises white and buff fine- to very fine-grained well sorted, friable, highly cross bedded sandstone beds with brownish-gray very soft argillaceous shale and siltstone interbeds. In the subsurface, the sandstones are mainly parallel laminated, fine, micaceous, bioturbated, with beds 5 to 20 cm in thickness, interrupted by siltstone interbeds. Three main sandstone bodies separated by shale-siltstone bodies are present in many wells.

In the westernmost part of the basin fine- and medium-grained cross-stratified sandstone beds are present in many places. Sandstone beds with a clay-ferruginous matrix and oolitic ironstone beds (siderite, chamosite, hematite) are well developed in the margins and western part of the basin. Small white peloids of cellophane and kaolinite-chamosite are scattered through the oolites. The upper third of the Manacapuru Formation consists of as much as 20 m of bioturbated brown sideritic mudstone with a few layers of ferruginous micaceous siltstone containing scattered, white phosphatic kaolinite peloids.

The Manacapuru Formation rests apparently conformably on the Pitinga Formation and is unconformably overlain, with low angle, by the

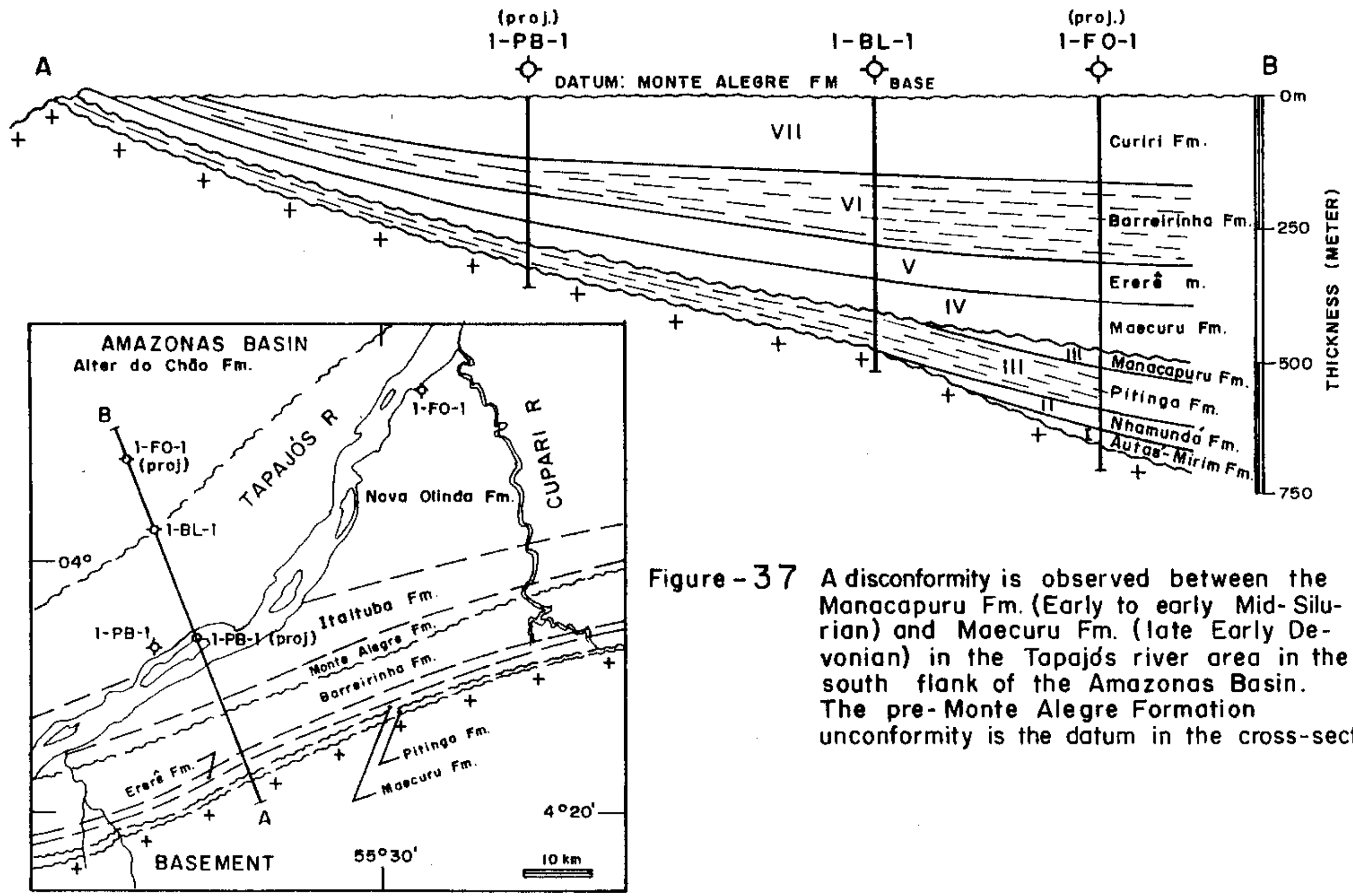


Figure - 37 A disconformity is observed between the Manacapuru Fm. (Early to early Mid-Silurian) and Maecuru Fm. (late Early Devonian) in the Tapajós river area in the south flank of the Amazonas Basin. The pre-Monte Alegre Formation unconformity is the datum in the cross-section.

Maecuru Formation (Figure 37), in the entire basin and by the Monte Alegre and Alter do Chão formations in basin extremities.

At the base of the Manacapuru Formation, the presence of *Arthrophycus* sp. and *Lingula* sp. (Lange, 1967) suggests shallow, sometimes brackish, nearshore water environment. The unit was dated as Early Llandovery by Lange (1967) and Daemon and Contreiras (1971a,b) based on palynological studies. However, they did not find any guide fossil for a detailed dating of the section overlying the graptolitic Pitinga shale. Here, on the basis of its regressive character and correlation with northwest African and Saharan sections, which were relatively contiguous to that of the Amazonas Basin in Silurian time, the Manacupuru Formation is inferred as deposited in the Late Llandoverian time at its base. The top of the unit may have been deposited in the Early Wenlockian time (early Mid-Silurian). Worldwide regression continued from the Late Llandoverian up to the Devonian, by which time the shorelines had retreated about 1,500 km from southern Algeria to northern Morocco (Berry and Boucot, 1973).

The Manacapuru Formation records a general regression in the basin with some transgressive oscillations. In the westernmost part of the basin, fine-grained, cross- and parallel laminated, micaceous argillaceous, bioturbated (skolithos) sandstones are interpreted as deposited in a littoral environment.

Medium-grained, cross-stratified sandstone beds are considered as deposited in deltaic distributary channels and carbonaceous shales may have been deposited in interdistributary lakes and bays. At the north and south flanks, deltaic and fluvial deposits were weakly preserved; only fine-grained, mcaceous, cross-stratified and cross-laminated sandstone beds and shale interbeds were deposited in a low energy shore face environment. In the central parts of the basin there is no well control but an offshore environment is postulated. The presence of ironstones suggests deposition in shallow water at the margins of the basin. Distributary delta abandonment or the onset of transgressive cycles may have inhibited clastic supply resulting in the concentration of ironstones in sand-starved environments. During regressive phases the ironstones may have been remobilized and redeposited together with sandstones in the high energy regressive environment.

According to Hallam and Bradshaw (1979) vegetation and warm humid climate were important factors in the formation of Mesozoic ironstones. However, in northern South America and northwest Africa, where no land plants existed at that time, the Lower Silurian and Lower Devonian ironstone deposits were laid down in high latitude under cold climate. In Northwest Territories, Canada, Young (1976) also reported the presence of Precambrian iron-formations intimately associated with glacigenic deposits. This is a strong indication that ironstones may be deposited in cold as well as warm climates, and that tropical land vegetation is not required for ironstone formation. For many investigators, the limestone-type texture of iron formations points out to an origin of iron formations by limestone replacement (Kimberley, 1974; Dimroth, 1979). However, there is no evidence for such a replacement in the Siluro-Devonian ironstones of Brazil, but it seems that clay may have been originally replaced by siderite or chamosite. Many isolated clay chips in sandstone beds now are fully sideritized.

The Manacapuru Formation is correlated lithologically with the Atafaitafa Formation (Bennacef and others, 1971) of the Algerian Sahara. The Atafaitafa unit also contains three sandstone bodies and two intercalated shale bodies and a number of ironstone interbeds, some of which were interpreted as paleosols (Bennacef and others, 1971).

The climate at that time was cold periglacial as indicated by the presence of ice-caps in the Guaporé shield (Paraná Basin, Maack, 1947; Andean Basin, Crowell and others, 1980, 1981; South Africa, Cape Basin, Rust, 1973). The Andean tillites of the Cancañari Formation were dated as Late Llandoverian by Berry and Boucot (1972) on the basis of graptolites and brachiopods and as

Early Wenloçkian by Crowell and others (1981) on the basis of stratigraphic position and acritarchs, so the Andean tillites may be as old as the Manacupuru Formation beds.

According to Cramer (1971) the acritarch component of Florida formations and the Amazonas Trombetas Group, as well as lithological content, are almost identical. The large geologic dissimilarities between the currently contiguous Appalachians and Florida suggest that Florida may have been attached to the Guyana shield and to the northwestern African craton, instead of the North American craton, in Silurian times. The similarity of the sedimentation of the Trombetas Group with coeval deposits in the Taoudini and Central Sahara basins (Figure 34) is certainly not fortuitous and may result from their having been in juxtaposition as is shown in figure 35.

URUPADI GROUP

Santos and others (1975) proposed the name Urupadi Group to include the Trombetas, Maecuru and Ererê formations. I do not agree with this subdivision because there is an unconformity between the Trombetas and Maecuru units. Here, the Trombetas Formation was raised to the category of group, and considering the unconformity between Silurian and Devonian rocks the Urupadi Group encompasses only the Maecuru and Ererê formations.

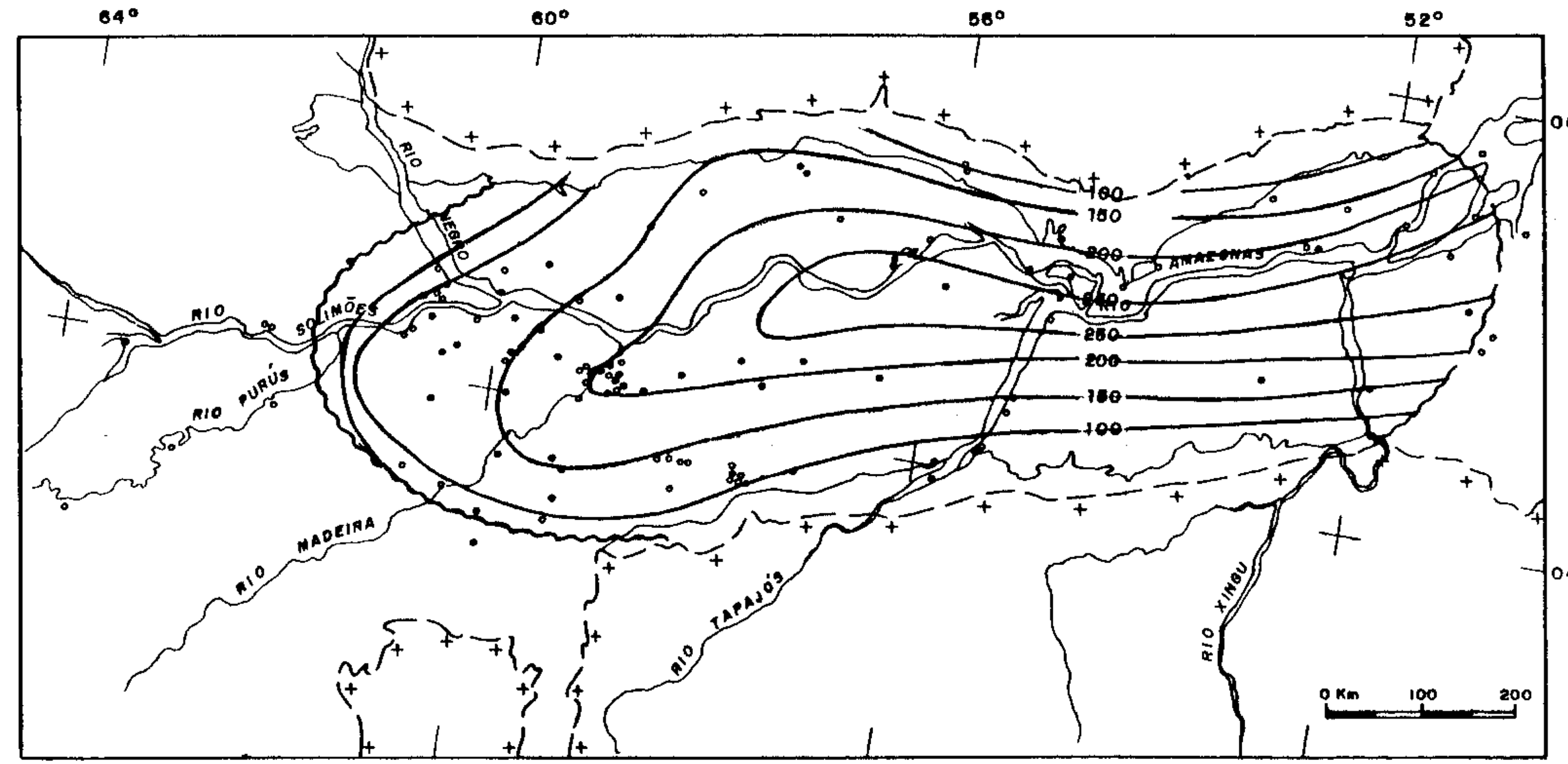


Figure 38 Isopach map of the Maecuru Formation - Amazonas Basin

MAECURU FORMATION

The Maecuru designation was proposed by Derby (1878) to refer to sediments of Devonian age situated above the Trombetas Group.

Lange (1967) formally subdivided the formation into the lower Jatapu Member and the upper Lontra Member. Its maximum thickness is estimated to be over 250 m in central parts of the basin. The Jatapu Member interfingers laterally and grades upward to the Lontra Member. The Lontra Member of Caputo and others (1971) includes a basal section of the overlying Ererê Formation as defined by Lange (1967). Figure 38 shows the isopach contour map of the Maecuru Formation.

The Jatapu Member, as much as 200 m thick, is characterized by gray, highly micaceous, fine-grained, thin-bedded, sideritic, highly bioturbated sandstone beds locally with regular and irregular shale and siltstone interbeds. Some medium-grained sandstone beds as well as ferruginous shale, hematite and siderite beds are mainly present in the basal part of the sequence. Ironstones and bioturbation occur throughout the section indicating a low sedimentation rate during deposition of the Maecuru Formation.

The Lontra Member, as much as 150 m thick, consists of white to light gray, strongly cross-stratified fine-grained to conglomeratic sandstone beds with a few siltstone interbeds.

The Maecuru Formation overlies unconformably the Trombetas

Group and is conformably overlain by the Ererê Formation and in the Coari arch flanks unconformably by the Monte Alegre Formation and in the western and eastern basin flank extremities by Alter do Chão and Almeirim formations.

The unconformity between the Trombetas Group and the Maecuru Formation is a paraconformity and was discovered with the aid of fossils and regional well log correlation. Several geologists have not detected the unconformity in the field due to the absence of basal conglomerates, erosional surfaces or dfffferent dips at the Silurian-Devonian contact. However, correlation among several boreholes and surface rocks as well as paleontological and palynological data establishes the presence of a very low angle (one and half degree) regional unconformity (Figure 37).

Derby (1878) assigned the Maecuru Formation to the Early Devonian based on the brachiopods *Amphigenia elongate* and *Spirifer duodenaria*. The presence of *Tropidoleptus* indicates a Middle Devonian age. Lange (1967) reviewed the fossil content and the pertinent literature, concluding that the Lontra Member was laid down in the Late Emsian and Early Eifelian stages, that is, during late Early Devonian and earliest Middle Devonian. The basal Jatapu Member may be somewhat older than Late Emsian, and it is characterized by soft-body fauna. Up to the present, no shelly fauna have been found in the Jatapu Member. Lange (1967) placed the Maecuru Formation in the local biostratigraphic interval IV and in part of the V, considering its age as ranging from Emsian to Earliest Eifelian. A striking characteristic of this biota is its North American mid-continent rather than Appalachian affinities (Caster, 1952; Copper, 1977).

The Maecuru Formation records the initial Devonian deposition in the Amazonas Basin, as it took place in southern northwest Africa and southern Sahara after a hiatus of about 20 m.y. between (Wenlockian) middle Silurian and Emsian (late Early Devonian).

From the southern flank of the Amazonas Basin, facies of the lower part of the Lontra Member grade into the Jatapu Member facies towards the central parts and the northern flank of the basin. The Jatapu Member consists of almost fully structureless, strongly bioturbated, micaceous sandstone beds with fine-laminated micaceous sandstone interbeds, suggesting low current activity and very low rates of deposition, perhaps below fair-weather wave base. A few subhorizontal stratified coarser sandstone beds may indicate sporadic storm activity. In the eastern outcrop belt, the Jatapu Member consists of bioturbated gray silt-shale beds indicating a deeper water environment than that in the western part of the basin. This is in agreement with an important epeiro-genic uplift in the eastern side of the basin since Carboniferous times. Such uplift may have triggered the removal of shallow water sediments far away from present basin margins, resting deeper water sediments in that region. The presence of intense bioturbation in the section, rich in siderite matrix, indicates low sedimentation rates. This, as well as its paleogeographic setting, rules out the possible action of tidal or oceanic currents. According to Stewart and Walker (1980) meteorological currents, probably storm-generated bottom-return flows, seem to be the most likely process for the occasional introduction and more rapid deposition of coarser sand in environments normally dominated by bioturbation. In the northern outcrop belt, the Jatapu Member shows deeper water deposits, suggesting that shoreline sediments, deposited far away from the present outcrop belt area, may have been removed by erosion.

The Lontra Member consists of very coarse- to medium-grained sandstone beds and represents a major fan delta system progradation interrupted by a fast, short-lived transgression followed by another fan-delta system progradation. The fan-delta concept was developed by Fisher and McGowen (1967), Fisher and others (1969), Fisher and Brown (1972) and Brown (1973). The fan-delta consists of an accumulation of coarse debris brought down by braided rivers which debouch in the sea, with a proximal deposition of essentially sandy-conglomeratic material grading basinward into medial and distal sandstone and shale beds.

In the southern outcrop belt medium-grained to conglomeratic cross-stratified sandstone beds with paleocurrents of toward N60ºW (Caputo and Andrade, 1968) are interpreted here as derived from a braided fluvial system. The coarse delta-fan sediments change to fine- to medium-grained fan-delta front sandstone toward the depositional axis of the basin and progressively to highly bioturbated siltstone and shale profan and offshore beds. A fast transgression caused the deposition of thin beds of bioturbated marine shale in the fan-delta and fan-delta front. Renewed progradation caused the deposition of another fan delta, displaced basinward. The fan-deltas are river-dominated, suggesting low energy in the marine depositional milieu.

In the upper part of the formation, fan-delta front and fan-delta deposits encroached the northwestern and the south flanks of the basin as a result of sea-level fall. Only in the north flank marine fossils were found in fan-delta front sandstone beds. These fossils make up the entire fauna described in the formation.

During the deposition of the Maecuru Formation the climate may have been cold, suggested by the lack of carbonates, evaporites, red beds, reefs, and aeolian desert sands. The Maecuru marine fauna suggests cold climate, but not so cold as in the Paraná Basin, which was classified as subarctic by Copper (1977). No glacial-derived beds were observed during the deposition of the Maecuru Formation. In the Lontra Member the presence of coarser sediments

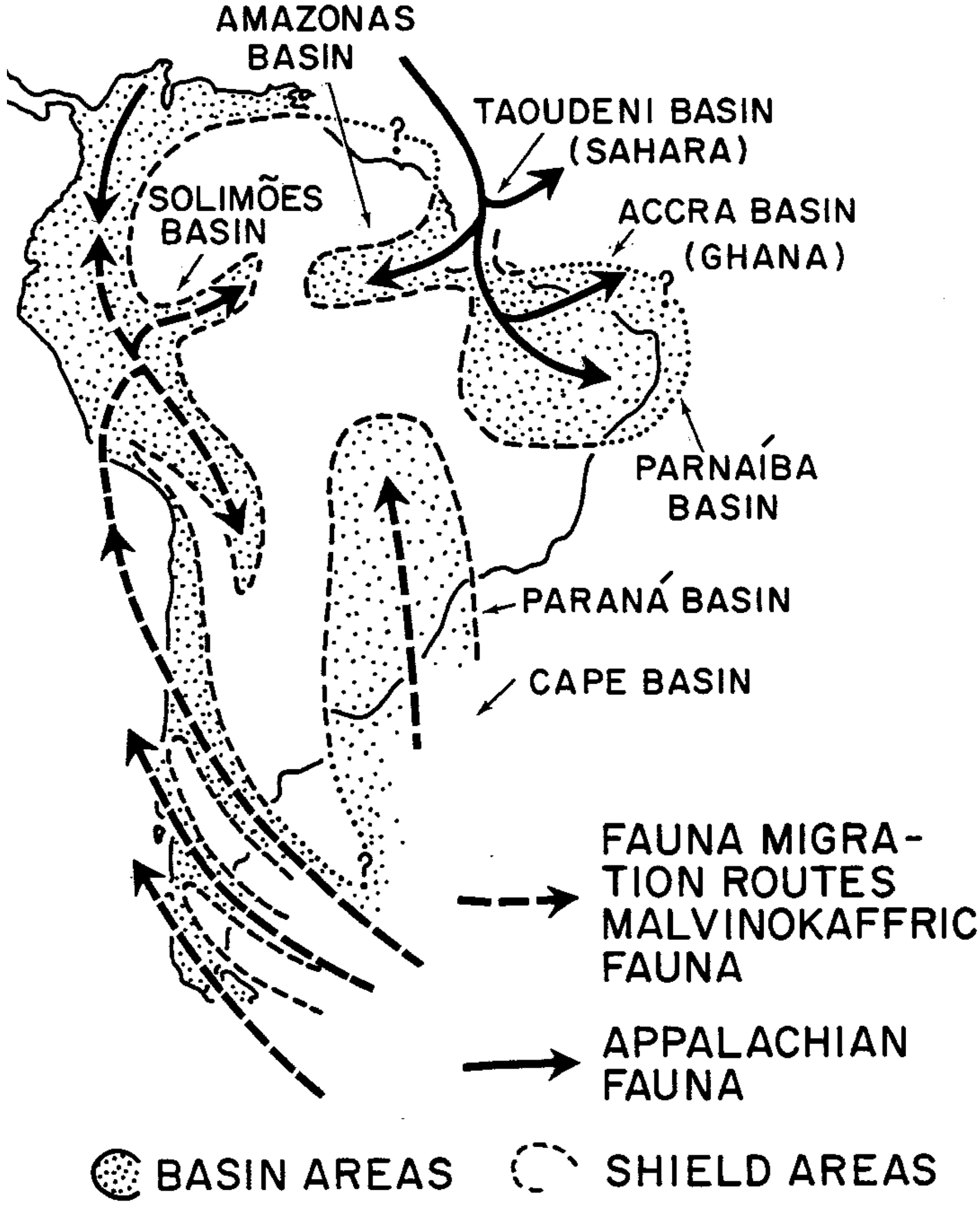


Figure 39 Paleogeographic map of South America during Emsian and Eifelian Stages (Modified from Harrington, 1967).

is due to a general sea-level fall rather than to tectonism in the source areas as interpreted by Carozzi and others (1973).

In the Solimões Basin the oldest Devonian rocks, presently found in the deepest parts of the basin, were deposited in the Givetian Stage (early Mid-Devonian), while in the Amazonas Basin, the oldest Devonian sediments were deposited in the Emsian Stage. This has a great paleogeographic significance because this indicates that the Amazonas sea did not transgress from the Andean area to the Amazonas Basin as previously suggested by paleogeographic reconstructions (Harrington, 1962, 1968). In the Amazonas Basin the Devonian transgression may have entered by the east side of the basin. The fauna of the basin is related to that of North America and northwest Africa (Caster, 1947a,b, 1952).

According to Hollard (1967), in the Taoudini Basin an invasion of North American fauna occurred in Emsian and Givetian stages. This may indicate that the Amazonas and southwest Taoudini basins were contiguous because they were colonized by the same North American biota (Figure 39).

ERERÊ FORMATION

The name Ererê was proposed by Derby (1878) to designate a Devonian section composed of siltstone with shale and sandstone interbeds situated between the Maecuru Formation and the Curuá

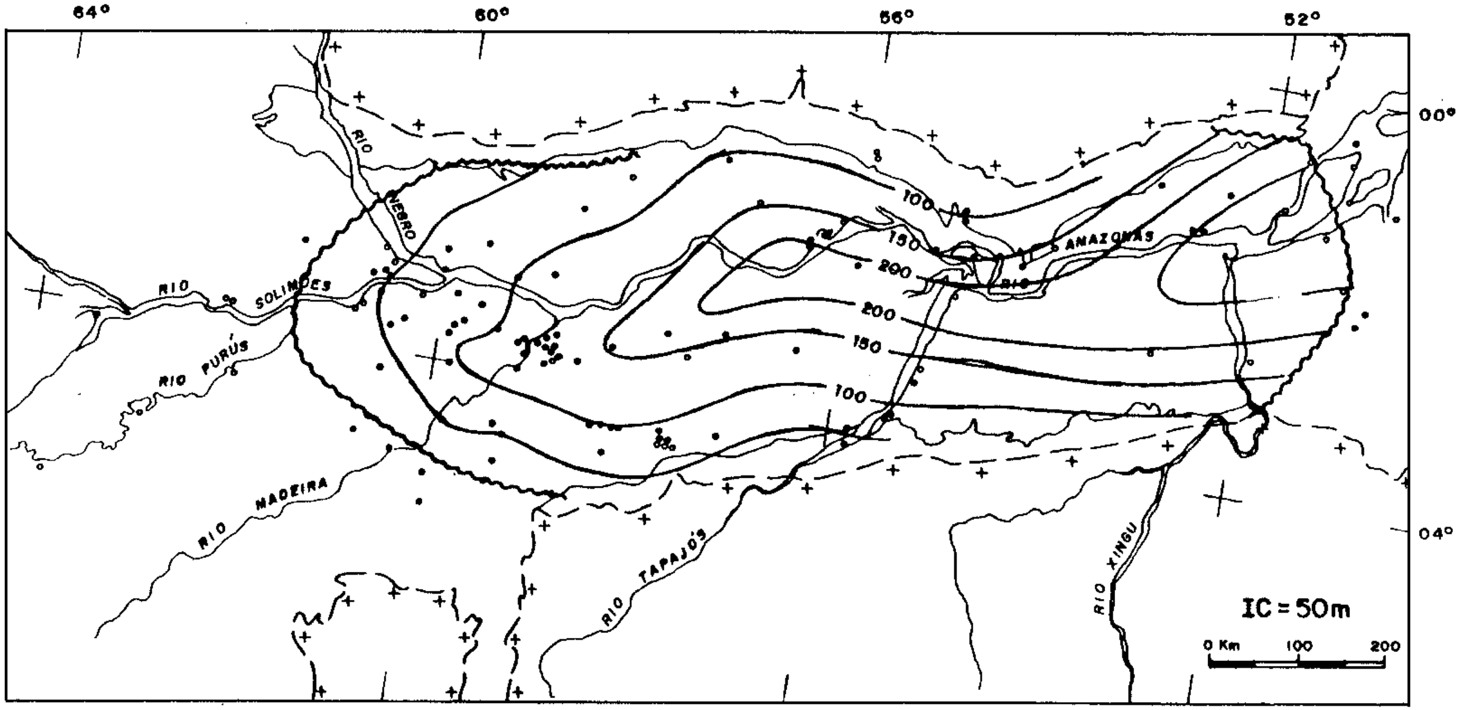


Figure 40 Isopach map of the Ererê Formation - Amazonas Basin

black shale. Bischoff, in 1957, called this section the Ariramba Member of the Maecuru Formation, but Lange (1967) established it as an independent formation. Its maximum thickness is estimated to be over 250 m in the eastern side of the basin. Figure 40 shows an isopach contour map of the Ererê Formation.

The unit comprises greenish-gray, micaceous, carbonaceous, thin-bedded, moderately bioturbated, shaley siltstones which have thin interbeds of light-gray very fine-grained, argillaceous, and minor very thin calciferous sandstone beds. Upwards the sandstone interbeds become more frequent and coarser-grained. In the southern outcrop belt some fine- to very coarse-grained sandstone bodies are well developed. Very thin and sporadic lime-dolomite beds (10 cm in thickness) are present in the northern flank, but sideritic beds or siderite-cemented sediments are rare.

The unit overlies conformably the Maecuru Formation and is conformably overlain by the Curuá Formation. In the Purus arch area, the Ererê Formation is unconformably overlain by the Monte Alegre Formation (Late Carboniferous) and in the outcrop belt extremities it Is overlain by Mesozoic and Cenozoic red beds.

The Ererê fauna is composed of brachiopods, gastropods, trilobites, pelecypods, ostracodes, scolecodonts, and conodonts, most of them collected in the Monte Alegre area. This area is located in a basin flank that was uplifted in Mesozoic times, so the strata show offshore characteristics with a much deeper water fauna than that of the Maecuru Formation that was collected in the north outcrop shallow water area. Several new faunal and floral elements entered the Amazonas Basin, while many components of the earlier Maecuru biota disappeared (Copper, 1977).

The formation was first dated by Rathbun (1874) as Middle Devonian. Its fauna was intimately compared with that of the New York Hamilton sediments due to the presence of about six common species. Lange (1967) and Daemon and Contreiras (1971a,b) positioned the Ererê Formation in the biostratigraphic interval V, based on palynological data and fossil content, corresponding to the Eifelian and Givetian stages.

The Ererê Formation records a larger transgression and deepening in its lower part, and a moderate regression in its upper part, and in the Paleozoic outcrop area the Ererê beds show deeper water conditions than those of the Maecuru Formation. The formation was deposited in river-dominated delta front environments with no significant contemporaneous reworking by marine processes in the south margin. In the southern outcrop belt and in the eastern Coari arch fluvio-deltaic deposits are present, consisting of very-coarse cross-stratified sandstone bodies which may be distributary paleochannels. These change laterally to sandy siltstone of possible bay environment. This rock assemblage changes basinward to massive, bioturbated, fine-and very fine-grained sandstone beds interpreted as delta front deposits. In the middle of the basin, little-bioturbated shale and siltstone beds are considered as deposited in prodelta and offshore environments.

In the western side of the northern outcrop belt, shoreface and delta front facies are developed, while in the eastern side, offshore sediments predominate. In the northern outcrop belt shoreline and proximal deltaic deposits may have been eroded following uplift of the Guyana shield (Rio Branco arch of Amaral, 1974) after sedimentation. In the eastern area of the basin, the Guyana and Guaporé shields uplift was more intense than in other parts of the basin.

The absence of some fossil groups in the Ererê Formation was explained by Copper (1977) as related to either cold climatic conditions or to poor sampling. The presence of very thin dolomite beds, and some calcitic cementation in Ererê sediments, as well as the deposition of limestone beds for the first time in the Taoudeni Basin since the Ordovician Period, point to a climatic warming in the area in the Mid-Devonian time, although the Paraná Basin was under colder conditions. At Mid-Devonian time, the temperature gradient was opposite to that of the Late Ordovician time when the colder area was located in the southern Sahara.

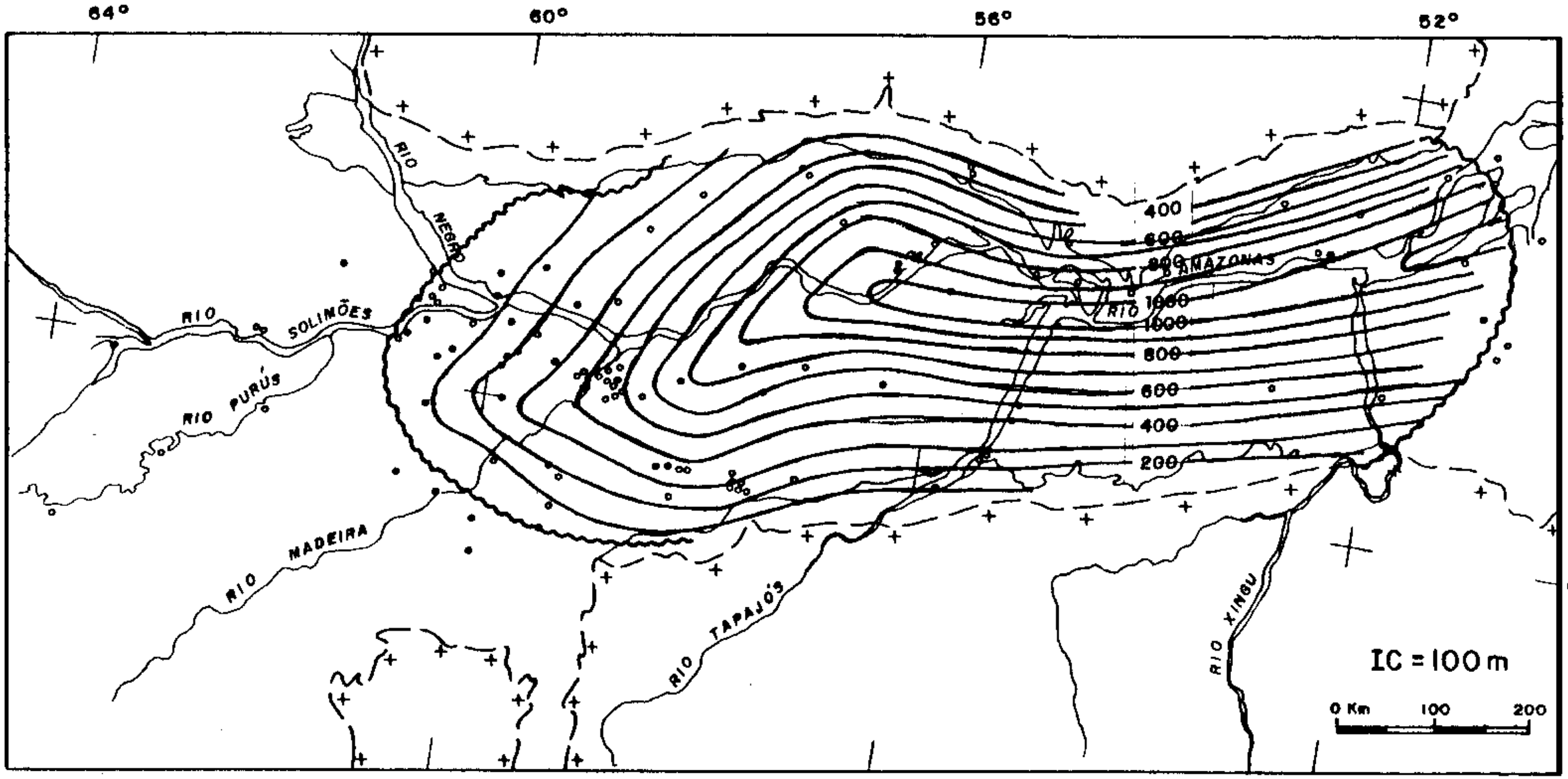


Figure 4 | Isopach map of the Curuá Group - Amazonas Basin

The presence of rocks deposited in fluvial and deltaic environments in the Coari arch suggests that land existed westward, preventing a marine connection between the Amazonas and Solimões basins.

CURUÁ GROUP

The name Curuá Group was proposed by Derby (1878) to refer to Devonian black shales overlying the Ererê Formation. Later the Curuá Group was considered as the Curuá Formation, and was subdivided into several members. Here the Curuá unit is again considered as a group subdivided into three formations, from the bottom upwards: Barreirinha, Curiri and Oriximiná.

The Barreirinha Formation is the best marker within the basin because of its high radioactivity, as detected in gamma ray logs. Its maximum thickness is estimated to be over 1,100 m in the eastern part of the basin. Figure 41 shows the isopach contour map of the Curuá Group.

BARREIRINHA FORMATION

The name Barreirinha was informally used at first by De Carvalho (1926) to encompass Devonian black shale beds cropping out along the Tapajós River banks. Figure 42 shows the isopach contour map of the Barreirinha Formation. The unit is present in

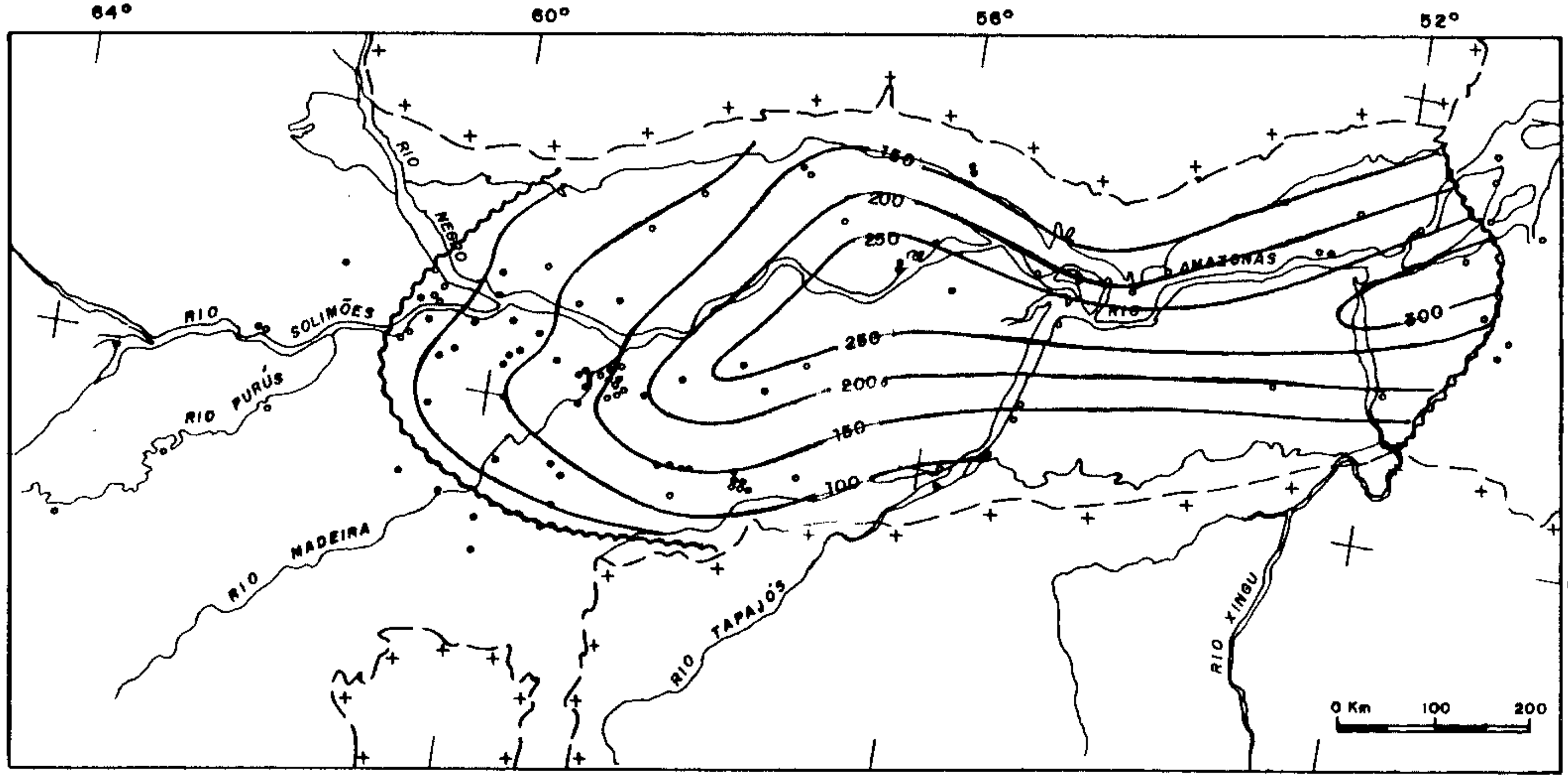


Figure 42 Isopach map of the Barreirinha Formation – Amazonas Basin

both outcrop belts and in the subsurface. It consists of uniformly strongly micaceous, pyritic, carbonaceous, bituminous, fissile (at the base), laminated dark gray to black shale beds with minor thin, silicified siltstone and very fine-grained sandstone interbeds 5-15 cm in thickness. The unit shows a high radioactive level in the lowermost part as well as two horizons with large concretions composed of Ca, Fe and Mn carbonates. Some concretions are about 2 m across. Approximately 280 m in thickness, the formation is extrapolated to the central parts of the basin, and in the subsurface the unit thins towards the Coari high flank more steeply than it does toward the marginal outcrops. The lower radioactive zone has the highest content of organic matter in the basin (up to 10%) as well as high contents of base metals such as S, Cu, Mo and V (Rodrigues, 1971). The most prevalent clay minerals are Kaolinite and illite.

In the well 2-LC-1-AM, the upper part of the Barreirinha Formation was considered as being composed of siltstone and medium-grained sandstone beds. These beds were interpreted by Carozzi and others (1973) as subaquatic channels filled by the action of turbidity currents. A new well (I-LC-2-AM) has shown that the coarse-grained massive section belongs to the overlying Curiri Formation, so that the Barreirinha Formation thins as it approaches the Coari high flanks without changing its facies. At least for the time being, no such coarse sediments were detected in the Barreirinha Formation. This sedimentary unit shows the same characteristics as those of the Chattanooga Shale of the interior basins of the United States.

The Barreirinha Formation rests conformably on the Ererê Formation and it is conformably overlain by the Curiri Formation. On the Coari arch flank, and in both western outcrop belt extremities, the unit is disconformably overlain by the Monte Alegre Formation (Upper Carboniferous) and in the eastern outcrop belt extremities, it is overlain by the Almeirin Formation (Tertiary).

The Barreirinha Formation is very poor in macrofossil, and bioturbation is absent because the black shale environment is not conducive to a normal benthic marine biota (Copper, 1977). The unit has yielded a very few brachiopods, conodonts, scolecodonts, and mollusks, fish teeth and spines, but it is rich in palynomorphs such as chitinozoans, acritarchs and sporomorphs. Important early Late Devonian (Frasnian Stage) fossils which date the formation are the brachiopods *Orbiculoideia lodensis* Hall and *Schizobolus truncates* Hall, also common to the Genesee Formation of New York State. The overlying basal Curiri Formation has yielded Early Famennian (mid-Late Devonian) plant fossils which could confirm the Frasnian age for the Barreirinha Formation. Palynological data also validate an early Late Devonian age (Daemon and Contreiras, 1971a,b).

The Barreirinha Formation, composed of black laminated shales, records the deepest water environment in the history of the basin. The black shales represent a sediment-starved anoxic basin related to a major worldwide transgression that may have swamped vast areas of deltas and coastal plains. During this transgressive phase, deposition of terrigenous coarse clastic sediments was at a minimum, that laid down was emplaced in drowned estuaries and river valleys. Coarse sediments may have been trapped in river valleys and were not able to reach the basin. It is possible that the Barreirinha shorelines were relatively muddy, similarly to the present Black Sea. In the Amazonas Basin outcrop area, the black shale sediments represent an offshore environment.

The Barreirinha transgression correlates very closely with the Taghanic onlap observed by Johnson (1970) in several basins in the United States or with the Frasnian transgression (Lunulicosta zone transgression) across Russian platform (House, 1975) and across northern Africa (Hollard, 1967).

Shaley transgressive deposits of the same age are registered in the Parnaíba (Andrade and Daemon, 1975), Paraná (Lange, 1967; Daemon and Quadros, 1970) and elsewhere in Gondwana (Figure 94). This is contrary to the idea suggested by Johnson (1979) that Gondwana was not extensively flooded by the sea due to epeirogenesis as the Euramerican platforms were flooded during Givetian and Frasnian times. The paucity of benthonic fossils is Interpreted as due to a high stress anoxic environment in the Amazonas and Solimões basins, lethal to most macroorganisms.

In the Solimões Basin, from Middle Devonian to early Late Devonian times, a sandy transgressive unit (Juruá Formation) was laid down with characteristics very distinct from those of the Barreirinha black shales. It is presently unknown whether the Solimões and Amazonas basins were connected during the Frasnian Stage. The Barreirinha Formation thins very sharply toward the Coari high, suggesting no communication between the Amazonas and Solimões basins. However, if any marine communication has existed between the Amazonas and Solimões basins it may have occurred during the Frasnian time when the transgression reached its maximum.

The upper part of the Barreirinha Formation is richer in silt-sized particles and less radioactive and bituminous than its lower part. The more abundant silt-sized particles may be attributed to an increase in aeolian activity in the source areas during a general increase in gradient temperature which could have promoted strong winds, or to an increase in bottom water circulation caused by the sinking of colder and dense oxygenated less saline water in the basin margins in response to a climatic deterioration. High influx of poor saline water, low evaporation and probable narrow connection to the ocean may explain the low content in boron in the Barreirinha illite clay minerals (Carozzi and others, 1973). The relatively low salinity environment is consistent with a very cold climate. The same kind of large Ca, Fe and Mn carbonate concretions (1 to 2 m across) found in the cold water Saharan graptolitic Silurian shales (Beuf and others, 1971) are present in the Late Devonian Barreirinha Formation.

If it is considered that the climate was glacial-arctic during the deposition of the overlying Curiri Formation, the upper part of the Barreirinha Formation may have been subarctic cold.

CURIRI FORMATION

The term Curiri Member was proposed by Lange (1967) for a Devonian section composed of diamictite, shale, and siltstone beds occurring at Igarapé Curiri (Curiri Creek). As much as 350 m in thickness is extrapolated to the deepest parts of the basin. In this stuy the Curiri Member is raised to the category of formation. At the time that Lange (1967) proposed the Curiri unit, its beds were only known in the type-section and in the subsurface. Caputo and Andrade (1968) mapped all the southwestern outcrop belt area, where they found that the unit is exposed eastward from the Nambi River up to the Cupari River. Exposure descriptions provided by Bemerguy (1964) and Macambira (1977) indicate the presence of this formation in the southeastern outcrop belt, so it is exposed in a belt of about 500 km in length in the south flank of the basin.

In the north margin, the unit is recognized along a few rivers, but possibly more exposures exist and they will be found with more detailed mapping. The diamictite beds are found in most boreholes which reached the underlying Barreirinha Formation, but in some wells, mainly in the central parts of the basin, the formation is only represented by shale and siltstone beds, most of them rich in *Spirophyton* trace fossils. Figure 43 shows the isopach contour map of the Curiri Formation.

At the base, the Curiri section is composed of dark gray to light gray, brown or buff silty shale alternating with light gray highly micaceous siltstone and very fine-grained argillaceous sandstone beds, strongly bioturbated by a trace fossil named *Spirophyton.* These bioturbated beds mark the base of the formation. In the central parts of the basin, the same section is composed of micaceous dark gray bioturbated shale. The basal Curiri beds are also characterized by an algae, or primitive land plant, called *Protosalvinia* sp., which marks a well defined zone in the section. For example, the Tracoá River bank outcrops, for a long reach, consist of very fine-grained, micaceous, clayey, cross-laminated sandstone beds with more than 50% of the rock com-

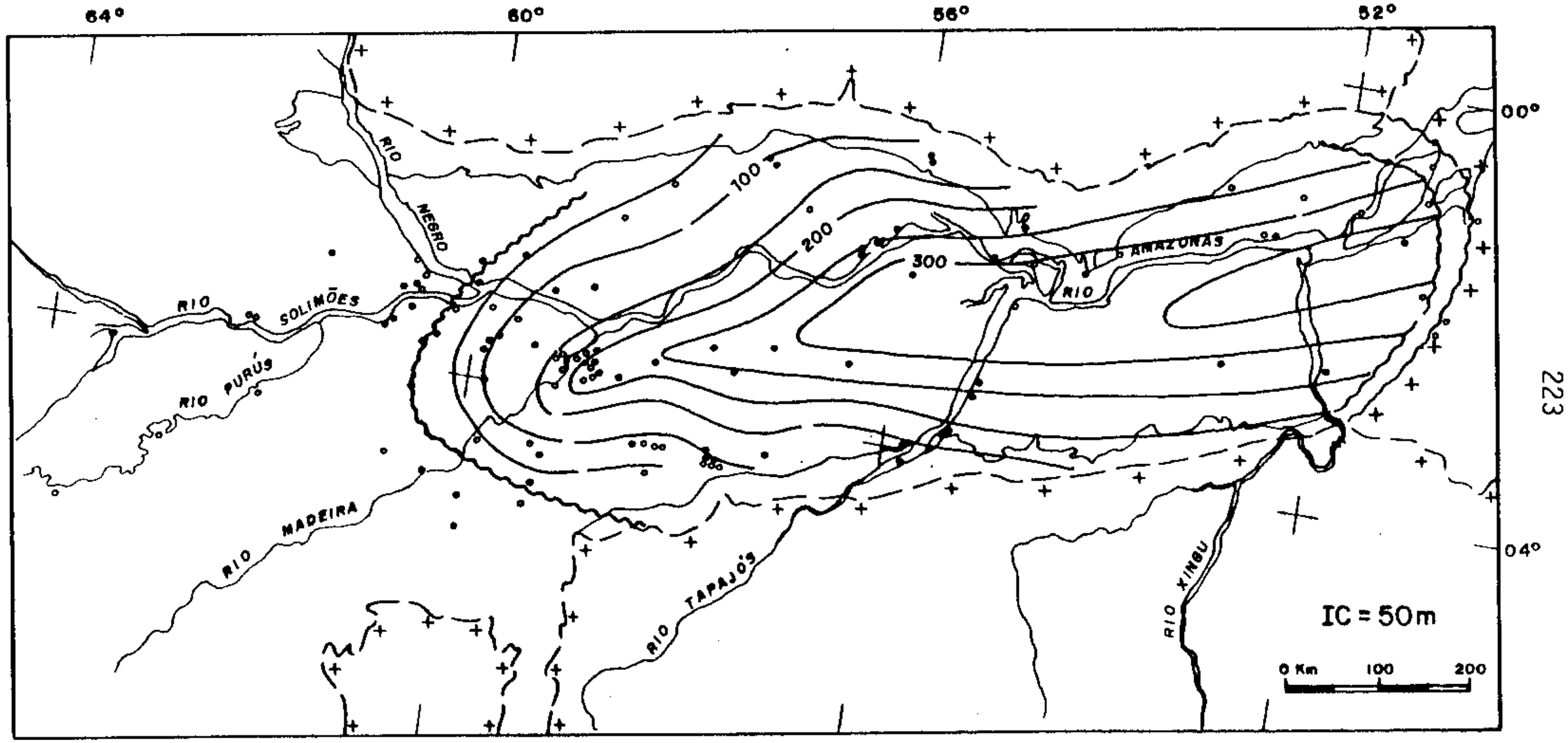


Figure 43 Isopach map of the Curiri Formation - Amazonas Basin

posed of Protosalvinia remains, which seem to be a kind of shoal concentration.

Above the Protosalvinia zone, a micaceous light to dark gray massive diamictite section with scattered sand-sized grains, granules, cobbles and pebbles is present (Photo 3). In the subsurface, some diamictite intervals (Photo 4) change basinward to laminated shale with rafted granules and cobbles and to laminated shale free of clasts. Shale, rhyolite, quartz, quartzite, chert sandstone, basalt, and limestone are the most common clasts found. Some striated pebbles were recovered from the subsurface (Bouman and others, 1960; Rodrigues and others, 1971) and are also found in outcrop (Macambira and others, 1977; Caputo and Andrade, 1968; Caputo and Crowell, in press).

Lenticular bodies composed of very fine- to fine-grained, cross-laminated and cross-bedded sandstone beds are generally deformed and mixed with diamictite masses. The sandstone shows scattered pebbles, microfolds and microfaults. The microfaults place sandstone and diamictite beds in irregular contact. Ludwig (1964) interpreted the diamictite beds as a result of turbidity currents and deformations in sandstone beds as convolute folds related to turbidity currents too. However, according to Reineck and Singh (1980) convolute bedding is remarkably continuous, in spite of the intensively folded internal laminae, so that faulting and slippage are not normally associated with convolutions. Deformations observed in these lenticular sandstones are not here interpreted as convolute folds. The deformed sandstone beds also show cross-stratification that are incompatible with a turbiditic origin as interpreted by Ludwig (1964). No graded beds were found in the Curiri Formation suggesting turbidity current activity.

The Curiri diamictite beds are present in two horizons and are separated by sandstone and shale beds. Diamictites together with sandstone bodies change laterally and upwards to shale with pebbles or to shale, free of clasts, but in some places rich in the trace-fossil *Spirophyton* in the flanks and central parts of the basin.

Thin sections show fairly fresh, angular grains, many are corroded, and ranging in size from silt to coarse sand. They "float" in a silty and clayey matrix and are rarely in direct contact with each other (Photo 5). Most quartz grains are composite, show normal extinction within silica overgrowths, and some are hematite stained. K-feldspar grains may comprise 5 to 10 percent, plagioclase 5 percent, rock fragments 15 to 20 percent, biotite 1 to 3 percent and siderite one percent. Sericite, muscovite and chlorite are the most common minerals in the matrix, which has locally been recrystallized to muscovite. Common heavy minerals are green tourmaline and yellow zircon.

Illite is the main clay mineral found in the subsurface (flanks and in central parts of the basin), followed by kaolinite in the margins (Carozzi and others, 1973).

The Curiri Formation overlies conformably the Barreirinha Formation and it is conformably overlain by the Oriximiná Formation. Along the eastern Coari arch flank, the Curiri Formation is unconformably overlain by the Monte Alegre Formation and also by the Alter do Chão Formation. In the eastern part of the basin the Almeirim Formation overlies unconformably the Curiri Formation. The upper part of the Curiri Formation is also truncated by the Monte Alegre Formation in the outcrop belt, so that the unit is only complete in the basin flanks.

The age of the Curiri Formation was determined on the basis of palynological data and algae fossils. The genus *Protosalvinia* is dated as mid-Late Devonian (Early Famennian) by Phillips and others (1972), Niklas and Phillips (1976) and Niklas and others (1976) and considered as restricted to the Famennian Stage by Gray and Boucot (1979). The forms of *Protosalvinia arnoldii, P. ravena* and *P. furcata* are found in the Amazonas and Appalachian interior basins, while *Protosalvinia braziliensis* and *P. bilobata* have only been found in the Amazonas Basin up to the present. *Protosalvinia arnoldii* was also found in the Parnaíba Basin (Niklas and others, 1976).

In the *Protosalvinia* zone the trace-fossil *Spirophyton* is also present, and characterizes the basal beds of the Curiri Formation. Daemon and Contreiras (1971a,b) dated the section corresponding to the Curiri Formation as Famennian based on the spores *Hystrichosporites* and *Ancyrospora* and forms of *Hymenozono-triletes lepidophytus* Kedo (= *Rotispora lepidophyta* (Kedo) Playford (1976) new combination) that characterize the biostratigraphic interval VII. The presence of the *Convolutispora*, *Vallatisporites* and *Reticulatisporites* genera and the disap-pearance of earlier forms characterize the biostratigraphic inter-val VIII (Late Famennian) corresponding to the Upper part of the Curiri Formation. Daemon and Contreiras (1971a,b) supported their age determinations by reference to Lanzoni and Magloire's (1969) work. Lanzoni and Magloire (1969) studied the Algerian Sahara Late Devonian-Early Carboniferous palynomorphs, linking their palynological intervals to marine macrofossil zones. Owens and Streel (1967) also pointed out that the spore *Hymenozonotriletes lepidophytus* Kedo indicates a Famennian age.

The Curiri Formation records glacial and interglacial periods in the area. Ice-sheets reached the basin in the Mid-Famennian time, but the development of ice-caps may have commenced earlier in highlands. The presence of nearshore very fine sands rich in *Protosalvinia* sp, may record the beginning of a regressive phase

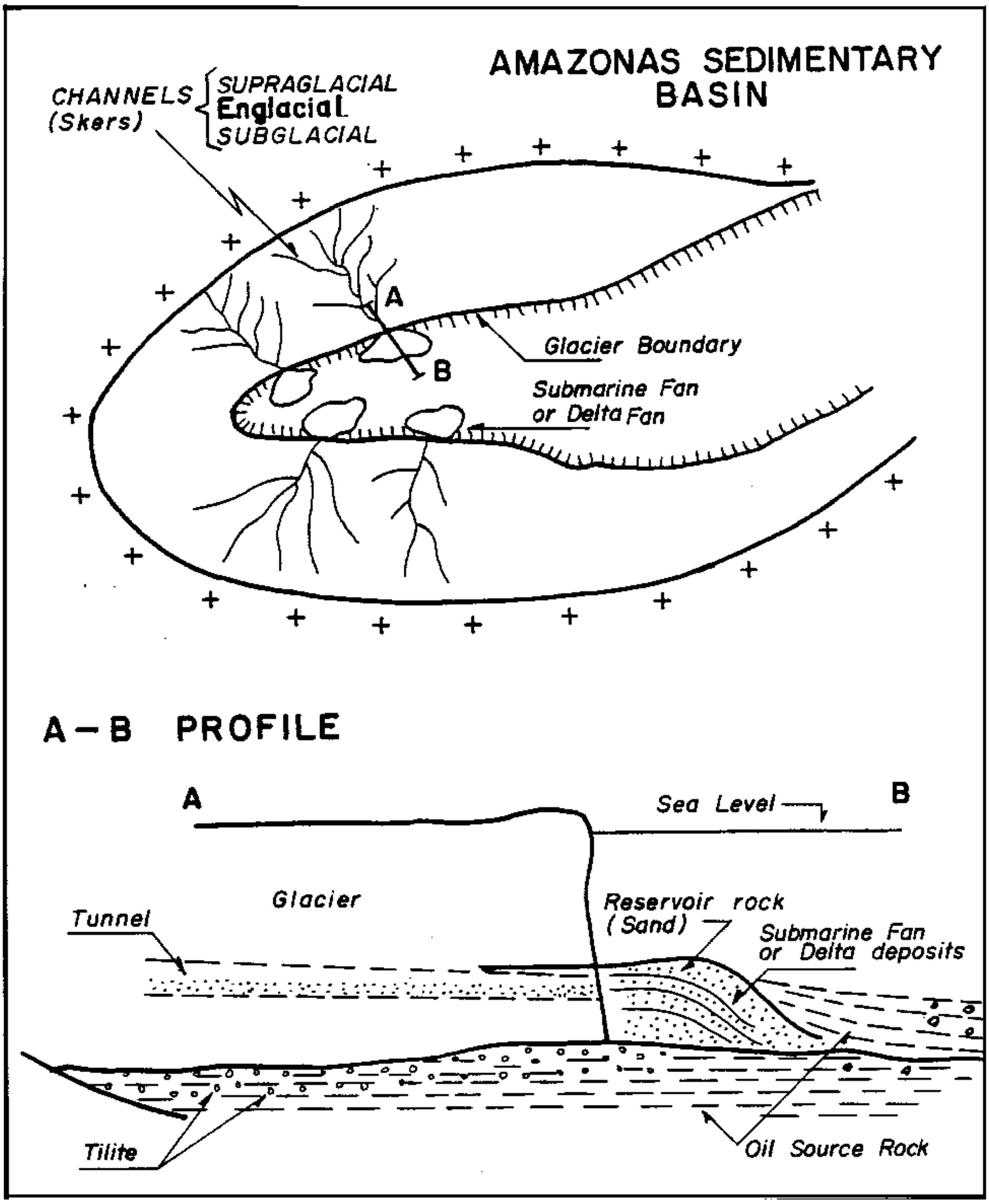


Figure 44. Oil accumulation model for Curiri Formation sands

in the basin. Continental tillites (Photo 3) with striated clasts (Photos 6, 7, 8) were deposited along the outcrop Paleozoic belt area, and in some zones of the basin flanks. Subglacial and englacial channels (eskers) represented by lenticular deformed sandstone bodies were developed in the region (Photo 9). The channels are directed from its edges toward the basin axis, as determined by many wells along basin flanks. It is possible that the subglacial and englacial streams built submarine fans in the central parts of the basin (Figure 44). In the flank areas, glacial and shallow glaciomarine sediments may have been deposited around glacial lobes and in central areas glaciomarine sediments may have been laid down in an offshore environment.

The relatively low salinity environment that started during deposition of the upper Barreirinha Formation persisted during the deposition of the Curiri Formation (Carozzi and others, 1973).

The shield area and the basin may have been depressed by ice-loading, so that the regression was not so pronounced in the Amazonas Basin as elsewhere, although outside the Brazilian basins a fast and large worldwide regression took place. Along the periphery of ice-sheets, a forebulge, supposedly formed due to the outflow of asthenospheric material from beneath the ice sheets, may have also contributed to the non-deposition of sediments during Famennian times in the Andean foreland basins.

In the Sahara region, at that time, a great amount of regressive ferruginous and chloritic oolites as well as hematite beds were deposited (Freulon, 1964). No limestone beds were laid down, probably due to a strong climatic deterioration. In the Late Frasnian and probably in the Famennian stages, in the Taoudeni and Saharan basins, many siltstone beds were laid down which may be interpreted as loess deposits although Nahon and Trompette (1982) considered them as a result of tropical weathering.

In the Sahara region, the Famennian sea retreated more than 1,500 km in relation to the Frasnian sea. Along the western side of northwestern Africa, the Famennian sea retreated about 500 km in a northward direction according to present coordinates (Freulon, 1964) due to the worldwide regression tied up to the Famennian glaciation in South America and Africa (Accra Basin).

The Amazonas Basin was situated close to the Appalachian interior basins during Late Devonian times, because the strong faunal similarities imply a close geographic proximity between these two areas separated by only a narrow ocean Any paleoreconstruction of Gondwana and Laurasia has to take into account the faunal relationship between both areas in the Late Devonian time.

ORIXIMINÁ FORMATION

The Oriximiná Member was proposed by Caputo and others (1971, 1972) for a section composed of thick interbedded light gray, very fine-to coarse-grained sandstone blankets and dark to black shale beds, with subordinate diamictite beds in the lower part of the section. In this work the Oriximiná Member is raised to the category of formation. The upper part of the Curiri Formation interfingers with the lower part of the Orximiná Formation, as the Oriximiná marginal sandstone bodies pinch out toward the center of the basin where Curiri offshore or lagoonal siltstone and shale predominate. Carozzi and others (1973) considered the lower Oriximiná section composed of shale, sandstone and diamictite beds, as belonging to the Curiri Formation, but here the Curiri Formation (with sandstone lenses) and the Oriximiná Formation (with sandstone blankets) are considered as originally defined. Despite the convenience in assembling the Oriximiná section in one unit, I feel that several genetic units are present. Unfortunately it is difficult to subdivide this part of the section because there are only few wells and cores in the central part of the basin. Figure 45 shows the isopach contour map of the Oriximiná Formation.

The Oriximiná Formation, up to 430 m in thickness, consists of white to light gray, pyritic, argillaceous, fine- to coarse-grained laminated or cross-bedded sandstone bodies with bioturbation at the top of the coarse-grained sandstone beds and with widespread bioturbation in the fine-grained sandstone beds. The sandstone bodies show variable thickness. The shales are dark

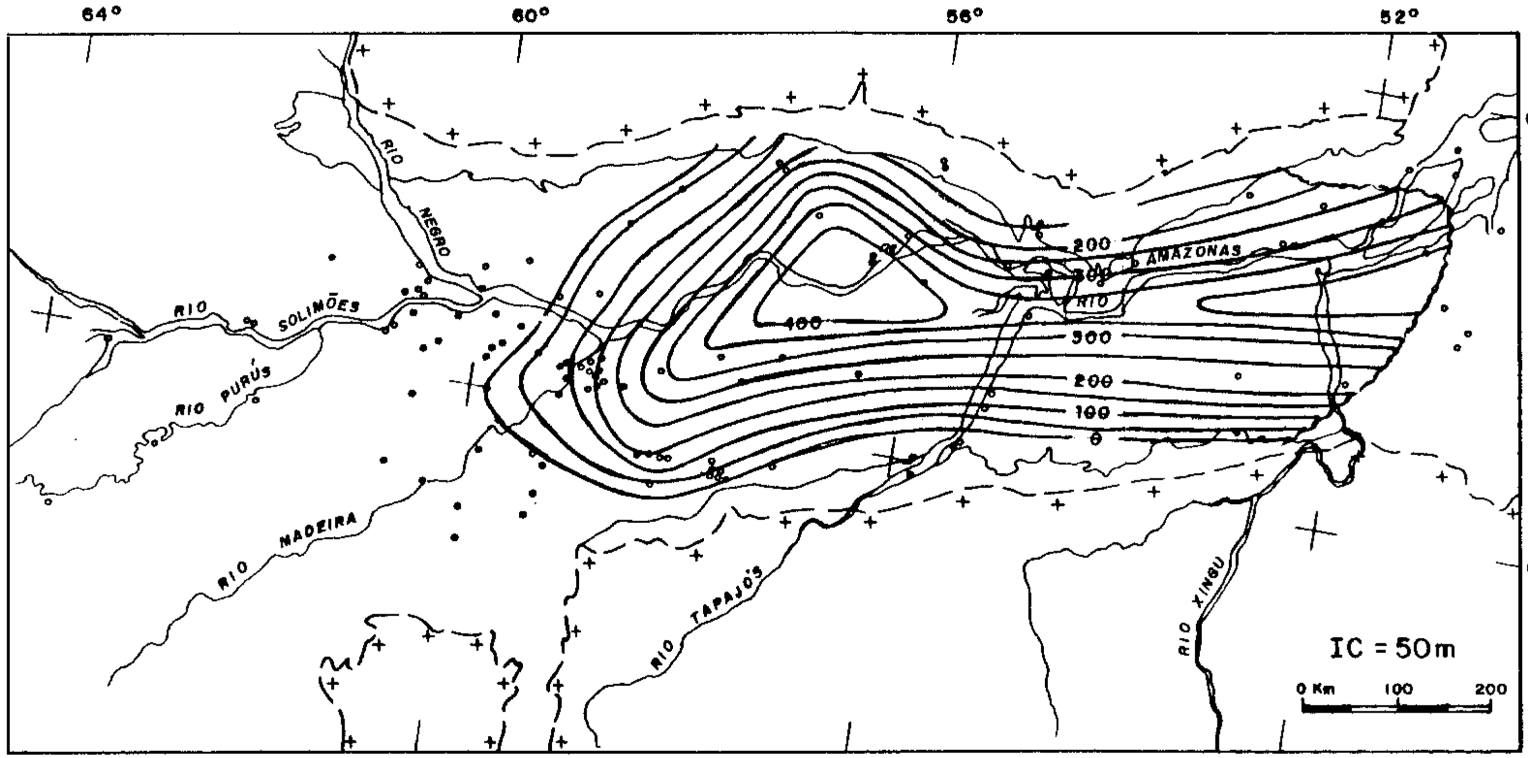


Figure 45 Isopach map of the Oriximiná Formation - Amazonas Basin

gray to black, fissile, micaceous, little bioturbated, with gray siltstone interbeds.

The diamictite is similar to those of the Curiri Formation, but in the Oriximiná Formation they occur only above the lowermost sandstone of the section.

The Formation is not exposed in the outcrop belt, except in The Monte Alegre dome where diamictites were not found. The Oriximiná Formation is present in the subsurface in the entire basin. It overlies conformably the Curiri Formation and it is conformably overlain by the Faro Formation in the central parts of the basin. But, on the basin flanks it is unconformably overlain by the Faro and Monte Alegre formations.

Few macrofossils were recovered from cores. The borehole 1-UA-1-AM at the depth of 3,017 m yielded the fossils *Orbiculoidea* sp., *Chonetes* sp., *Lingula* sp., and *Strophomenaceae* genus and species undetermined. All these fossils are not suitable for dating but they provide some indications about environmental conditions.

Strata defined as Oriximiná Formation were dated as Late Devonian to Early Carboniferous (Late Famennian to Late Tournaisian stages) by Daemon and Contreiras (1971a,b), based on palynology, corresponding to the upper part of the biostratigraphic interval VIII to X, as their paper indicates.

The formation consists of an alternation of restricted shallow-marine or lagoonal, deltaic, and fluvial deposits, with at least one glacial advance recorded in its lower part. Fine- to medium-grained feldspathic quartzitic arenite predominates. It comprises many coarsening-upward sequences 5 to 40 m thick, each of which is thinner and finer-grained toward the basin center. In the lower part of the section, the sandstone bodies grade into siltstones and shales toward the center of the basin where *Estheriae* were recovered. These fossils indicate a fresh to brackish water environment in the central part of the basin. The rarity of acritarchs, some of which could be reworked older forms, is indicative of some intervals with non-marine deposition in the western part of the basin. In the eastern part there is no fossil control, but transitional and restricted marine conditions may have prevailed. The upper several meters of a coarsening-upward cycle consist of fine-grained ripple-bedded sandstone with shallow-marine to brackish fossils such as *Lingula* and *Orbiculoidea* in the central deep parts of the basin.

These sediments are overlain by cross-bedded, coarse grained quartz arenites with coal debris which are interpreted as deltaic distributary channel deposits. Several river systems were developed on the basin flanks. Braided river systems, characterized by thick sand bodies, and also the meander river systems characterized by thin argillaceous sandstone bodies; siltstone and shale interbeds were developed.

At least one glacial advance took place in the basal part of the sequence that can be documented by diamictites overlying fluvial cross-bedded and cross-laminated fine- to medium-grained sandstone beds. The diamictites in the central part of the basin overlie and underlie siltstone and shale beds of the Curiri Formation.

During the deposition of the Barreirinha Formation, plant debris were supplied to the basin, and this supply increased during the deposition of the Oriximiná Formation. In the middle part of the section its regressive character increased; the western part of the basin was probably covered by a meandering fluvial system. The upper part of the section records a general retreat of the fluvial and delta systems in the basin and marine or lagoonal shales were deposited in the inner flanks and central parts. At the Devonian-Carboniferous boundary regressive sandstone beds are present in the Oriximiná Formation.

The lower and middle sections of the Oriximiná Formation correlate with the Longá Formation of the Parnaíba Basin and the upper section correlates with the Lower Poti Formation of the same basin.

The climate during deposition of the lower part of the section may have been arctic glacial, but the appearance of plant debris in the middle section indicates climatic warming and the development of land plants in the basin periphery.

The end of the Strunian Stage (uppermost Famennian) in the Sahara region is also characterized by regressive beds (Conrad and others, 1970), but in the early beginning of Carboniferous time a new worldwide transgression took place.

FARO FORMATION

The Faro Member was proposed by Lange (1967) to designate a thick section located at the top of the Curuá Formation. Caputo and others (1971, 1972) raised the unit to the rank of a formation independent with a boundary above that indicated by Lange (1967). The Faro Formation, up to 330 m thick, consists mainly of two thick sandstone bodies and two shale bodies. Locally, the lower shale body may pinch out, forming a lower very thick sandstone body and an upper very persistent shale body. Figure 46 shows the isopach contour map of the Faro Formation.

The sandstone beds are of mainly fine-grained but locally medium- and coarse-grained, white to light gray, parallel or cross-stratified, and sometimes with thin carbonaceous shale and siltstone interbeds.

The lower shale unit is composed of dark gray to black carbonaceous laminated pyritic shale with interbeds of rich mica-

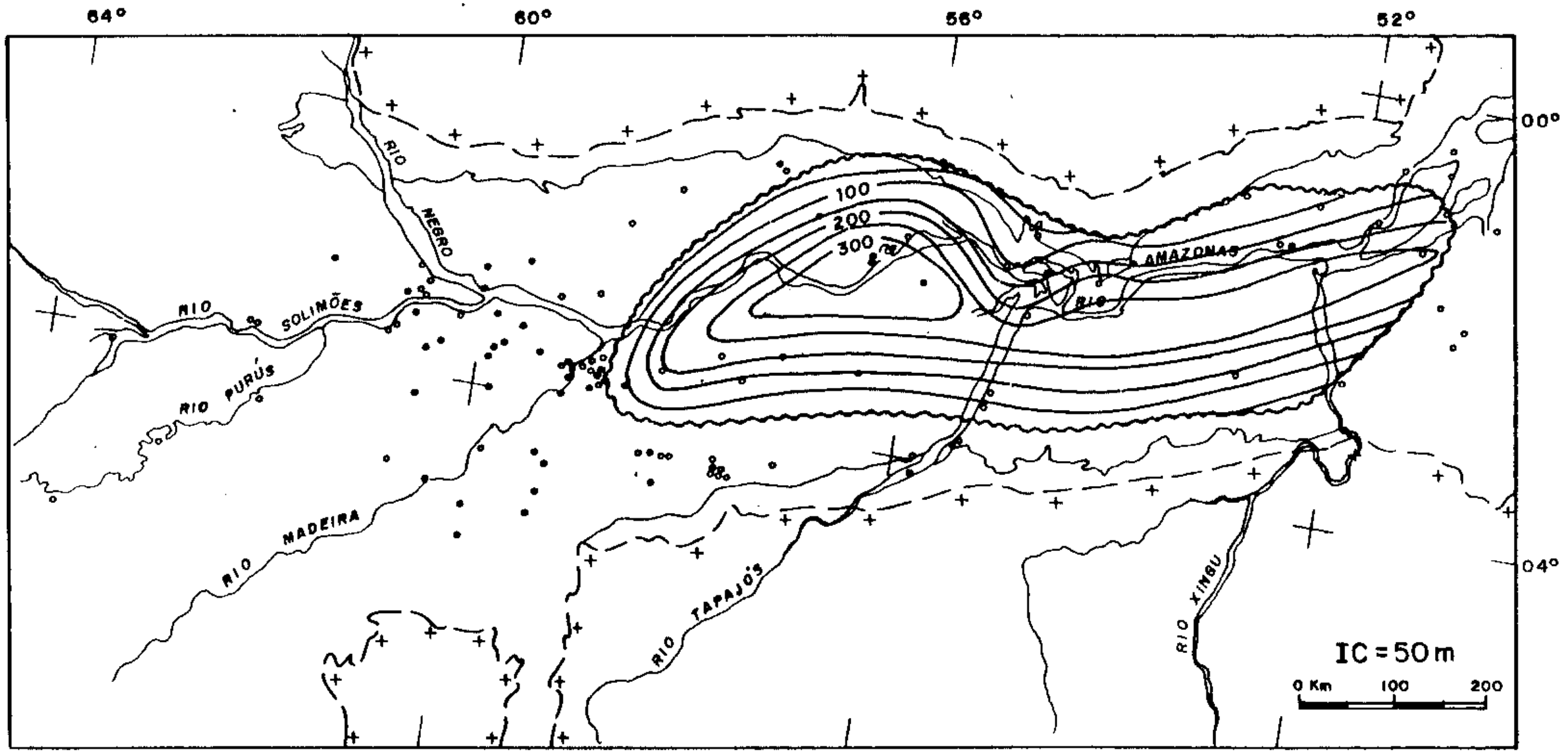


Figure 46 Isopach map of the Faro Formation - Amazonas Basin

ceous sandstone. In this lower shale unit, a diamictite bed, with sand-, granule- and pebble- sized clasts dispersed in a massive, micaceous, silty and clayey groundmass was described in the well 1-MA-1-PA, core 21. The upper shale unit is mostly composed of black carbonaceous, pyritic, laminated shale with some gray siltstone and thin fine-grained sandstone interbeds.

In the middle of the basin and towards its northern flank the Faro Formation overlies conformably the Oriximiná Formation, but along the southern flank, the Faro Formation covers the Oriximiná Formation disconformably and records a period of subaerial exposure and erosion.

Sedimentation continued along the central parts of the basin while erosion took place in its margins for a long time. The Faro Formation is unconformably covered by the Monte Alegre Formation of Late Carboniferous age in the whole basin, except in the Monte Alegre dome, where Mesozoic basic intrusions have uplifted the area and erosion has exposed the Faro Formation.

The age of the formation was determined from data obtained by Daemon and Contreiras (1971a,b) on palynological grounds.

The section comprising the Faro Formation was deposited in part of the local biostratigraphic interval XI and in XII corresponding to the Early Carboniferous time (Visean Stage). The base of the Faro Member of Lange (1967) is older (Tournaisian ?) than the base of the Faro Formation of Caputo and others (1971, 1972). Some degree of interfingering exists between the base of the Faro Formation and the top of the Oriximiná Formation.

The Faro Formation represents a generalized regression during a sedimentary cycle that began in the Emsian Stage (late Early Devonian) and finished in the Visean Stage (mid-Early Carboniferous). These regressive deposits correlate very well with a worldwide drop in the sea level which culminated in the subsequent Namurian Stage (Late Mississippian).

According to Daemon and Contreiras (1971a,b) the acritarchs found in the biostratigraphic intervals XI to XII are reworked. The Faro Formation contains a large quantity of carbonized plant debris as well as coal films and streaks.

The section is interpreted as having been deposited on a proximal alluvial plain in the western part of the basin and around its periphery, where bed-load channel deposits (braided systems) predominate. In the eastern side of the basin, mixed-load channels carried significant amounts of fine sediment and interbedded sandstone bodies dominate the section. In this same side of the basin, the shale fraction increases, indicating the development of suspended-load channel deposits associated with broad flood plains and lakes. Fluctuations in the depositional environment are attributed to changes in base-level of erosion.

The large amount of coarse clastic sediments may be related to the erosion produced by ice-caps in the shield area as well as to the worldwide regression. Much of the Faro Formation may represent periglacial outwash deposits.

The lower shale unit with diamictite beds very similar to those of the Curiri Formation may have resulted from a glacial advance in the Amazonas Basin in Visean time. The upper carbonaceous shale unit is interpreted as lake deposits which onlapped the Faro Formation itself. The shale unit covers part of the Oriximiná Formation at the basin flancs. The apparent absence of a gap in the sedimentation between the Oriximiná and Faro formations in wells in the north flank (well 2-NA-1-Pa) suggests that the northern basin edge was located far away from its present northern boundary in comparison with its present southern boundary where there is a gap corresponding to the top of the Oriximiná Formation. This stratigraphic relationship suggests that the northern Guyana shield area was uplifted some time after the deposition of the Faro Formation. This uplift resulted in removing a large part of the north margin of the basin. The north basin edges extended beyond its present margin.

The drainage to the Phoibic Ocean (ocean between Laurasia and Gondwana formed after the Acadian Orogeny and closed during the Hercynian Orogeny, McKerrow and Ziegler, 1972) may have been interrupted due to an uplift beyond the eastern boundary of the present Amazonas Basin. This uplift may have generated a lake at the end of the Faro Formation sedimentation. The continued uplift in the eastern extremity of the basin directed the drainage system westwards during the Middle Carboniferous time (Namurian Stage), when a widespread erosion occurred in the basin.

The great amount of mica in the sediments suggests mechanical weathering in the source areas, and this is compatible with dry climate. The presence of diamictites of possible glacial origin suggests an arctic to subarctic climate. Moreover glacial rocks in the State of Rondônia and Bolivia in Early Carboniferous (Visean) time hints at the presence of ice-sheets in the Guaporé shield, and that some ice lobes may have extended into the Solimões and Amazonas basins at times of glacial maxima. In interglacial times outwash clastic sediments may have reached these basins. The poor coal development in the unit may suggest that the climate was not humid enough for a good development of coal in the basin, because high moisture is the primary requirement for growth of luxuriant vegetation in either cold or warm climates.

In northern Brazil, the sediments laid down from Late Ordovician to middle Early Carboniferous are deprived of all the generally accepted warm climatic indicators. This suggests that the climate was predominantly cold during the mentioned time.

TAPAJÓS GROUP

The Tapajós Group includes the Monte Alegre, Itaituba and Nova Olinda formations (Dos Santos and others, 1975). I propose to include the Andirá Formation in the Tapajós Group. Therefore, the Tapajós Group consists of the Monte Alegre, Itaituba, Nova Olinda and Andirá formations.Its maximum thickness is estimated to be over 2500 m in central parts of the basin. Figure 47 shows the isopach contour map of the Tapajós Group.

MONTE ALEGRE FORMATION

The name Monte Alegre was informally used at first by Freydank (1957) to describe a Carboniferous section mainly composed of sandstone beds and since then the designation has been accepted by most geologists. Figure 48 shows the isopach contour map of the Monte Alegre Formation. The unit commonly consists of a basal conglomerate composed of granules and cobbles with great petrographic variety. Where the conglomerate is missing the unit is made up of white, cream to light green, with small- and large-scale cross-bedding, kaolinitic, friable, subrounded to rounded, frosted and pitted grains, bimodal, mainly medium-grained sandstone beds. Fine-grained sandstone beds are more common in the subsurface and coarse-grained sandstone beds are sometimes observed in outcrops. In the subsurface, the unit up to 140 m

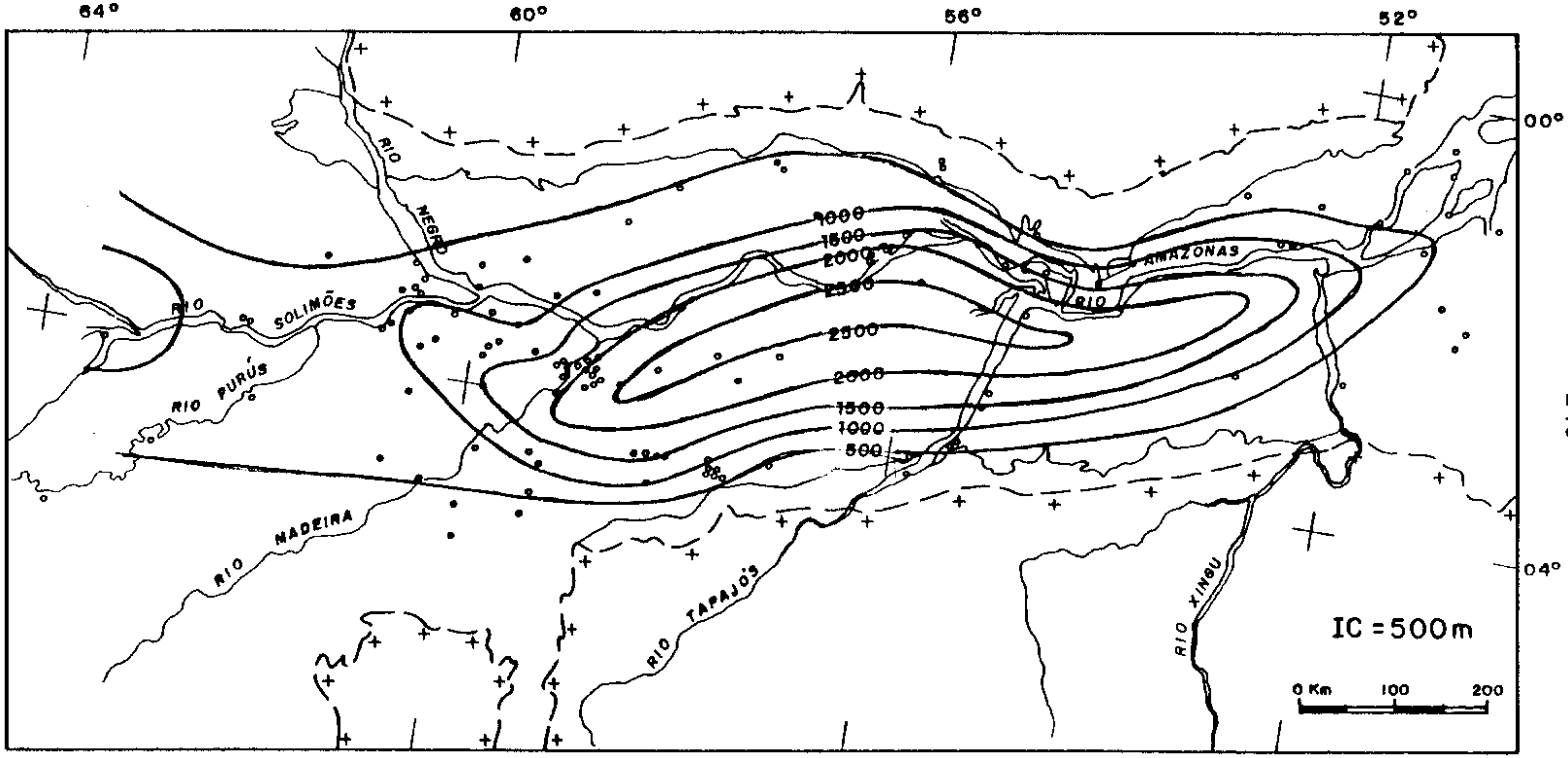


Figure 47 Isopach map of the Tapajós Group-Amazonas Basin

thick, contains red, brown to gray shale, white to cream-colored limestone and dolomite interbeds at the top of the formation. The Monte Alegre Formation truncates unconformably all underlying units with an angle of about 7 m per km, as determined in the outcrop area by Caputo and Andrade (1968) but the angle is smaller in the subsurface. In some parts of the basin, the lower contact with the sandy section of the Faro Formation is difficult to determine.

The upper contact is conformable, transitional in many places or abrupt with the Itaituba Formation. In subsurface the upper contact is normally placed below an anhydrite bed very constant across the basin. The subsurface contact may be a little displaced in relation to the contact seen in outcrop, but this difference seems insignificant.

The basal sandstone beds of the formation are generally unfossiliferous. The shale beds may contain *Lingula*, fish and plant remains and the limestone and dolomite interbeds in the central parts of the basin at places contain brachiopods, bryozoans, foraminifera and conodonts. Especially interesting is the presence of the conodont *Streptognathodus* which suggests that the unit is not older than early Late Carboniferous (Bouman and others, 1960).

The genus *Millerela* indicates an age between Late

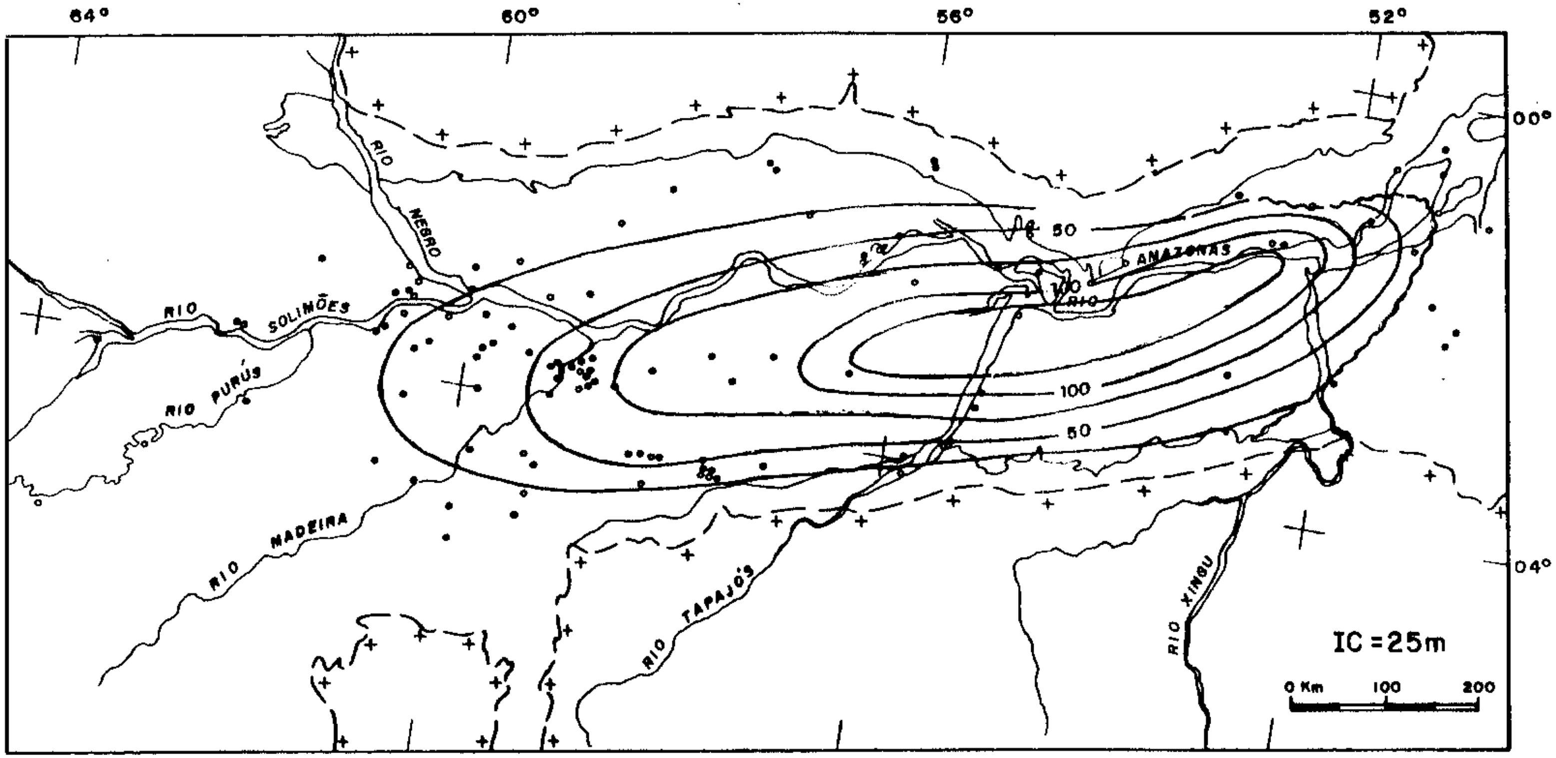


Figure 48 Isopach map of the Monte Alegre Formation - Amazonas Basin

Mississippian and Late Pennsylvanian (Daemon and Contreiras, 1971a,b). Petri (1952, 1956, 1958), on the basis of fusulinaceans, recognized a Middle Pennsylvanian age for the overlying Itaituba Formation.

According to Daemon and Contreiras (1971a,b) the presence of saccite spores which are significant from the Westphalian C/D onwards (late Middle Pennsylvanian), and the absence of disaccite spores, which are important from the Stephanian (Late Pennsylvanian) onwards, indicates that the age of the formation is Westphalian C/D. It is possible that its lower portion may have been laid down during the Westphalian C.

In most of the outcrop area the Monte Alegre Formation was deposited in a fluvio-aeolian environment as can be interpreted from large-scale cross-stratification, bimodal sand-grain distribution and fine grain-size.

On the flanks and central parts of the basin, in the upper part of the section, the presence of unbroken shells of brachiopods and entire fronds of *Fenestella* and other bryozoans is indicative of a quiet shallow-water marine environment (Bouman and others, 1960). The marine transgression which entered the basin from the west is characterized by a low energy milieu. This sea may have been very shallow and tideless. In Peru and Bolivia, the sea first occupied a back arc basin, and later it ingressed in the

Solimões Basin. From the Solimões Basin it invaded the Amazonas Basin, and at the maximum of high-stands of sea-level, flooded the Parnaíba Basin, reaching distances of more than 3,000 km from the open sea.

From Ordovician to Earliest Carboniferous times sedimentation was mainly controlled by sea level and climatic changes, framed in a weak tectonic background. In the Late Visean time, while the Marajó area (arch) was being uplifted the Coari arch was being down-warped, the drainage systems were directed towards the west, giving way to the new transgression of the sea from the west.

This tectonism was very important in eastern South America from Early Carboniferous to Mesozoic time and it may have been related to the uplift the Gurupá arch that preceded the rupture of western Gondwana.

The presence of limestone, dolomite, red beds, aeolian sandstone and correlation with evaporites in the Solimões Basin suggest a warm and dry climate in the region.

At Westphalian D time, the Amazonas Basin area may have been located in the trade wind belt of high evaporation, far distant from polar zones, as can be deduced by its rock record. At that time the northwestern edge of Gondwana ice-sheets may have reached only the Paraná and southern part of Andean basins. The climate changed radically in the area, from Late Carboniferous time on.

ITAITUBA FORMATION

The name Itaituba Series was proposed by Hartt (1874) to define Carboniferous limestone rocks which had been discovered by Coutinho (1863, in Agassiz, 1866) along the Tapajós and Cupari river banks. Later, the upper part of the section which is composed of limestone and also halite was separated from it and given the name Nova Olinda Formation. Figure 49 shows the isopach contour map of the Itaituba Formation. The Itaituba Formation is a heterogeneous cyclic unit composed of limestone, dolomite, shale, siltstone, sandstone and anhydrite beds. The total thickness of the unit ranges from 110 m at the outcrop area to 420 m in the middle of the basin.

The limestone and dolomite beds consist of calcarenites with. bioclastic or calcisiltic matrix, biocalcisiltstone, micrites,. oolitic and lithoclastic calcarenites (Carozzi and others, 1972). Anhydrite beds are white, gray, red, and mainly nodular. The. maximum composite anhydrite thickness reaches about 180 m, mainly. in the upper part of the section, and its total volume is as much. as 3,185 x 10^4 km^3. Shales are black, blue, green, brown and. variegated.

The siltstone and fine- and rarely medium-grained sandstone. beds are brown, green and, cream-colored, cross-laminated, with argillaceous matrix and calcareous or anhydritic cement.

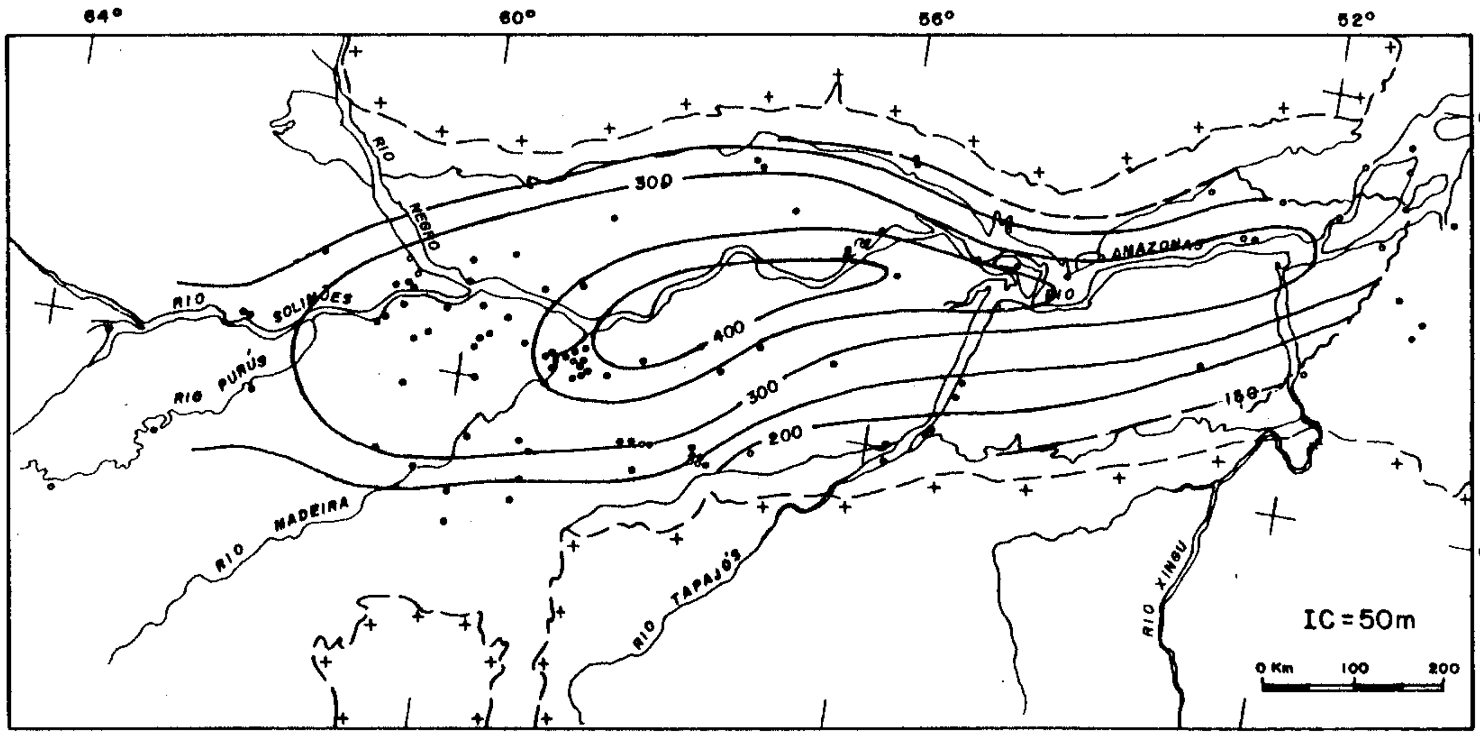


Figure 49 Isopach map of the Itaituba Formation - Amazonas Basin

The limestone beds predominate in the lower half of the unit and anhydrite beds in its upper part. The anhydrite beds are rarely exposed due to present moist conditions in the area, so that only clastic and rare carbonate sediments crop out. Slump structures are common in the unit as well as in the overlying Nova Olinda Formation due to anhydrite solution.

In the southern outcrop belt the Itaituba Formation comprises thick limestone beds, while in the northern outcrop belt clastic intercalations are common.

The Itaituba Formation overlies conformably the Monte Alegre Formation and its basal contact is placed below its lowermost anhydrite bed. In the Coari arch, in some places the Monte Alegre Formation is missing, so that the Itaituba Formation may overlie unconformably older formations, including the Precambrian Prosperança Formation. The upper contact is conformable and it is placed at the base of a clastic section of the overlying Nova Olinda Formation. On the extremities of the outcrop area, the Itaituba Formation is unconformably overlain by the Alter do Chão and Almeirim formations.

The Coari arch migrated eastward in Late Triassic to the present position of the Purus arch. (See the section of the Itaituba Formation in the Solimões Basin, figure 21.).

The Itaituba Formation limestone beds contain, by far, the richest biota fossil found in the Amazonas Basin formations. It has

Andean affinities (Mendes, 1959, 1961). The most frequently found fossils are brachiopods, gastropods, bryozoans, ostracodes, conodonts, foraminifera, trilobites, corals, cephalopods and fish remains. The most frequently found fossils in shales are the thin-shelled crustaceous *Estheriae* and plant remains which indicate fresh to brackish water (Bouman and others, 1960). The presence of the foraminifer genus *Plectogira* and conodont genera *Streptognathus* and *Ideognathodus* suggested to Bouman and others (1960) an Early Pennsylvanian age for the section. Fusulinaceans found in the basal part of the unit were used by Petri (1952) to date the formation as Middle Pennsylvanian. According to Daemon and Contreiras (1971a,b) the basal beds are considered as laid down in the Westphalian D Stage and the upper beds in the Stephanian Stage on the basis of palynological and micropaleontological data as discussed in the Monte Alegre Formation section.

The unit was placed in the biostratigraphic intervals XIII and XIV by Daemon and Contreiras (1971a,b).

The Itaituba Formation records a general transgression with many sea-level fluctuations which began during the deposition of the upper Monte Alegre Formation, and general regressions represented by the thick anhydrite beds capped by the basal continental clastics of the Nova Olinda Formation.

The Late Carboniferous sea came from the west, at a time when the east end of the basin had been uplifted during the Hercynian Orogeny. This region started to collaps in Jurassic-Cretaceous times generating the Marajó graben.

Carozzi and others (1973) and Carozzi (1979) interpreted the section as deposited mainly in intertidal and supratidal environments, and only a small part of it as deposited in a shallow subtidal environment along the basin axis. The extent of the area, of about more than 3,000 km in length from the back arc Andean belt (Tarma Formation) to the Parnaíba Basin (Piauí Formation) makes the tidal model unlikely. If the shallow water covering the supposed huge intertidal area had to be removed from the entire basin during low tides it would be improbably to evacuate the basin in a few hours. It would be impossible to achieve the required high velocities considering that the limestone beds were generally laid down in a low energy milieu.

Here the environment of deposition is envisaged simply as a result of sea-level changes and seasonal and climatic fluctuations. Each normal cycle started with lagoonal and lacustrine clastic deposition in the greater part of the basin and aeolian and fluvial clastic deposition at its basin margins. Then, the sea transgressed across wide areas of the flats depositing limestone beds. In addition, the thickness of limestone beds depends on the degree of permanence of sea water in the basin and climatic conditions. In the summer, when evaporation exceeds precapitation the water level may drop considerably generating a steady sea water inflow to the basin. At times, the degree of evaporation may have reached such high levels that it triggered the precipitation of anhydrite beds. The thicknesses of the anhydrite beds may have been a function of the degree of sulfate concentration in the water, since many limestone beds may have been replaced by anhydrite bodies (Szatmari and others, 1975).

During the deposition of the Itaituba Formation, sedimentary cycles ended with the deposition of subareal and subaqueous gypsum and anhydrite beds. The scarce deposits of halite found in the unit (only a few centimeters thick) show that there is a deficit of halite deposits, because sea water normally contains about 20 times more NaCl than $CaSO_4$ when in solution. This suggests that a denser salty water outflow took place during Itaituba deposition. Small transgressions and regressions are observed in a few meters of limestone beds (Szatmari and others, 1975) which indicates frequent sea-level fluctuations.

Clastic intercalations are thicker and more abundant in the eastern part of the Amazonas Basin as well as in the western part of the Parnaíba Basin, whereas carbonate and anhydrite beds, respectively, are dominant from the western Amazonas Basin to the Andean basins (Figure 50). Quartz-silt particles in the carbonates suggest aeolian activity rather than aqueous current activity as mentioned by Carozzi and others (1972).

During the deposition of the Itaituba Formation continental sediments tended to crowd out marine deposits in the easternmost portion of the basin because the area was being uplifted. The major facies changes of the Itaituba Formation is the change from carbonate to clastics toward the east end of the basin.

The restricted connection with the Paleopacific Ocean at the west of the continent was probably through a chain of islands since the Tarma Formation from Peru, shows considerable amounts of anhydrite intercalations (Benavides, 1968) which also suggests an environment with occasional restrictions.

The Late Carboniferous Amazonas Basin was intermittently connected with the Parnaíba Basin. The time equivalent section laid down in the Parnaíba Basin is largely continental with only few thin tongues of marine fossilifreous limestone.

The climate during deposition of the Itaituba Formation may have been arid and the area of the Amazonas Basin may have been located in the belt of trade winds of high evaporation. The edges of ice-sheets centralized in South Africa (Crowell and Frakes, 1972) were also present in the Paraná Basin, during the deposition of the Itaituba Formation evaporites.

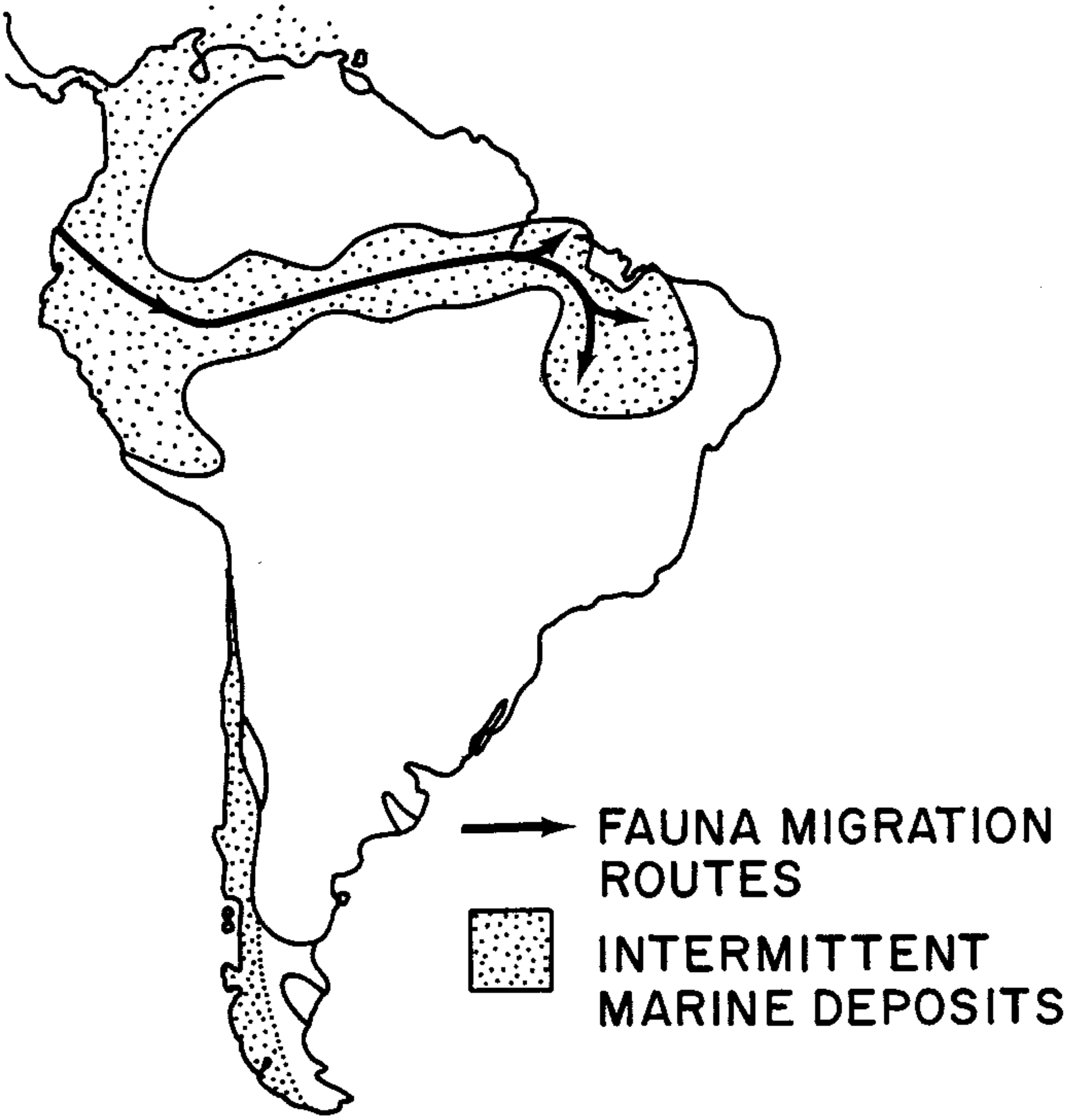


Figure 50 Paleogeographic map of South America during Westphalian "D" and Stephanian Stages (Modified from Harrington, 1962).

NOVA OLINDA FORMATION

The Nova Olinda Group was proposed by Kistler (1954) to designate an evaporite sequence occurring above the Itaituba Formation. Later the unit was considered as a formation. The lower portion of this unit occurs along the outcrop belt, but most of it is covered, occurring only in the subsurface. The total thickness of the sequence reaches about 1,650 m in the central parts of the basin. In the outcrop area the thickness measured is incomplete due to the cover of the Alter do Chão and Almeirim formations, but thicknesses of about 250 m (Urupadi and Cupari rivers) were measured by Caputo and Andrade (1968). Figure 51 shows the isopach contour map of the Nova Olinda Formation.

The unit is composed of halite beds and the same type of cyclic rocks which occur in the underlying Itaituba Formation. The halite beds range in color from pink to red, sometimes Interlaminated with anhydrite and shale interbeds. Each halite bed has a variable thickness from centimeters up to 100 m. The major concentration of halite is in the upper part of the unit. The composite total thickness of the salt beds is about 505 m in the central parts of the basin (Figure 52), making up about 30% of the total thickness of the formation. The total volurne of halite was calculated with the aid of a computer to be as much as 3.282 X 10^4 km^3.

The anhydrite beds are white to light gray, nodular, massive or interlaminated with halite or shale. The anhydrite beds show a variable thickness, but for the most part they range from 5 to 40 m, but their total composite thickness is about 280 m, making up about 17% of the total section. The total volume calculated is as much as 3.623 x 10^4 km^3. The difference in volume between anhydrite and halite beds, despite the presence of thicker halite beds, is due to the larger extent of the anhydrite strata than that of halite beds as well as facies-changes from halite to anhydrite beds toward the basin edges. The limestone beds are finer-grained and poorer in fossils than those of the underlying Itaituba Formation.

The sandstone beds are mainly fine-grained, cross-laminated, variegated, argillaceous and the medium-grained sandstone beds are cross-bedded with regular sorting.

The shale and siltstone beds are black, gray, green, brown and mainly red in the upper part of the section. In the upper third of the section potash salt occurs in two central areas of the basin with deposits comparable to those of Saskatchewan, Canada (Szatmari and others, 1975).

The Nova Olinda Formation overlies conformably the Itaituba Formation and the contact is placed below a clastic section of about 25-35 m in thickness, which overlies anhydrite or limestone

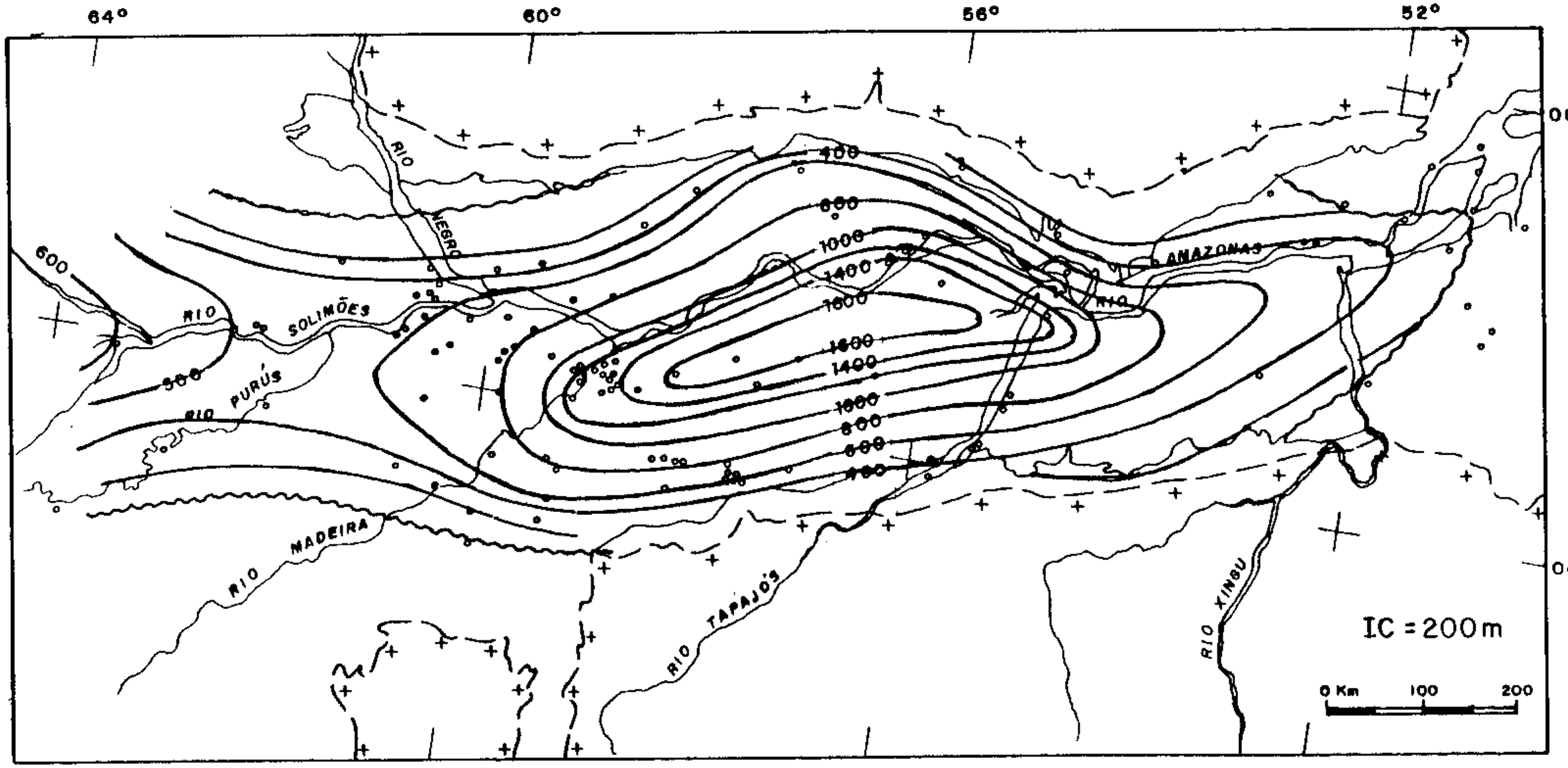


Figure 51 Isopach map of the Nova Olinda Formation–Amazonas Basin

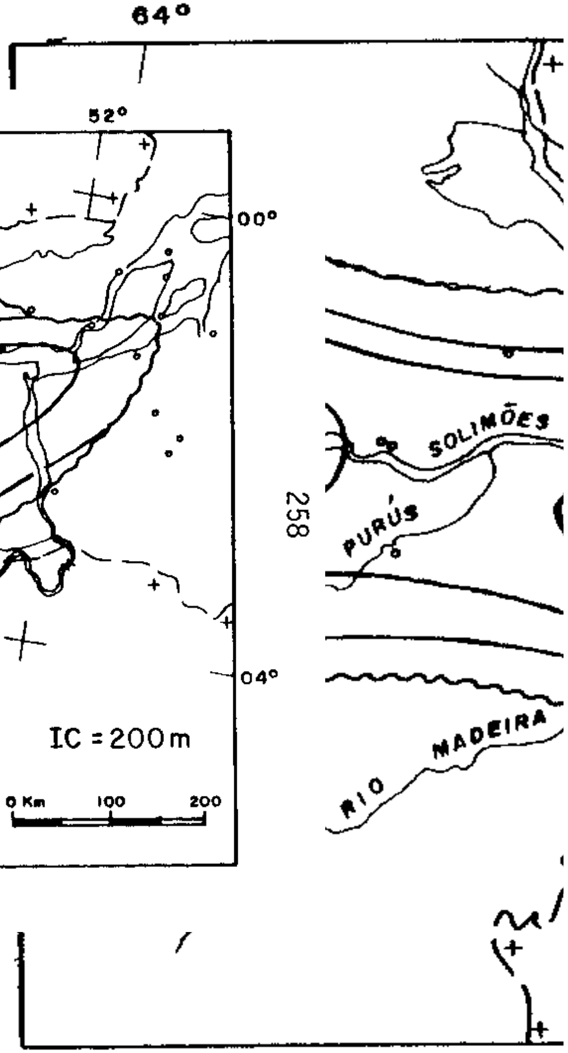


Figure 51 Isopacl

beds of the Itaituba Formation. Its upper contact is apparently conformable with the overlying Andirá Formation red beds. In the basin flanks and outcrop area this unit is unconformably overlain by the Alter do Chão and Almeirim formations.

The fossil content of the marine limestone beds consists of brachiopods, mollusks, foraminifera, conodonts, crinoids, corals, and crustaceans. Some dwarf fossils were observed in the unit. The shale horizons contain plant remains and Estheriae suggesting fresh or brackish water deposition. The age of the unit was determined as Latest Carboniferous to Middle Permian by Daemon and Contreiras (1971a,b) on the basis of palynological studies. The formation was placed in the upper part of the biostratigraphic interval XIV and in the biostratigraphic interval XV.

Most of the fossiis identified in the Nova Olinda Formation were obtained from wells, but a few fossils from outcrops were studied.

In the Nova Olinda depositional time a rather steady current of sea water must have moved eastward through the Solimões and Amazonas basins.

The Nova Olinda Formation records many evaporitic cycles with several subcycles beginning with continental beds at the base, overlain by limestone and ending with salt beds covered by continental beds of another major cycle. Several minor incomplete

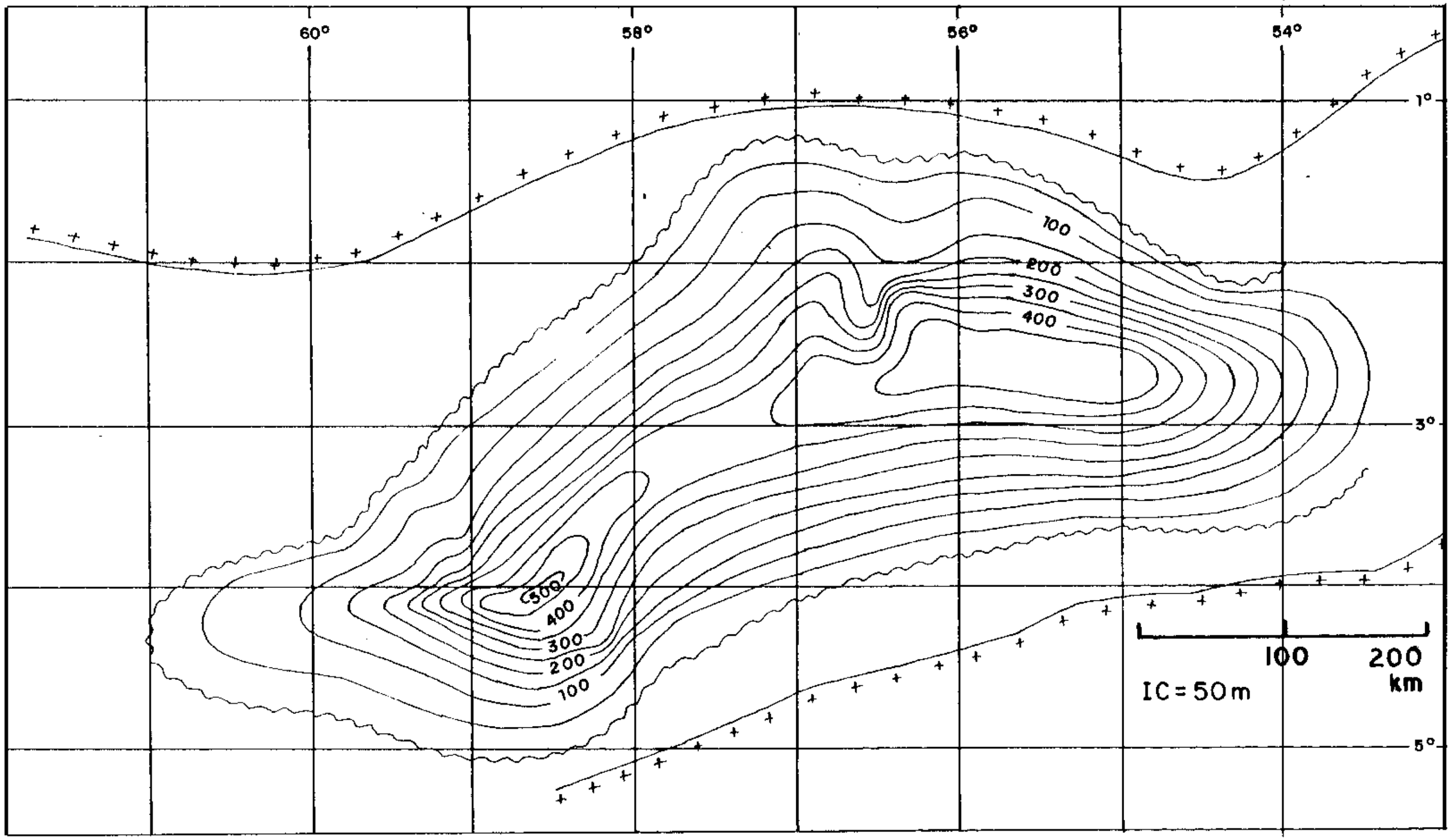


Figure 52 Isopach map of halite beds from the Nova Olinda Formation - Amazonas Basin

cycles were also recorded in the unit. The formation is characterized by development in stages, which led from marine or non-marine carbonate sedimentation to sulphate and from sulphate to chlorine deposition. The deposition of thick halite beds in the Nova Olinda Formation suggests a larger restriction than that occurring during deposition of the underlying Itaituba Formation. The large amount of halite also indicates that the outflow of denser solutions was smaller than that which took place during deposition of the Itaituba Formation. The presence of potash deposits indicates that a great volume of water may have been trapped in the basin and may have evaporated without significant dense water outflow. The sill of the basin is inferred to have become very shallow or narrow due to widespread sea-level fall.

The basin was subsiding in the western part and rising in the eastern part, and the thickest and purest salt deposits were deposited far from the rising area. Towards the Marajó highs, clastic material appears in the form of impurities and farther east in the form of clay and sandstone beds where the thickness and frequency of carbonate and anhydrite beds decrease and coarse-grained sandstone beds as well as shale and siltstone beds increase.

The halite beds were confined to the western half of the basin and clastic deposition took place with the growth of the Marajó arch. Rivers may have transported a large amount of clastic material into the eastern side of the Amazonas Basin, freshening the water and preventing chemical precipitation near the shore. The maximum anhydrite and halite thicknesses are found in the western part of the basin where broad downwarping was caused by salt loading. This widespread downwarping may have concentrated brines and may have caused a basin amplification in its width increasing the evaporation surface. While the uplift of the eastern part may have caused erosion, and a reduction in its width. This may explain the fan-like shape of the basin.

The distribution of salt and clastic deposits indicates that the basin was supplied with sea water only from its western end. The strata of highest salinity are found farthest from the areas where solutions entered the system of basins from the Andean area to the Parnaíba Basin. Equivalent evaporites of the Solimões Basin are free of halite deposits indicating a lower brine concentration than that of the western Amazonas Basin. The equivalent Copacabana Group section of Peru is almost free of anhydrite and halite deposits (Reyes, 1972), suggesting a proximity to normal marine water.

During deposition of the lowermost Itaituba Formation in the Solimões Basin, salt deposits were accumulated in it. At this time the Coari arch was the eastern boundary of the Solimões Basin and the site of highest degree of salinity was located in the eastern Solimões Basin, that is, in the part farthest from where saline solutions ingressed the basin. However, when the Coari arch was transgressed by the sea, the zone of maximum salt concentration migrated to the western Amazonas Basin where thick halite deposits accumulated in the Nova Olinda Formation.

During deposition of the Itaituba and Nova Olinda formations the northern flank received more clastics than the southern flank. Moreover, the position of the depocenter was displaced about 70 km southwards. Carozzi and others (1973) interpreted this displacement as due to transcurrent movements in a NE-SW direction. It is unlikely that during the Early Permian time transcurrent faults occurred in the Amazonas Basin because from Carboniferous to Jurassic times diabase intrusions took place in the area suggesting distension rather than shearing. A probable explanation for the depocenter displacement and larger clastic supply in the northern flank could be an epeirogenetic uplift in the Rio Branco arch (Amaral, 1974) located in the Guyana shield that gave origin to the Takutu rift.

Contrary to a rather widely held opinion that the climate of the Amazonas Pennsylvanian to Permian times must have been cold because of glaciation during these periods in south Brazil (Paraná Basin), the direct evidence of the thick-shelled, profuse fauna and extensive carbonate and halite deposition indicates a dry and warm climate throughout the northern Brazilian basins. Evaporation of water in Solimões and Amazonas basins caused a continental inflow of the less saline waters of the Paleopacific Ocean.

The Amazonas Basin during deposition of the Nova Olinda Formation was located in a trade wind belt of high evaporation. It is interesting to notice that the major halite and silvinitic accumulation took place in the Mid-Permian time when in the Paraná Basin ice caps had already evacuated the basin. In that time, only in Australia and probably in Antarctica, glaciation was active in Gondwana (Crowell and Frakes, 1971). In Early Permian times the halite deposits were less developed.

ANDIRÁ FORMATION

The name Andirá Formation was introduced by Caputo and others. (1971) to describe a section composed of red beds located at the top of the Paleozoic sequence. A thickness of 645 meters was measured in the well 1-AD-1a-AM (Andira). Its maximum thickness is estimated to be about 720 m in central parts of the basin. Figure 53 shows the isopach contour map of the Andirá Formation. The Andirá Formation comprises red, brown, light green, light gray to dark gray and black, micaceous, soft siltstone, shale and mudstone beds with intercalations of chert and limestone beds.

Subordinate sandstone beds are fine-grained, argillaceous,

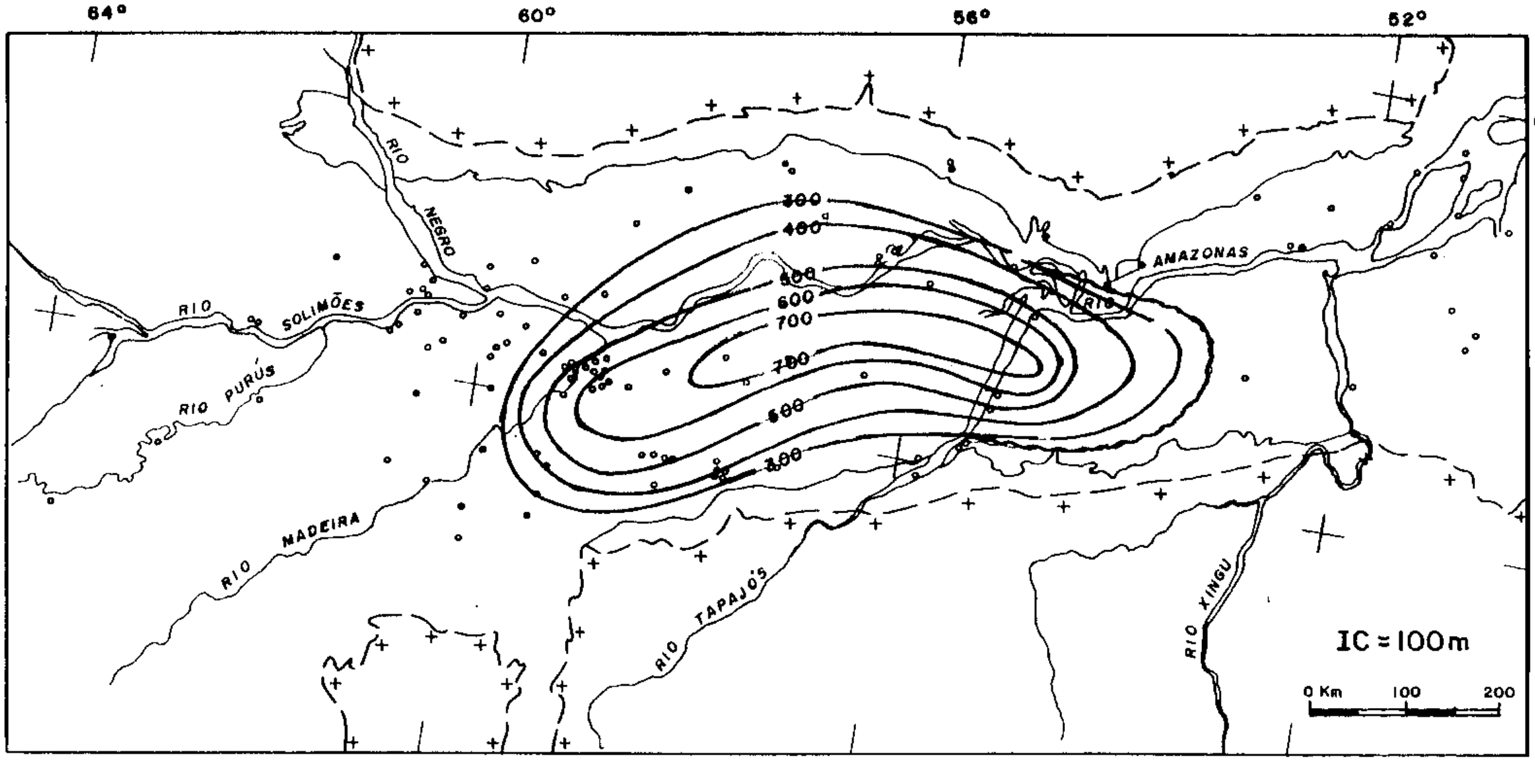


Figure 53 Isopach map of the Andirá Formation-Amazonas Basin

calcareous, laminated and micaceous with parallel- and cross-stratification. Towards the east of the basin, a change to a dominantly sandy facies is present.

In the middle of the section there are light gray, white or brown, argillaceous, pyritic limestone which grade to dolomite and chert beds and also are present two pink or black anhydrite beds, very characteristic in the section.

The Andirá Formation overlies conformably the Nova Olinda Formation and is unconformably overlain by the Alter do Chão and Almeirim Formations with a low angle of as much as 20 minutes (5m/km)).

This unit is only present in the subsurface in the flanks and central portions of the basin. It is absent in most parts of the eastern side of the basin where it was uplifted and eroded.

No index macrofossils have been found in the Andirá Formation and the biota comprises some Estheridae, fish scales and ostracodes, which suggest fresh or brackish water environment (Bouman and others, 1960).

The age of the formation was determined by Daemon and Contreiras (1971a,b), on the basis of palynological data, as Late Permian, corresponding to the biostratigraphic interval XVI. The spores are similar to those found in Late Permian sediments of the Paraná Basin. It is inferred here that sedimentation may have ended in the earliest Triassic time. The most common sporomorphs found are according to Lima (1982,· written communication): Sporomorphs: *Protohaploxypinus* sp., *Vittatina* sp., *Punctatosporites* cf. *rotundus* Bhardwaj, *Lueckisporites cf. virkkiae* (Pot. and Kl.), *Hamiapollenites*.

From Late Permian to Middle Triassic time a major regression took place on Earth resulting in a worldwide sea withdrawal from the shelves and from interior basins (Hallam, 1977).

The Andirá Formation records the transformation of an evaporitic marine to a continental basin where sedimentation took place without a marine influence. Lake and playa deposits in the central parts of the basin may have existed at the base of the formation as indicated by *Estheridae* and black shales and fresh water limestone beds. On the periphery of the basin fluvio-aeolian sediments and mud and debris flows may have predominated. Much of the edges of the formation were removed by later erosion. Loess sediments in the source areas may have been important as suggested by the great amount of siltstone beds in the section. In the upper beds, lacustrine conditions may have graded eastward into fluvial and aeolian conditions as suggested by larger grain size sediments in the eastern edge of the basin.

The Amazonas Basin began to shrank during the deposition of the Andirá Formation, so that some previously deposited Paleozoic formations, may have been eroded at its edges and redeposited in its flanks and central parts. During deposition of the Andirá Formation the eastern end of the basin was rising as indicated by the development of medium-grained sandstone beds and small thickness of the section in that direction, while the western portion was subsiding. The sea receded towards the west abandoning also the eastern Andean belt where the Andean Late Permian Mitu Group composed of red beds and halite was deposited above the Early to Mid-Permian Copacabana Group.

Szatmari and others (1975) pointed out that the Paleozoic post-evaporitic sequence of the Solimões (Fonte Boa Formation) and Amazonas basins (Andirá Formation) is a distal molassic deposit resulting from erosion of the Andean ranges during and immediately succeeding the Tardihercynian orogenic phase which lasted from 265 to 260 m.y. ago. This orogeny is important in Chile, Argentina, Bolivia and Peru where post-tectonic Late Permian molasse units overlie folded rocks with angular unconformity.

In northern Bolivia the deformations are less Intense and in northern Peru the deformation phase is not recognized and upper Permian strata rests conformably on the lower Permian sediments {Dalmayrac and others, 1980). In Colombia, in the eastern Andes, the Late Hercynian orogeny is not important, and correlative molasse deposits as old as Triassic-Jurassic were laid down far away from the Solimões Basin (Cediel, 1972). Drainage from the

Solimões and Amazonas basins was directed toward the Andean area, so it is more likely that the source areas of the Andirá Formation were not located in the Andes. Probably, source areas of the Andirá Formation were situated in the Guyana and Guaporé shields as well as in the Marajó region in Late Permian time.

The climatic conditions during deposition of the unit may have been the same as those prevailing during the accumulation of the Nova Olinda Formation. Deposition of this unit is viewed as located in the trade wind belt of high evaporation, as suggested by the presence of anhydrite, limestone, red beds and aeolian deposits. During Late Permian time, world warm and hot climatic belts may have expanded due to the shrinkage of the polar ices. In the early Late Permian (Kazanian Stage) evidence for limited glaciation exists in Tasmania (Crowell and Frakes, 1971), but there is no evidence for glaciation in Gondwana in late Late Permian time.

PENATECAUA DIABASE

The name Penatecaua diabase was proposed by Issler and others (1974) to designate tholeiitic magmatic rocks which were emplaced in the Amazonas Basin from Permian to Jurassic times. The diabase rocks are made up mainly of plagioclase, pyroxene, ilmenite and magnetite minerals displaying an ophitic texture. The Paleozoic and older sediments were intruded by diabase dikes and sills, with thicknesses that range from a few meters to more than 200 m. The composite sill thickness reaches more than 700 m. The dike thicknesses range from about 5 to 25 m in the outcrop area, but seismological surveys have shown dikes as thick as 1 km or more. Dikes which exceed 20 km in length were mapped in the southern outcrop belt by Caputo and Andrade (1968). The dikes follow N25ºE and N40ºE trends in the southern outcrop belt and a few dikes follow an E-W direction (Caputo and Andrade, 1968). In the northern outcrop belt a N25º-35ºE trend is the most common, but also N10º-30ºW, N55º-65ºW and N70º-85ºE trends were observed (Bischoff, 1963). No extrusive rocks are known related to the Penatecaua event in the Amazonas Basin.

The analysis of these rocks by radiometric determinations allowed Thomaz Filho and others (1974) on the basis of K/Ar method to establish the beginning of the magmatism in Early Permian and the end of it in the Jurassic time. They recognized at least 2 cycles of magmatic activity; the first from Early Permian to Late Triassic time and the second in the Jurassic time. No Cretaceous basic igneous activity was observed in the Amazonas Basin, although it was observed in the Parnaíba and Paraná basins.

The first magmatic cycle of the Amazonas Basin is correlated with tectonic events related to the opening of the North Atlantic

Ocean and the second is correlated with the opening of the South Atlantic Ocean (Thomaz Filho and others, 1974). The long-lasted magmatic activity may have elevated the ground surface preventing sedimentary deposition during the igneous activity when erosion took place. The drainage system may have continued towards the Andean belt area during that time and perhaps when isostatic equilibrium was achieved, subsidence began again in the Cretaceous Period.

ALTER OO CHÃO FORMATION

The name Alter do Chão Formation was first used by Kistler (1954) to define red beds that overlie the Paleozoic section of the Amazonas Basin. The term Barreiras derived from Late Tertiary sediments occurring in the eastern Brazilian coast has erroneously been applied to this formation. I favor the name Alter do Chão in order to designate the Cretaceous red beds section that are separated from Tertiary red beds by an unconformity. Figure 54 shows the isopach contour map of the Alter do Chão and Almeirim formations.

The formation, up to 600 m thick in the middle of the basin, consists of sandstone, siltstone and mudstone beds with some conglomerate interbeds. The sandstone beds are fine- to medium-grained, red, variegated, kaolinic, argillaceous, soft and generally cross-bedded with cut and fill structures. Subordinate, compact, silicified, white to cream-colored and yellow sandstone beds (Manaus Sandstone, Albuquerque, 1922; Travessões Sandstone, De Carvalho, 1926) with scattered quartz granules and pebbles are also present in the section. Mudstone and siltstone beds are soft, brick red and variegated. The conglomerates fill paleochannels in the other types of sediments of the formation. Many limonitic bands occur throughout the section, which is generally massive. At the top of the unit, in plateaus, bauxite deposits are present.

The Alter do Chão Formation partly overlies Paleozoic sediments disconformably. On the western side of the basin it is unconformably overlain by the Solimöes Formation and eastward from the Monte Alegre dome the unit is truncated by the Almeirim Formation.

The Alter do Chão Formation is relatively poor in fossils. Plant remains were found in the Paituna hill, in the Monte Alegre area and fish and dinosaur teeth and vertebrate bone fragments were found in wells drilled by Petrobras. Dicotyledonous plants indicate, at least, an age as old as Late Cretaceous. Price (1960) identified a large Theropoda tooth (Carnivorous reptile dinosaur) from the borehole 1-N0-1-AM (Nova Olinda), indicating a Late Cretaceous age for the lower part of the section.

Some core samples from wells that I gave to Daemon and

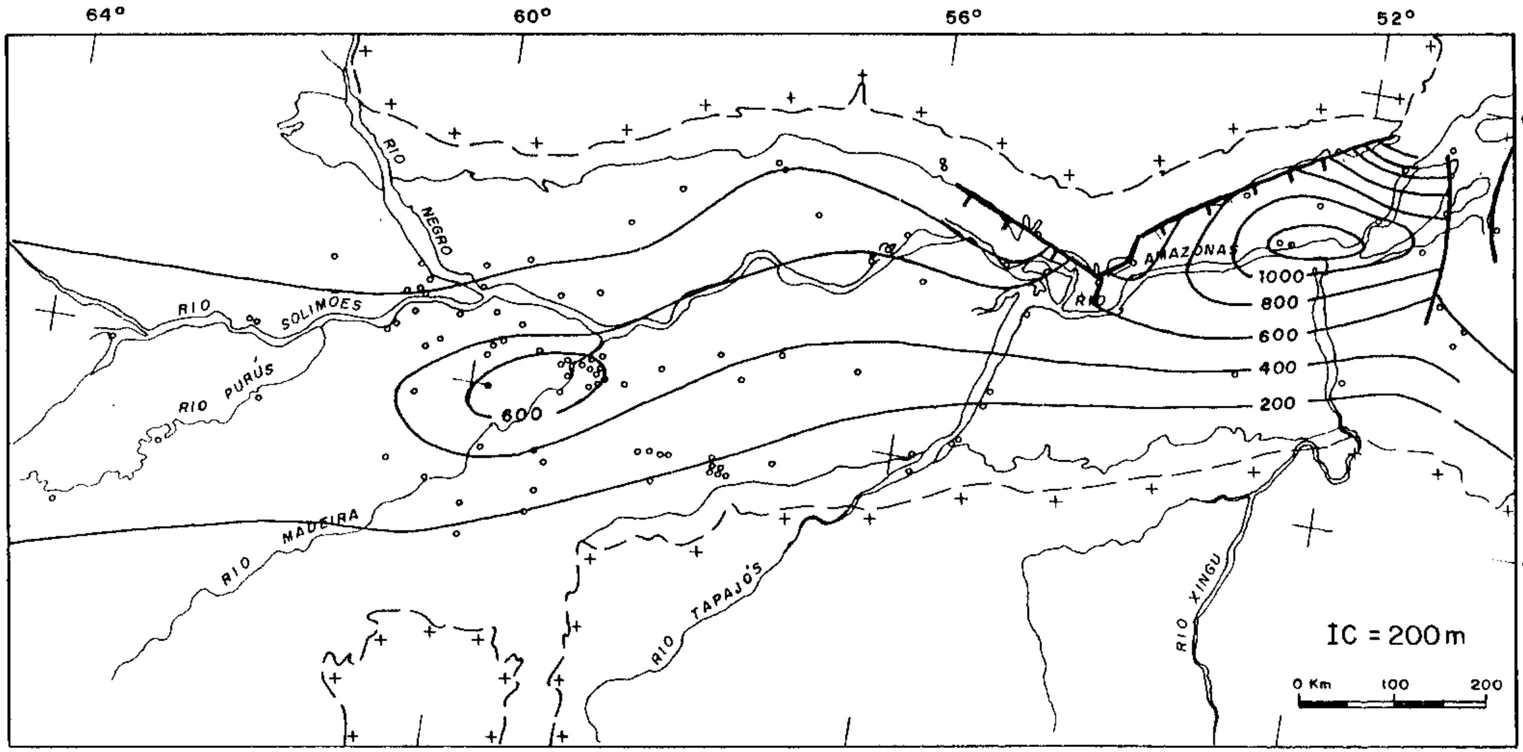


Figure 54 Isopach map of the Alter do Chão and Almeirim Formations-Amazonas Basin

Contreiras (1971a,b) furnished a Late Cretaceous age and later Daemon (1975) assigned a Middle Albian to Turonian age to the Alter do Chão Formation on the basis of palynological grounds.

The nature of the sediments, red color and continental fossils indicate that the Alter do Chão was deposited in alluvial fans and alluvial plains.

The source areas for the Alter do Chão clastics were located in the Guyana and Guaporé shields as well as the estern Marajó rift shoulder (Gurupá arch). The proximal conglomerate beds seen in the outcrop area suggest that basement was the primary source of sediments, although part of the Paleozoic section also may have furnished some detrital material to the Alter do Chão Formation. The elevated area of the Gurupá arch was under erosion while the Marajó rift was being filled with sediments. Part of the western flank of the Gurupá high is now the western shoulder of the Marajó graben and is called Gurupá arch separating the Amazonas Basin from the marginal rifted Marajó Basin. In the western Marajó basin shoulder and in part of the eastern Amazonas Basin, the Alter do Chão Formation is absent with only Tertiary sediments overlying Paleozoic and basement rocks. This suggests that the Marajó area was uplifted and the deposition of the Alter do Chão Formation took place westward from the western side of the Marajó Basin.

The drainage system continued toward the Solimões, Acre and Andean basins.

The uplift of the Gurupá arch may have blocked the main drainage system from the Parnaíba Basin, since Permian to Early Cretaceous time, so its drainage system may have been directed northtward to the Marajó Basin and Atlantic Ocean. This conclusion is contrary to Petri and Campanha's (1981) view who stressed the idea that the river system responsible for the deposition of the Alter do Chão Formation may have drained towards the coastal Barreirinha Basin, north of the Parnaíba Basin. The Amazonas river headwater was located in the Gurupá area. The Alter do Chão Formation was laid down under tropical conditions. The absence of anhydrite in the red beds may suggest humid conditions during their deposition, and the absence of coal may indicate that the enviroment was not sufficiently moist for its development. Some iron bands may indicate periods of laterite formation under seasonal, more moist conditions.

ALMEIRIM FORMATION

The Almeirim Formation is here proposed to designate a thick Tertiary section composed of red beds occurring in the eastern Amazonas Basin. The name is derived from the borehole Almeirim (2-AL-1-AM) in the eastern side of the basin where 1250 m of Tertiary sediments were drilled and the section may be completed with outcrops occurring in the Almeirim hills area. The Almeirim beds appear similar to the Alter do Chão beds, but are separated by a disconformity. The individualization of this unit is justified due to its paleotectonic and paleogeographic significance.

The formation consists of poorly consolidated sandstone, siltstone and claystone red beds. The sandstone beds are mainly coarse-grained to conglomeratic, cross-bedded, poorly sorted, red stained, kaolinitic and friable with most grains subangular. The siltstone and shale beds are generally red and brown, massive or poorly laminated and soft with some kaolinitic clay interbeds.

The Almeirim Formation overlies unconformably all older sediments and represents the last wide cover of the eastern Amazonas Basin. The boundary with the Alter do Chão Formation is not clear; some faulting may have been involved during its deposition as was observed in seismic lines made by Petrobras, but the Almeirim Formation also may onlap westward the Alter do Chão Formation. Laterally, the unit is synchronous with the Marajó Formation and the Pará Group of the Marajó Basin and in the Pará and Amapá shelves offshore. The section grades into marine sediments from the western shoulder of the Marajó Basin (Gurupá arch) to the offshore Pará and Amapá areas.

Cores from wells located above 50 m from the base of the formation yielded palynomorphs which were determined by Daemon and

Contreiras (1971a,b) to be as old as Paleocene to Eocene. Possibly in the central and eastern part of the basin the Almeirim sedimentation may have started earlier.

Sporomorphs: *Spinizonocolpites echinatus* Muller and *Bombacacidites* sp.

The Almeirim Formation represents deposition in alluvial plain environments in which the rivers may have been predominantly braided. Part of the drainage system, after the collapse of the Marajó Graben may have been directed toward the marginal rifted Marajó Basin and the source area may have been located in the Guyana and Guaporé shields as well as the western Amazonas Basin.

From Oligocene or Miocene times onwards, the Andean belt may have also contributed with large amounts of sediments due to the total reversal of the drainage system to the Atlantic Ocean.

The sustained uplift in the Marajó area which probably began in Carboniferous time is interpreted as resulting from uplift doming which preceded the rifting that opened the North Atlantic Ocean.

According to a model proposed by Kinsman (1975) an early domal uplift occurs over a deep mantle plume which ruptures and forms an extensional fault system with grabens and horsts. The crustal elevated areas are thinned due to erosion and extension and when cooled they subside below sea level forming the continental shelves and slopes. Both early features exist above sea level, as the underlying lithosphere continues cooling, the continental shelves and slope subside below sea level. The break in the continental slope may mark the boundary between continental and oceanic crust. Due to a great supply of sediment to the Mouth of Amazon River in Tertiary time, the eastern Amazonas Basin was not invaded by the sea due a peripheral bulge during subsidence as a result of the cooling of the lithosphere.

Thus, the Almeirim Formation records the subsidence that took place when the lithosphere was cooling. While the eastern side of the Amazonas Basin was subsiding the western side of the Amazonas Basin was rising, so that there is Alter do Chão Formation sediments occurring in uplifted plateaus as high as 250 m in the Tapajós area.

In Tertiary time, the Amazonas Basin enjoyed a moist tropical climate as evidenced by red beds without evaporites. However, the Tertiary Solimões Formation of the Solimões Basin contains some anhydrite crystal that suggests a drier climate towards the west (Acre Basin) prior to Pliocene times.

CHAPTER 10. PARNAÍBA BASIN

Pioneer geological surveys carried out in the Parnaíba Basin date from 1870 (Rodrigues, 1967), but systematic works started in 1913 and 1914 by Small with reconnaissance surveys. Paleontological and geological studies were carried out by Lisboa (1914), Duarte (1936), De Paiva (1937), Caster (1947a,b, 1952), Plummer (1948), Campbell and others (1948), Campbell (1949) and Albuquerque and Dequech (1950), Kegel (1951, 1953, 1956), Santos (1945, 1953), Blankennagel (1952) and Blankennagel and Kremer (1954). Regional works were conducted by Kegel (1953), Mesner and Wooldridge (1964), Bigarella and others (1965), Beurlen (1965), Barbosa and others (1966), Rodrigues (1967), Aguiar (1969, 1971), Carozzi and others (1975), Mabesoone (1977), Lima and Leite (1978), Schneider and others (1979), and Carozzt (1980).

From 1954 to 1967 many detailed stratigraphic and structural field surveys were carried out by geologists and by me working for Petrobras. Recently, geological surveys are being undertaken by public institutions (Departamento Nacional da Produção Mineral, Campania de Pesquisas de Recursos Minerais, and Universities).

Different lithostratigraphic subdivisions have been advocated by pioneer investigators, but, unfortunately, other researchers have discarded them, pointing out that some rock assembages are not extensive mapping units. However, these local facies differences are important to understand the overall basin stratigraphic framework and should be considered. Other problems are that some investigators do not take into account subsurface data when mapping, while other geologists disregard field work when studying subsurface data. The stratigraphic column of the basin is shown in Figure 26, beginning with the Serra Grande Group deposited during Ordovician-Silurian time. Figure 55 shows the Parnaíba Basin boreholes index map; figure 56 shows a cross-section along the Parnaíba Basin and figure 57 shows the basement contour map the Parnaíba Basin.

SERRA GRANDE GROUP

The term Serra Grande was proposed by Small (1914) to describe a section as much as 900 m thick composed of sandstone, conglomerate and limestone. The underlying folded limestone beds were excluded by Kegel (1953) due to the presence of an angular unconformity between the sandstone and limestone beds. The Serra Grande section is one of the most controversial units of the stratigraphic column of the Parnaíba Basin; many subdivisions are far from being unanimously accepted and different age assignments have been proposed for it. Here the unit is considered as a group and is subdivided into three formations which are from bottom upwards: Ipu, Tianguá and Jaicós Formations. All three are

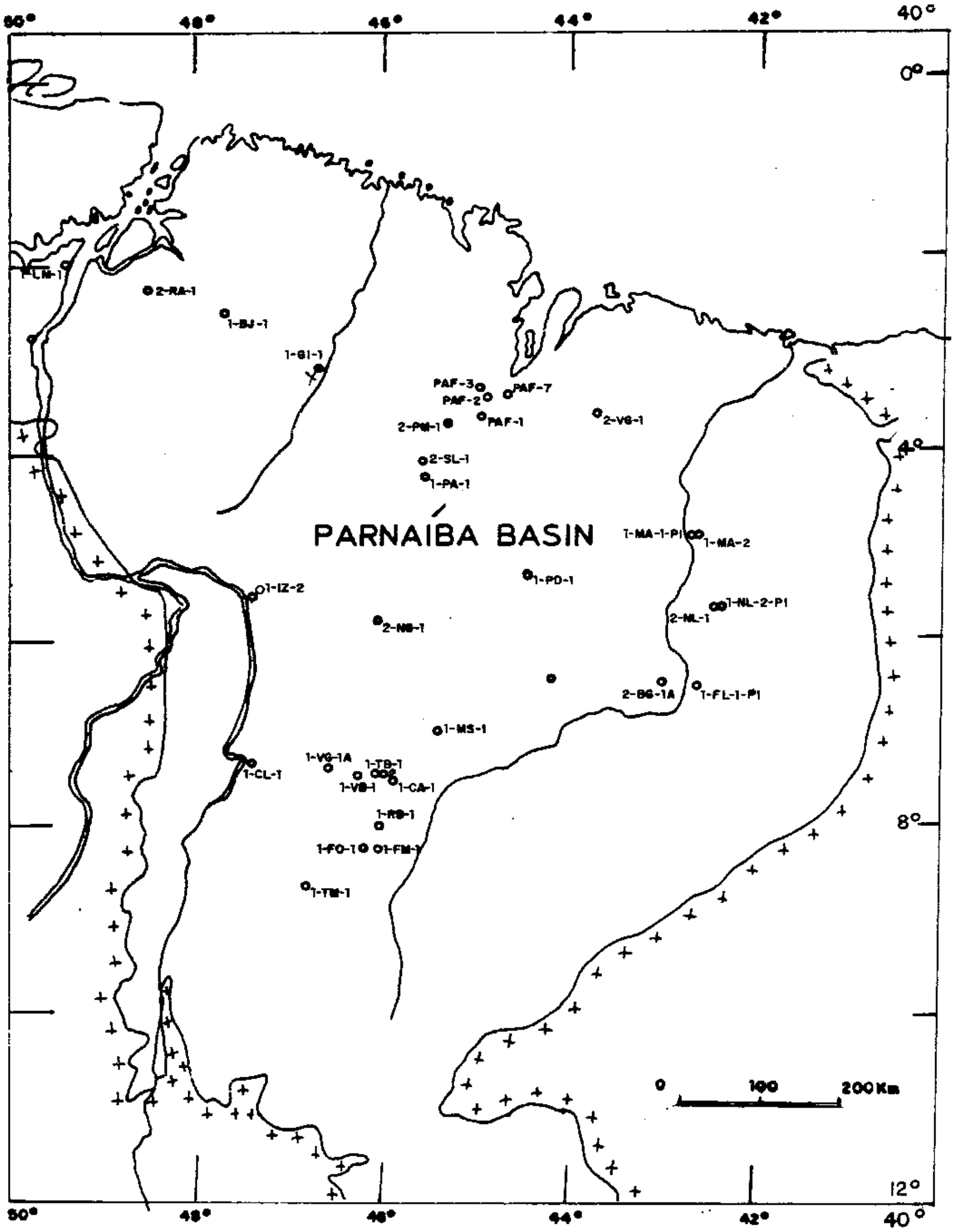


FIGURE: 55 Parnaíba Basin bore holes index map.

recognized in the northeastern outcrop belt and in the subsurface. Outside of the present basin boundary several of the Serra Grande Group remnants are isolated in small grabens and have been given different designations. They will also be considered in this study due to their great paleogeographic value. Figure 58 shows the isopach contour map of the Serra Grande Group.

IPU FORMATION

The Ipu Member was utilized by Campbell (1949) to designate the basal section along the Serra Grande scarp. This term was discarded by Kegel (1953) because it was considered to be the same as the whole Serra Grande section. Here it is used in the category of formation to define the basal section of the Serra Grande Group. Carozzi and others (1975) used the term Mirador Formation instead of Ipu Formation. However, the Mirador Formation is an older sedimentary unit of the basement. The type-section is located in the Serra Grande scarp near the town of Ipu. Figure 59 shows the isopach contour map of the Ipu Formation. I describe the Ipu section as follows. It is composed of pebbly sandstone, conglomerate and sandstone beds up to 300 m in thickness. The pebbly sandstone is massive and cross-bedded, white to cream-colored, friable to well-cemented and contains rounded scattered quartz-pebbles of variable size up to 5 cm in diameter. Conglomerate beds are recurrent in the section and are

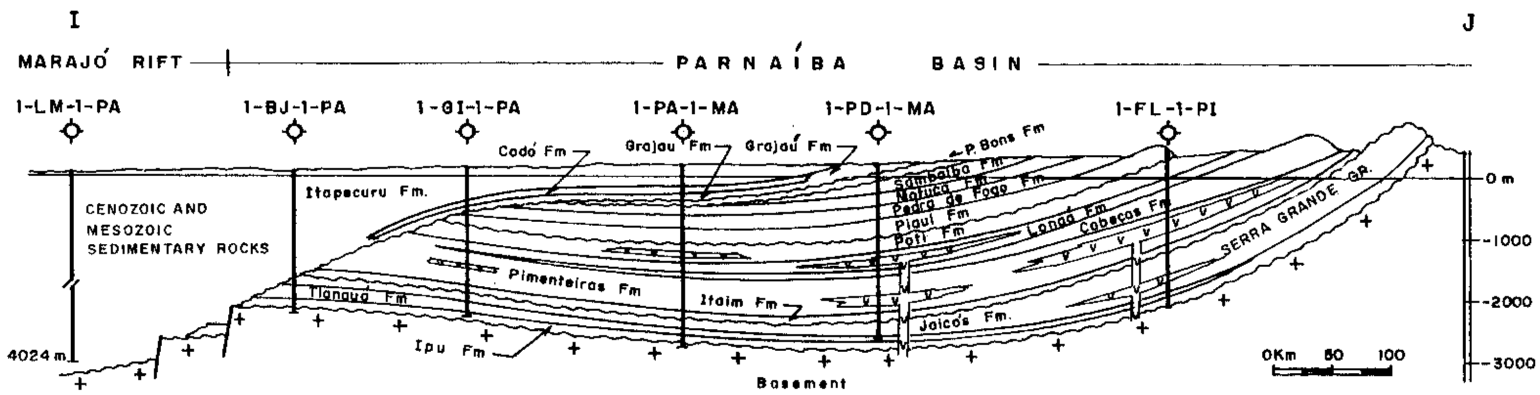


Figure-56 Longitudinal cross-section of the Parnaíba Basin composed mainly of white to light-gray quartz and quartzite boulders with a sandy and clayey matrix. At several places, rounded boulders over 20 cm in diameter are observed in the section. In the Serra Vermelha area, diamictite beds as thick as 6 meters contain quartz, quartzite, fine-grained sandstone and crystalline clasts within a massive sandy and clayey matrix, and in the Serra da Capivara area these diamictite beds are as thick as 20 m.

The sandstone beds are fine- to coarse-grained, massive to strongly cross-bedded, poorly sorted, argillaceous and micaceous, white to light gray.

White grains, considered as of kaolinitic nature, are actually agglutinated micro-quartz grains according to sedimentological analysis (Beurlen, 1965).

Kegel (1953) described some faceted pebbles from the section, but he did not observe striated pebbles in it.

In the subsurface, the coarse clastic rocks change to predominantly fine- to medium-grained, light-gray to light-green sandstone beds with some micaceous sandstone interbeds. The unit thins westward to Tocantins arch, whereas it thickens in a northeastern direction (Ipu area). On the north side of the basin gray diamictite beds occur at the top of the section in wells, as for example, the 1-CI-1-MA (Cocalinho stratigraphic well number 1) drilled by Petrobras.

The eastern outcrop belt of the basin is located 600 km away from the northeastern Brazilian coast. In the area between the coast and the Parnaíba Basin outcrop belt several remnants of the Serra Grande Group have been described by Mabesoone (1977) in downfaulted depressions as at the Mirandiba, Rio do Peixe, Cariri valley, Jatobá and Sergipe-Alagoas Basin areas.

In the Cariri valley part of the Serra Grande conglomerate described by Mabesoone (1977) is considered as a true diamictite composed of dispersed rounded and angular pebbles and cobbles derived from igneous and metamorphic rocks of the basement, supported by a sandy and clayey groundmass.

At the margins of the Jatobá Basin the basal section also contains thick conglomerate beds comprising the Tacarutu Formation which is equivalent to the Serra Grande Group. In the Sergipe-Alagoas coastal Basin, the lowermost section comprises the Atalaia and the Mulungu members of the Batinga Formation. The Mulungu Member consists of gray or red-maroon sandy diamictites with inclusions of chaotically dispersed fragments of quartz, quartzite, granite, phyllite, gneiss, micaschist and other metasedimentary rocks in a silty and sandy groundmass (Schaller, 1969; Rocha-Campos, 1981a) with some shale interbedded beds. It grades upwards and laterally into finely-stratified pebbly sandstone or tillite and rhythmite (Rocha-Campos, 1981) crowded with

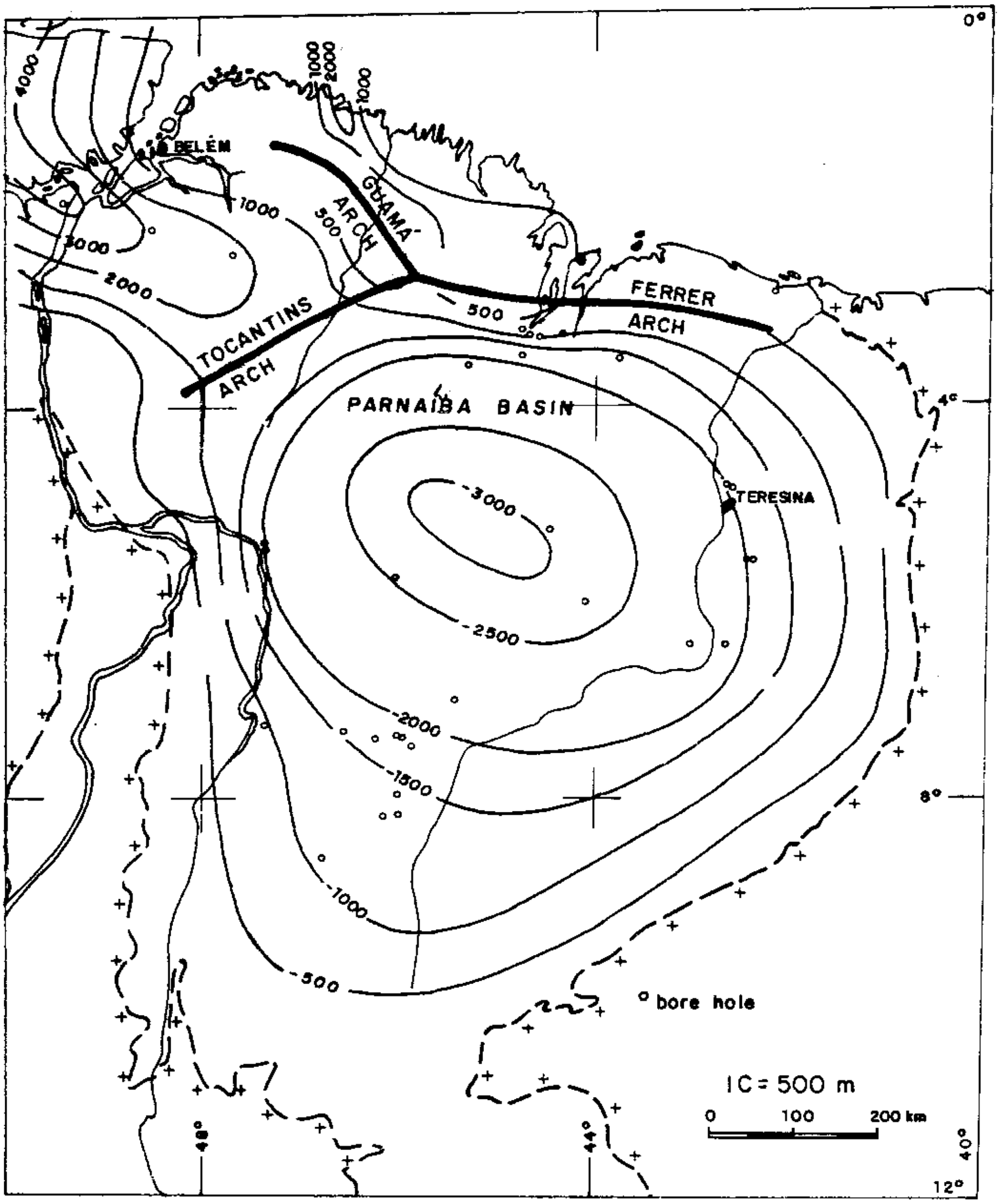


Figure 57 Structural contour map of the Precambrian basement.

centimeter-sized clasts. The tillite has abundant and unoriented clasts; some are faceted and striated, particularly those clasts of soft constitution (Rocha-Campos, 1981a).

Despite the wide extent and thickness of the Ipu and its equivalent formations outside the Parnaíba Basin, the unit shows a remarkable facies everywhere in the outcrop area, characterized by fining upwards coarse clastics, and some diamictite beds.

There is a difference in elevation in the basement of more than 300 m along the basin margins, which suggested to Kegel (1953) that the basement surface was not planar. However, I find that the basement was an almost plane surface when deposition started; the elevation differences are due to differential subsidence in the depocenter, now exposed (Figure 59). A broad uplift in the eastern and northern sides of the basin from Carboniferous to Mesozoic times, removed a large part of the section of the basin exposing its northeastern trending depositional axes (Figure 58) in the Ipu area, where the basement shows the lowest elevation and the Ipu Formation the largest thickness. The deep erosional retreat has removed a large part of the basin, so that a roughly N-S cross-section of the Serra Grande Group is displayed along the eastern basin escarpments.

In the Sergipe-Alagoas basin, close to the northeast coast of Brazil, underlying the Mulungu Formation, the basement surface is

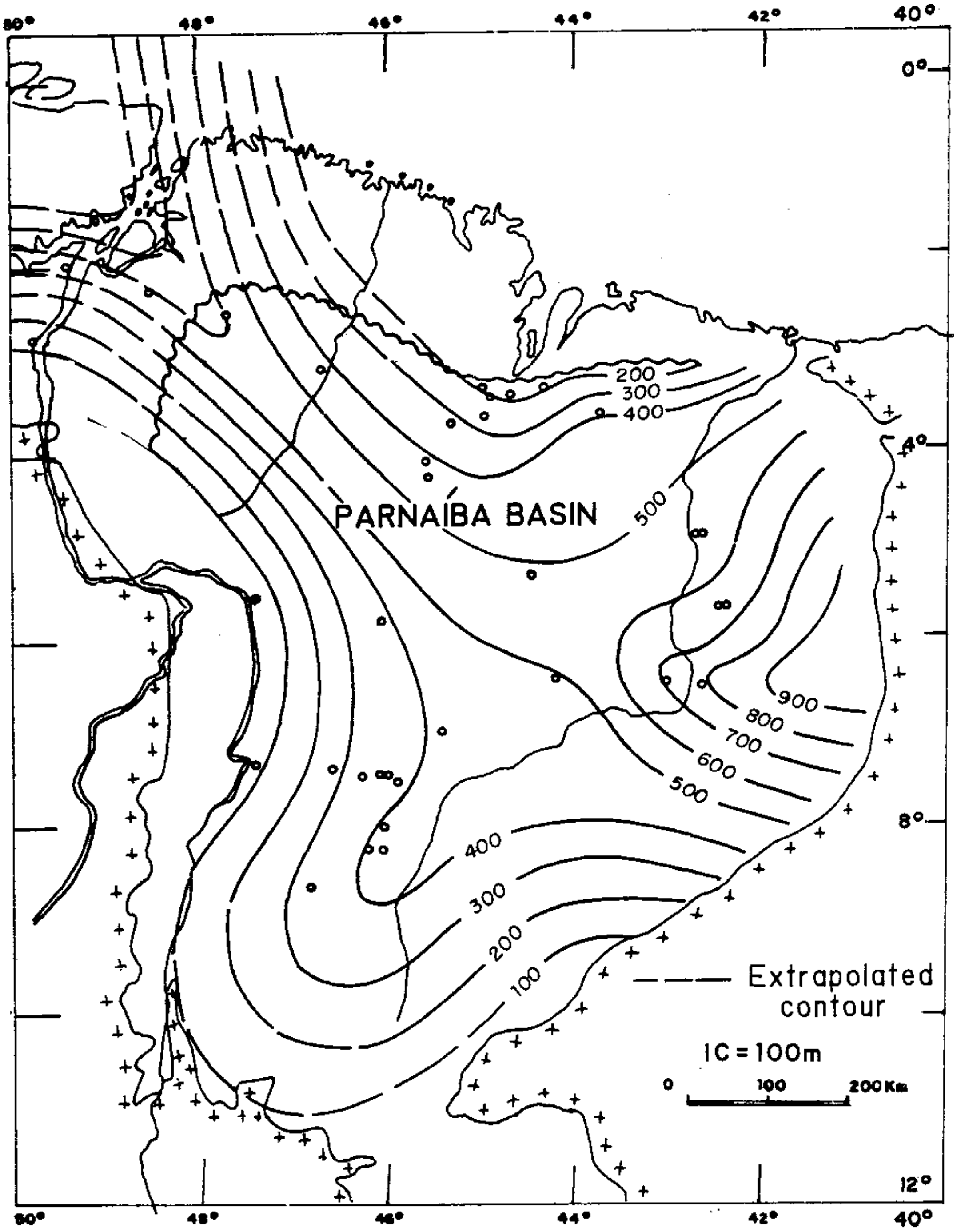


FIGURE:58 - Isopach map of the Serra Grande Group irregular and ondulating, showing parallel elongated, polished and finely striated low bosses, trending N45ºW (Rocha-Campos, 1981a).

The Ipu Formation rests unconformably on Precambrian to Early Ordovician igneous, metamorphic and sedimentary rocks. In most of the Parnaíba Basin the Ipu Formation is overlain conformably by the Tianguá Formation.

Only one fossil was found and tentatively identified as *Arthrophycus* sp. by Moore (1963), which would indicate an Ordovician-Silurian age. Age assignments for the unit were Cretaceous (Small, 1914); Carboniferous (Plummer, 1946, 1948); Early Devonian (Blankennagel, 1952; Kegel, 1953); Silurian-Devonian (Aguiar, 1971); Late Silurian (Mabesoone 1977; Mesner and Wooldridge, 1964 and Quadros, 1982); Ordovician Brito (1969) and Bigarella (1973b,c). Modern palynological studies proved that the overlying Itaím Formation is Emsian (late Early Devonian) to Eifelian (early Mid-Devonian) so the Ipu Formation cannot be younger than Emsian.

This study considers the Ipu Formation as old as Late Ordovician to Earliest Silurian. This is so because there is an unconformity between the Serra Grande Group and the Itaim Formation of Devonian age. In addition, the Ipu Formation is conformably overlain by transgressive Early Silurian shales of the Tianguá Formation. Finally, the diamictites of supposed glacial

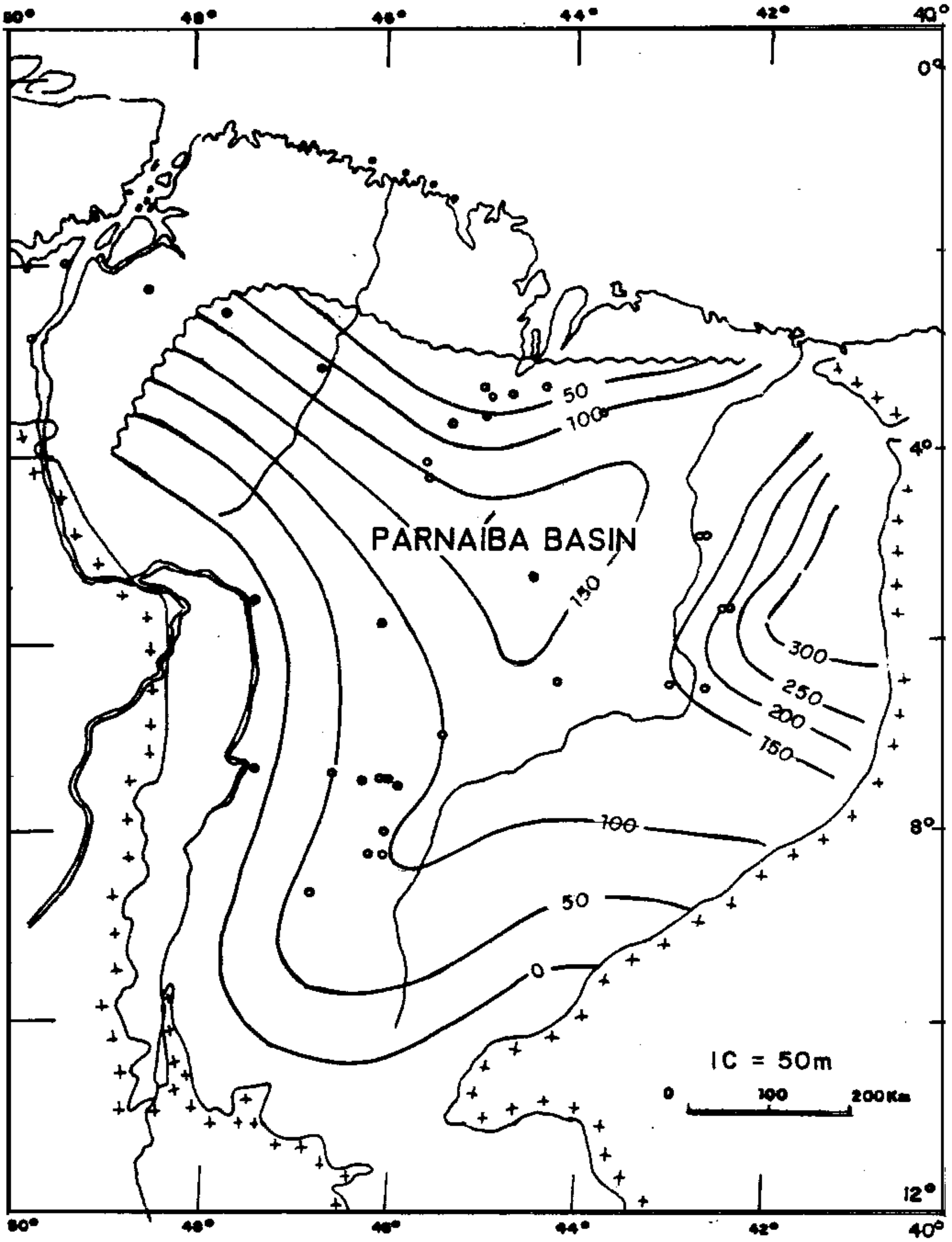


FIGURE 59 - Isopach map of the Ipu Formation

origin in top of the unit can be correlated with glacial diamictites, located in an equivalent stratigraphic position, existing in northwest Africa. This hints as well at a Late Ordovician to Early Silurian age.

In he Sergipe-Alagoas Basin, there is also controversy over the age of the Mulungu tillites. They have been considered as Late Carboniferous in age on the basis of spores (Florinites) found in the overlying shales of the Boacica Member of the Batinga Formation. However, in the adjacent Jatobá and Parnaíba basins, Late Carboniferous limestone, anhydrite, aeolian sandstone and red beds may have been deposited in low-middle latitude areas under arid conditions as suggested by its rock assemblage. These hot arid conditions are very distinct from those indicated by the cold environment Mulungu Member of the Sergipe-Alagoas Basin. An unconformity should exist between the Mulungu tillites and the overlying Batinga Formation. The correlation between the upper Ipu Formation of the Parnaíba Basin and the Mulungu Member of the Sergipe-Alagoas Basin is here proposed due to their sedimentary affinities. The Mulungu Member should be raised to the category of an independent formation whose age may be Late Ordovician to earliest Silurian (Figures 2, 94).

The Ipu Formation correlates in age with the Nhamundá Formation of the Amazonas Basin, although the general facies of both units are different in their respective outcrop areas. However, towards the Tocantins arch electric well logs of the Ipu Formation show good correlation with those of the Nhamundá Formation of the Amazonas Basin. The Ipu Formation also probably correlates with the Asemkaw and Ajua formations (described in the 13th chapter) of the Accra Basin of Ghana (Talbot, 1981) close to Volta Basin.

Kegel (1953) and Beurlen (1970) interpreted the deposits of the Ipu Formation (Lower Serra Grande Group) as being of fluvial origin. Bigarella and others (1965) regarded the unit as laid down in a nearshore environment. It is unlikely that the Ipu Formation was deposited in a single environment because its great thickness, width (more than 600 km) in the outcrop area and wide grain-size range. Carozzi and others (1975) interpreted its depositional environment as a result of sedimentation in coalescent deltas, whereas Mabesoone (1977) envisaged the Serra Grande Group (of which the Ipu Formation is the basal unit) as laid down in a submarine fan environment with relatively strong relief, at sites where escarpments and cliffs decrease in steepness. Mabesoone (1978) also viewed the Serra Grande as formed in a marine shallow water milieu in the outcrop area. Schneider and others (1979) regarded the basal Serra Grande clastics as deposited in alluvial and delta fan environments.

Taking into account the broad areal extent, the cratonic setting of the basin, the abundant cross-bedding, the scarcity of the clay fraction and lack of graded beds it is unlikely that the Ipu Formation was laid down in a submarine-fan environment according to Mabesoone (1977).

The lithological features of the Ipu Formation from the northeast Brazilian coast down to the basin suggest a wide environmental range from proximal glacial and glaciofluvial (outwash or sandur deposits) in the outcrop eastern flank and delta fan and delta fan front environments in the central parts of the Parnaíba Basin. The volume of sediments laid down in the eastern side of the basin and the ice load was so large that it induced strong subsidence in the present eastern outcrop area of the Parnaíba basin. The thickness of the Ipu Formation decreases toward the central part of the basin as observed in wells drilled in the area. In the western part of the basin the diamictite pinches out before reaching the outcrop area on the other side of the basin.

The formation also includes deposits possibly originating from the wearing down of highlands in Africa. In a Gondwanan reconstruction, South America has to be rotated 45º counterclockwise in order to fit Africa, so that any common trend to both continents found in South America is rotated 45 in relation to Africa. The ice striations trending N45ºW (Rocha-Campos, 1981a) in the Sergipe-Alagoas Basin correspond to supposed Late Ordovician-Early

Silurian ice striations trending N90W in Gabon (Micholet and others, 1971), Africa. The striation trend in the Sergipe-Alagoas basin is in good agreement with the well established glacial center in a region south of Sahara.

Ice striations are trending outward from a glacial center in a northern direction in the Sahara region and in a northwest direction in South America. The general westward trending paleocurrents (Bigarella and others, 1965) also suggest a source of sediments to the Parnaíba Basin from the southeast (east in relation to Africa).

TIANGUÁ FORMATION

The Tianguá Member of the Serra Grande Formation was proposed by Rodrigues (1967) to designate a section composed o gray shale and fine-grained sandstone beds occurring in the subsurface and in the northeastern outcrop belt. Carozzi and others (1975) raised it to the category of formation. Figure 60 shows the isopach map of the Tianguá Formation.

The Tianguá Formation was subdivided into three members from bottom to top which consist of (1) dark-grey, bioturbated, mica-ceous, sideritic, and carbonaceous mudstone beds with dark-gray shale and siltstone intercalations followed by (2) light-gray, feldspathic fine- to medium-grained sandstone beds and

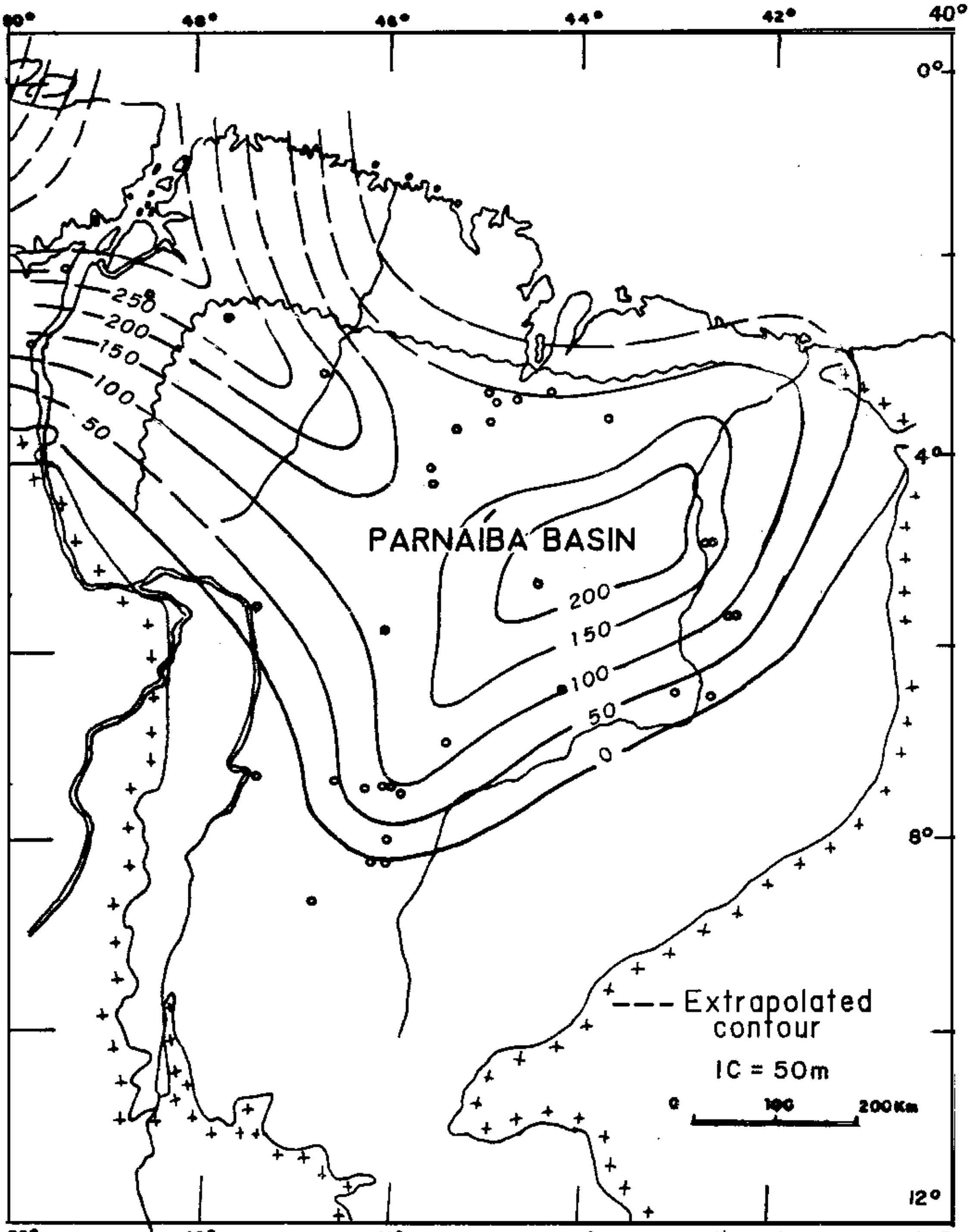


FIGURE 60 - Isopach map of the Tianguá Formation

(3) dark-gray to greenish dark-gray, micaceous, sideritic, bioturbated shale with sand grain inclusions (glaciomarine ?) and siltstone intercalations.

The formation overlies conformably the Ipu Formation and it is in turn conformably overlain by the Jaicós Formation. It crops out only in a small area from Tianguá to Ipueiras towns in the northeastern outcrop belt, while it changes its facies to sandstone towards the eastern and southern outcrop belts.

In the Tianguá area, the Tianguá Formation occurs 91 m above the base of the group and is about 22 m thick, consisting of interlayered siltstone and very fine sandstone. In the Ipu area, a covered interval occurring 233 m above the base of the Serra Grande Group is considered to be the Tianguá Formation with a probable thickness of about 25 m. In the subsurface its thickness increases up to 270 m toward a northwest direction. The unit is truncated northward towards the coast, where the shales do not show marginal facies as observed in Figure 61, suggesting a continuity toward Ghana.

Apart from track and trail fossils, no macrofossils were found. Mesner and Wooldridge (1964) assigned a Late Silurian age to their upper Serra Grande Member which corresponds here to the Tianguá Formation based on palynological studies carried out by Müller (1962). New studies carried out by Carozzi and others (1975) considered the Tianguá Formation as Early Silurian to Early

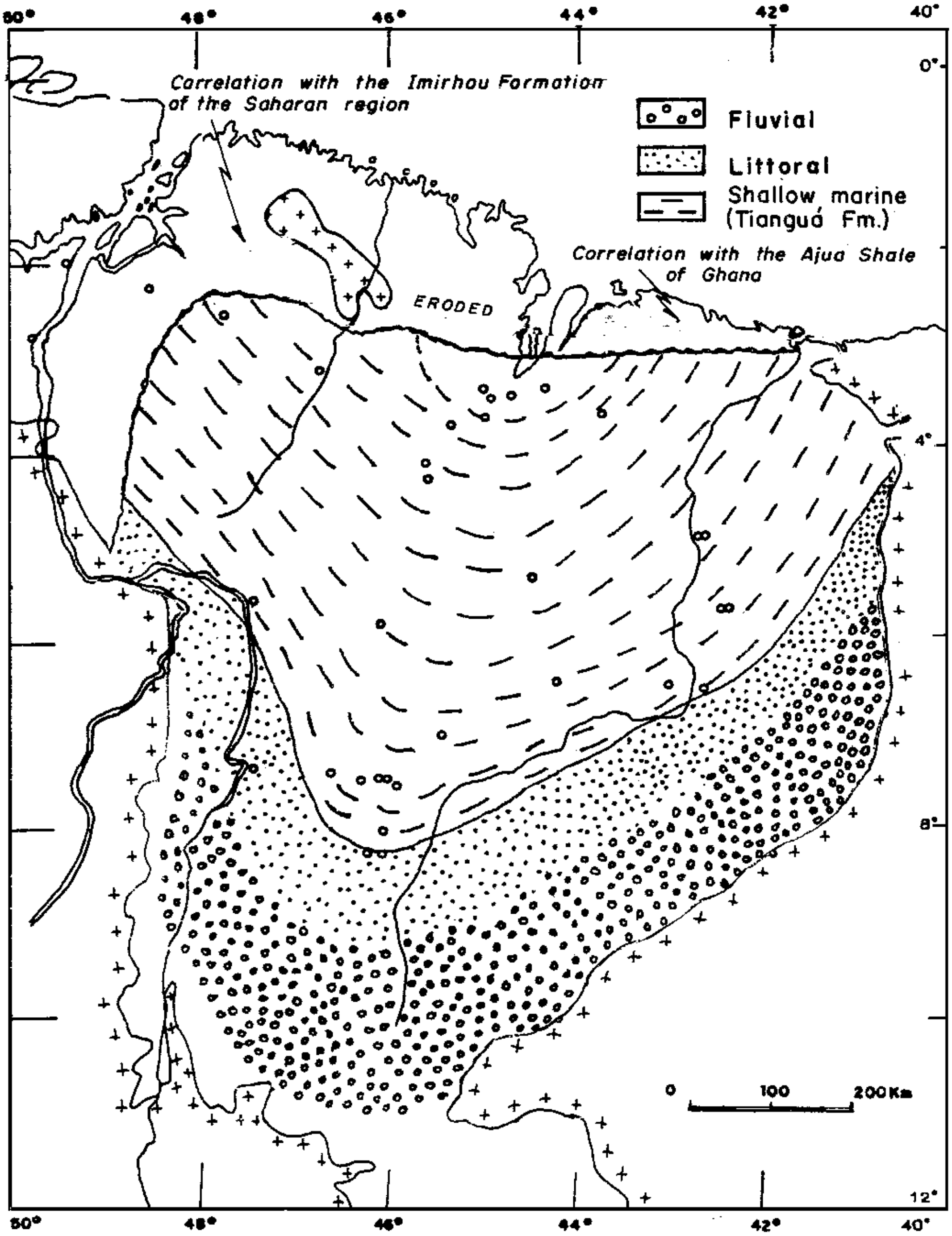


FIGURE 61 - Distribution of the Tianguá Formation (marine shale and siltstone beds) in the Parnaíba Basin based on bore holes.

Devonian (Early Emsian). They pointed out that no sedimentary break had been observed corresponding to the Late Silurian because of active sedimentation process and they stressed that the apparent chronostratigraphic gap in the Amazonas Basin was probably due to lack of cores, or to inadequate preservation of pollen since in both basins lithostratigraphic contacts are clearly gradational.

I do not agree with this explanation. An in-depth palynological study was carried out by Daemon and Contreiras (1971a,b) in order to determine the stratigraphical relationship between the Trombetas Group and the Maecuru Formation. No macrofossils or microfossils indicate a Late Silurian (Ludlovian) or Early Devonian (Gedinian and Siegenian) stages. The only Early Devonian stage represented is the Late Emsian. The unconformable contact between the Trombetas Group and the Maecuru Formation is well marked on the basis of palynological and paleontological data in the Amazonas Basin.

In the chronostratigraphic and lithostratigraphic column of the Parnaíba Basin presented by Carozzi and others (1975), Wenlockian, Ludlovian, Gedinian and Siegenian stages (from Mid Silurian to mid Early Devonian) are missing. This clearly suggests a gap in the sedimentation not explained by Carozzi and others (1975). The Tianguá Formation yielded similar palynomorphs

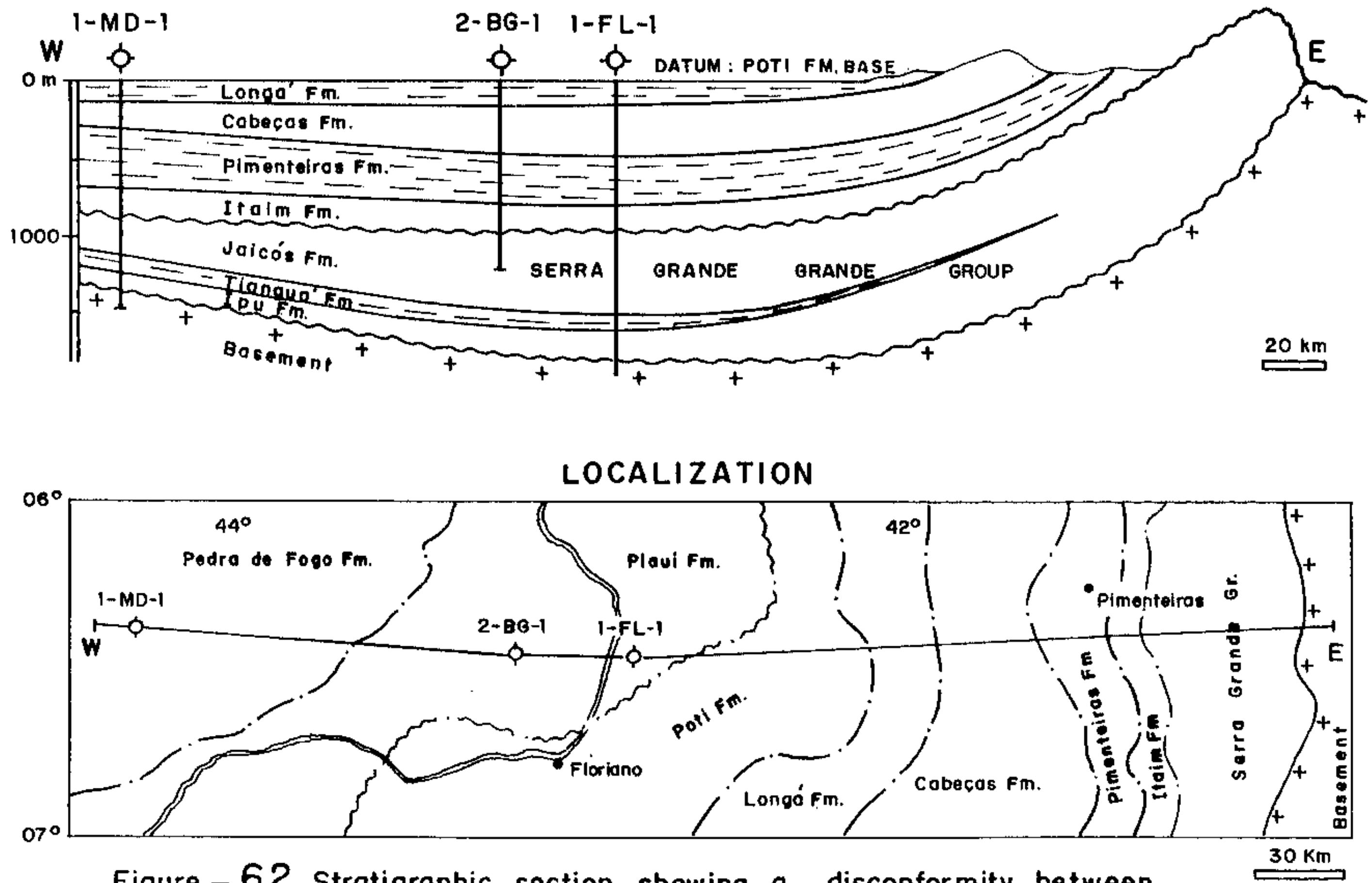


Figure – 62 Stratigraphic section showing a disconformity between Ordovician-Silurian beds (Serra Grande Group) and Early Devonian beds (Itaím Formation of the Canindé' Group) and the pinch out of the Tianguá Formation (Early to Mid-Llandoverian). Units jounger than the Longá Formation not represented in the cross-section. Parnaíba Basin.

to those of the Amazonas Basin biostratigraphic interval III of Daemon and Contreiras (1971a,b) according to Carozzi and others (1975), corresponding to the Early to the Midle Llandoverian stages (Early Silurian). Its palynological content is very similar to that of the Pitinga Formation of the Amazonas Basin. So, this suggests a same age assignment for both formations.

The Tianguá Formation is considered to have been deposited during the maximum worldwise transgression which occurred after the melting of most of the Late Ordovician-Earliest Silurian African and South American ice-caps, during late Early to Middle Llandoverian stages (Figure 61). It onlaps the Ipu Formation in the western side of the basin. In these conditions the Tianguá Formation is corre- lated with the Pitinga Formation of the Amazonas Basin and both units record the same transgressive event. The Tianguá Formation may also correlate with the Ajua Formation of Ghana, Accra Basin (Crow, 1952) and it pinches out toward most of the outcrop area (Figure 62).

The lower member of the Tianguá Formation is a result of fine clastic deposition in an offshore, pro-fan and lower shoreface milieu. The lateral equivalent environment may be fan delta front and fan delta recorded by fine-grained parallel laminated and coarse-grained sandstone beds respectively of the Ipu Formation.

The middle member may represent a fan delta front progradation resulting from a minor regression. The upper member shale probably records a deepening in the depositional environment. Ice retreat and advance may have played an important role in the transgressive and regressive units of the Tianguá Formation. Part of the Ipu Formation and part of the overlying Jaicós Formation represent lateral and marginal facies of the Tianguá Formation in most of the outcrop area, where shale beds are replaced by sandstone beds. Some thin sandstone interbeds in the Tianguá Formation may represent storm-derived deposits.

The presence of a large number of bioturbations in the shales indicates a low rate of deposition. The climate may have changed from glacial arctic to subarctic. The absence of limestones and other warm climatic indicators also corroborate the predominance of a very cold climate.

JAICÓS FORMATION

The name Jaicós Formation was proposed by Plummer (1946, 1948) to designate sandstone and conglomerate beds occurring in the Serra Grande escarpments, but because the designation Serra Grande (Small, 1914) had priority the term Jaicós became useless. Rodrigues (1967) revalidated it to designate the same section of Plummer, but Carozzi and others (1975) redefined the unit to

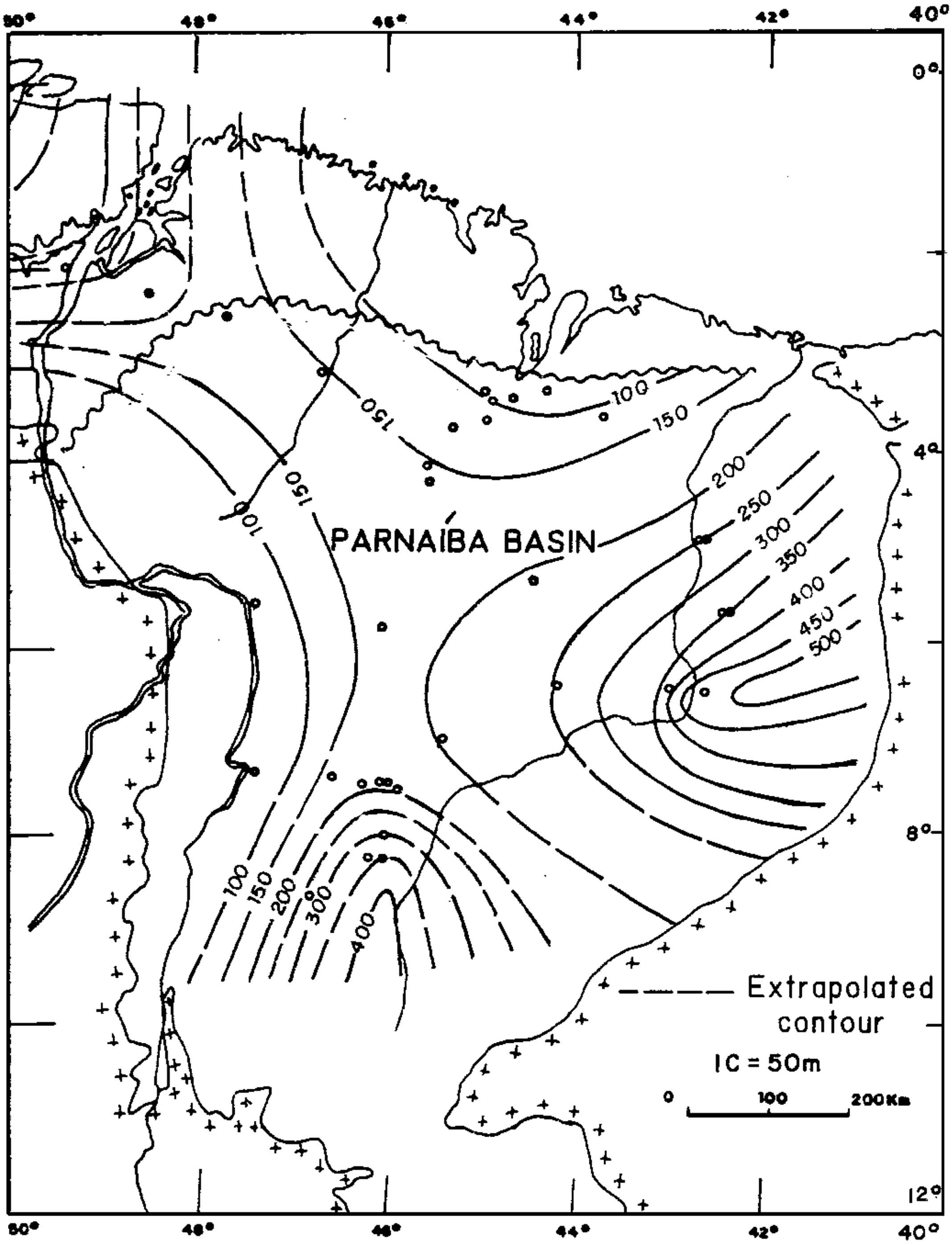


FIGURE 63- Isopach map of the Jaicós Formation

describe the section above the Tianguá Formation. Here Carozzi´s definition is maintained, although the position of its upper boundary is controversial. This section corresponds to the lower part of the Itaím Member of the Pimenteira Formation as considered by Mesner and Wooldridge (1964).

Its maximum thickness is estimated to be over 500 m in the eastern border of the Parnaíba Basin. Figure 63 shows the isopach contour map of the Jaicós Formation.

The unit is made up of gray, light-gray, cream-colored, brown and buff, angular to subangular coarse-grained and gravely sandstone with dispersed pebbles from 2 to 6 cm in diameter. The sandstone is massive or cross-bedded and lenticular, poorly sorted, and friable. Basal cross-bedded conglomerates are common, and the unit is mineralogically and texturally immature.

The Jaicós Formation rests conformably on the Tianguá Formation and is disconformably overlain by the Devonian Itaím or Pimenteira formations or by the Cretaceous Urucuia Formation. The disconformity between the Jaicós and Itaím formations is very difficult to identify, because both units are sandy, so many geologists do not recognize it.

Kegel (1953) indicated one place where the contact could be disconformable based on field criteria. The unconformity is better seen between the Pimenteira shales and Jaicós Formation when the Itaím Formation is missing.

No macrofossils were detected in the Jaicós Formation, but some sporomorphs determined in shale interbeds found in the sub-surface indicate a Silurian age for the unit (Müller, 1962, 1964). Carozzi and others (1975) considered the Jaicós Formation as deposited in Early and Mid-Emsian times, but overlying conformably the Llandoverian Tianguá Formation. Obviously, this stratigraphic relationship should be incorrect because a large time gap exists between Llandoverian and Emsian stages.

Here, on the basis of its sporomorphs, regressive character and correlation with Amazonian (Manacapuru Formation) and North African Atafaitafa Formation (Bennacef and others, 1971) of the Algerian Sahara, a Late Llandoverian to Earliest Wenlockian (late Early Silurian to early Mid-Silurian) age is indicated for the Jaicós Formation. This is in agreement with a worldwide regression which took place from Late Llandoverian to late Early Devonian. During this time shorelines migrated northward more than 1,000 km in Africa from southern Algeria to northern Morocco {Berry and Boucot, 1973), when the sea retreated from the Amazonas and Parnaíba basins.

For some time a Late Devonian diamictite section named the Cabeças Formation was confused with the Jatcós Formation, so that Devonian diamictites of the top of the Cabeças Formation were regarded as located at the top of the Jaicós Formation (Serra

Grande Group) (Blankennagel and Kremer, 1954; Malzahn, 1957; Bigarella, 1973). Actually, the Serra Grande diamictites are located at the top of Ipu Formation. In the outcrop area, the Jaicós Formation was considered as deposited in a fluvial (Kegel, 1953, Beurlen, 1965), shallow marine, (Bigarella and others, 1965; Bigarella 1973a,b,c, Mabesoone, 1978), submarine fan environment (Mabesoone, 1977). In a regional study it was considered as deposited in upper deltaic distributary, lower deltaic distributary and delta front by Carozzi and others (1975), or alluvial fan, delta fan and delta fan front by Schneider and others (1980). Here the sedimentation of the Jaicós Formation in the outcrop area is also considered as a result of deposition in alluvial and delta fan areas and also in delta fan and delta fan front in the subsurface. Some delta fan front deposits may occur in sections of the outcrop area, resulting from fluctuations in the sea-level. Part of the pro-fan and offshore deposits are represented by the Tianguá shales and siltstones. Some periglacial influence in the sedimentation is also inferred because ice-sheets were present in South America in the Andean area (Crowell, 1980, 1981) during Late Llandoverian time).

The general absence of warm climate indicators suggests a cold climate in the area during the deposition of the Jaicós Formation. Mabesoone (1975) also interpreted that the deposition of the Serra Grande Group occurred under cold conditions. The presence of abundant mica in the Tianguá lateral shales also suggests a poor chemical weathering regime in the source area compatible with a very cold climate.

CANINDE GROUP

The Canindé Group includes the Pimenteira, Cabeças and Longá formations (Rodrigues, 1967). The Itaím section was placed in the Serra Grande Group by Rodrigues (1967), but because it is more related to the Devonian section than to the Silurian it is here included in the Caninde Group.

Its maximum thickness is estimated to be over 1,000 m in the eastern side of the basin. Figure 64 shows the isopach contour map of the Canindé Group.

ITAÍM FORMATION

The Itaím Member of the Pimenteira Formation was proposed by Kegel (1953) to describe Devonian micaceous sandstone and silty sandstone beds occurring in the eastern side of the basin.

Blankennagel and Kremer (1954) did not recognize the Itaím Member, mapping it together with the Serra Grande Group. Mesner and Wooldridge (1964) considered as Itaím the section here considered as Jaicós and the section mapped as Itaím by Kegel (1953).

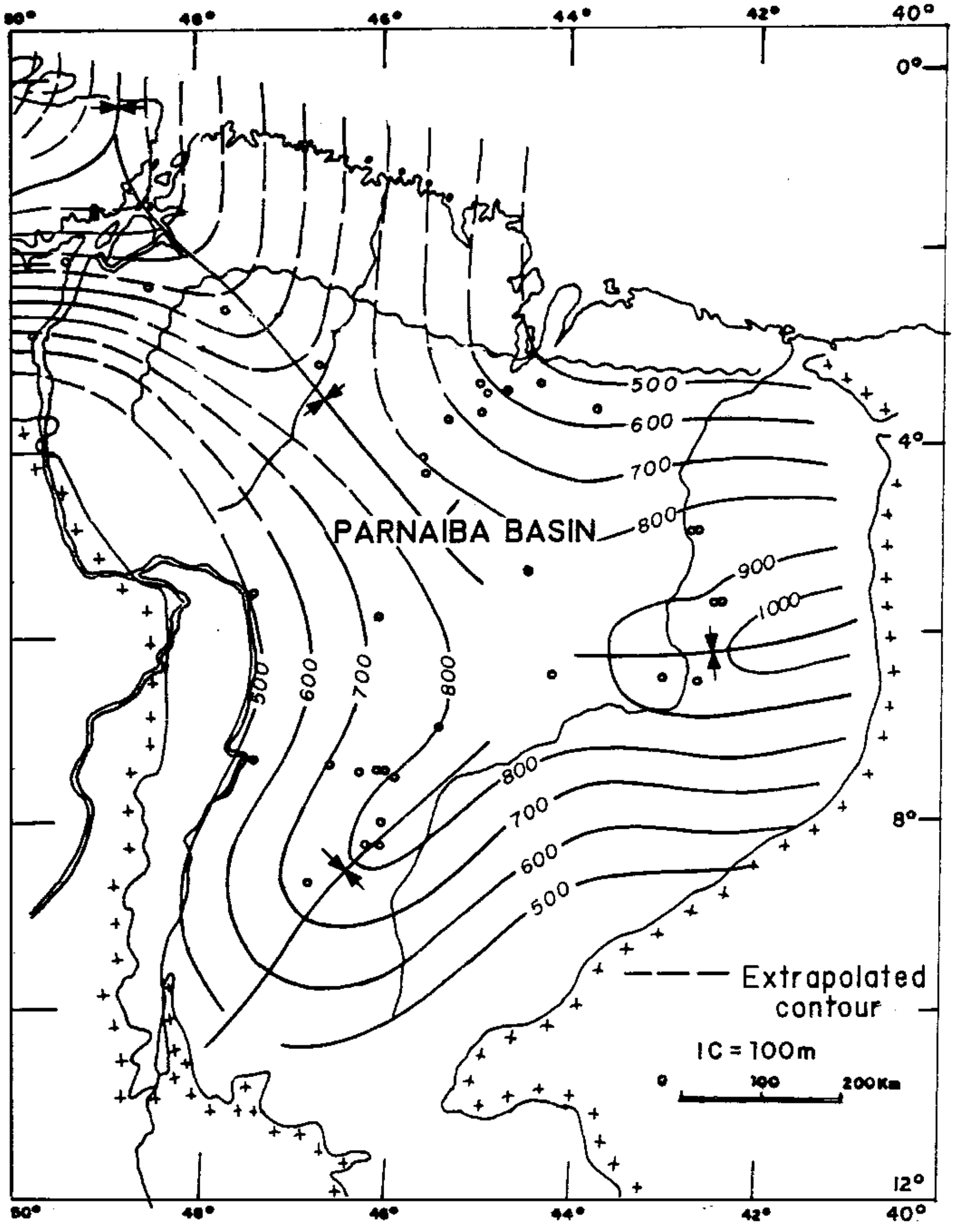


FIGURE 64 - Isopach map of the Canindé Group

Rodrigues (1967), followed by Carozzi and others (1975), considered the Itaím Formation as belonging to the upper part of the Serra Grande Group.

Here the Itaím section is mapped as an independent formation, belonging to the Canindé Group, because it is more related to the Devonian section than to the Silurian Serra Grande Group. Figure 65 shows the isopach map of the Itaím Formation.

The Itaím Formation, up to 250 m in thickness, consists of friable, moderately cross-bedded, cross-laminated and parallel bedded fine- to coarse-grained yellow, light gray, brown, purple or red, micaceous, sandstone beds with red, brown or purple shale and siltstone interbeds. Most of the oxidation colors are secondary due to the weathering of sideritic shales and nodules. Upwards, the degree of sorting, roundness, finer grained sandstone, number of shale beds and thicknesses increase, but at the top of the unit a well-sorted medium-grained sandstone body, almost free of shale, is present.

In the subsurface, the unit consists of sandstone beds and bioturbated shale and siltstone beds with fine- to medium-grained sandstone interbeds. The sandy intercalations increase toward the southeastern and eastern margins of the basin. Gray silty shale beds are frequent in the middle of the section. At the upper part of the section sandstone interbeds increase again as I observed in bore

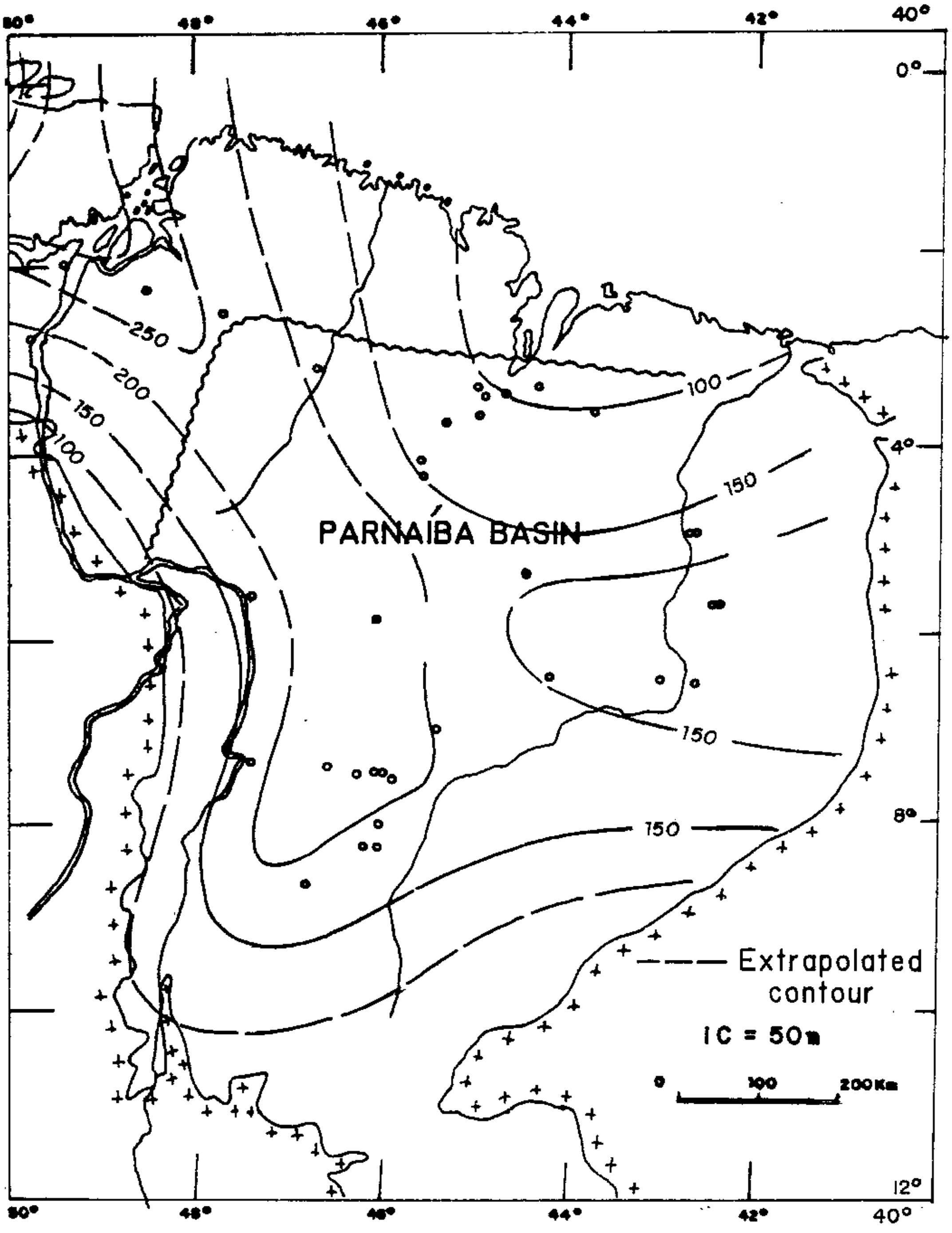


FIGURE 65- Isopach map of the Itaím Formation

is holes. Sideritic shale beds as well as oolitic, hematitic and chloritic shale beds are common in the section.

Small (1914) proposed an unconformity between the Serra Grande Group and overlying sediments. Kegel (1953) also commented that some observed features could be due to an unconformity at the base of the section, but because of the lack of more data, he considered there was no time gap in the stratigraphic column between Serra Grande and Itaím. He placed the Serra Grande Group in the Devonian Period, but also predicted that if an unconformity exist at the base of the section, the underlying Serra Grande Group (Formation) could have been laid down in Silurian times.

Beurlen (1965) pointed out that from the town of Picos southward, the Itaím beds pinchout so that the overlying Pimenteira shales are resting unconformably over the Serra Grande Group. In the western outcrop belt the Itaím Formation overlaps the Serra Grande Group and is overlapped by the overlying Pimenteira Formation. The contact between the Itaím sandstone beds with the Serra Grande sandstone beds is very difficult to detect as it also occurs between the Maecuru Formation and Trombetas Group in the Amazonas Basin. This is so because the lower Itaím Formation and the Jaicós Formation are very similar. The lower contact is placed in the section where sandstone beds with micaceous siltstone interbeds appear. A good characteristic of the uni its evidence of biological activity. The upper contact of the unit is either transitional where fine grained sandstone beds give way to shale beds of the Pimenteira Formation or is abrupt where medium-grained sandstone is capped by shales.

Few fossils are known in the outcrops, although the trace fossil *Spirophyton* sp. and bioturbated beds are common. *Spirophyton* sp. is known in North America from late Early Devonian to Pennsylvanian (Caster, 1952). Kegel (1953) listed some fossils from the Carolina well (1-Cl-1-Ma) that he supposed to be located at the top of the Cabeças Formation, which is placed higher in the column instead of at the top of the Itaím Formation, as I determined by well correlation. For age determination, important forms are: *Amphigenia* sp., (late Early Devonian and Mid-Devonian), *Eodevonaria* sp. (Early and Mid-Devonian), *Conularia undulata* Conrad (Mid-Devonian of the United States or late Early Devonian of Bolivia) and *Tentaculites stubeli* Clarke, also known in the Maecuru Formation of the Amazonas Basin (Kegel, 1953).

The presence of plant· remains and *Spirophyton* sp. does not support a Silurian or Ordovician age for the Itaím Formation as suggested by Brito and Santos (1964, 1965) and Brito (1969) on the basis of palynological data. This palynologically based dating could indicate a simple case of misidentification (with perhaps the Serra Grande Group being mistakenly identified as the Itaim Formation.

In the Carolina well (1-Cl-1-Ma) at the top of the Itaím Formation, considered as the top of the Cabeças Formation by Kegel (1953), the fossil assemblage suggests a late Early Devonian to Middle Devonian age. In the subsurface, the section yielded sporomorphs which place it in the biostratigraphic interval IV (Emsian and lowermost part of the Eifelian Stage). This corresponds to late Early Devonian to early Mid-Devonian age (Daemon and Contreiras, 1971a,b).

The Itaím Formation correlates with the Maecuru Formation of the Amazonas Basin. Both units indicate a simultaneous transgression in the area in late Early Devonian as occurred in the southern Sahara region, in the Tafassaset, where Early Devonian sandstone lies on top of the lower part of the Early Silurian shales (Biju-Duval and others, 1969). This transgression did not reach most of the present basin margins. Southward from the town of Picos the Itaím Formation thins. Its equivalent unit was not identified in the Accraian Series.

I interpret the Itaím Formation as representing a new depositional cycle in the Parnaíba Basin with two regressive phases at its upper part. The presence of abundant bioturbated beds and siderite enrichment under the form of oolites, nodules, and beds indicates low clastic supply and low reworking in profane and basin environments. Only during the regressive phases are a higher clastic supply and degree of noticeable reworking. The section seen at the surface suggests fan delta and fan delta front environment alternations. In the subsurface toward the west, pro-fan deposits are represented by shales. The two regressive sandstone bodies at the top of the unit may represent fan delta front environments. Some thin sandstone strata intercalated in pro-fan shales may indicate storm induced beds. The good sorting and roundness displayed by the sandstones may indicate that the unit was not a result of a first erosional cycle. The northeastern, eastern and southeastern regions were the main source of sediments for the basin. Toward the northwest the sandstone beds become thinner and less abundant and have a smaller grain size, increasing shale beds.

Although marine invasions occurred simultaneously in the Parnaíba, Amazonas and Paraná basins, no sea connection existed between the Paraná and Parnaíba basins, because Givetian sediments onlapped the Emsian-Early Eifelian Itaím Formation, part of the Serra Grande Group, and the basement, particularly in the western and southwestern parts of the basin. In the southwestern side of the basin, the overlying Pimenteira Formation is presently onlapping the Itaím and part of the Serra Grande Group. In the Paraná Basin, Emsian and Eifelian (late Early Devonian and early

Mid-Devonian) sediments are onlapping the basement and they are onlapped by Givetian (late Mid-Devonian) beds toward the northern basin margin (Northfleet and others, 1969). This stratigraphic relationship conflicts with the concept that during Emsian and Eifelian times the fauna could migrate from the Parnaíba to the Paraná Basin or vice-versa (Grabert, 1970). No confirmed equivalent Emsian and Eifelian marine fossiliferous rocks were found in the area between the Parnaíba and Paraná basins, only some sandstone beds precariously correlated with the Serra Grande Group continental beds were found.

The presence of plant remains may indicate that the Parnaíba Basin had a slightly warmer climate than that of the Paraná Basin area (Copper, 1977), where no plant remains were found in Emsian and Eifelian times. The climate in the Paraná and Parnaíba basins may have been subarctic as can be deduced from the poor fauna diversity and absence of warm climatic indicators. At this time, however, no glacial sediments have been identified in Gondwana. The ice-caps, if they existed, should have been restricted to highlands far from the mentioned depositional basins.

PIMENTEIRA FORMATION

The name Pimenteira was introduced by Small (1914) to designate shales of about 20 M in thickness in the village of Pimenteira.

Plummer (1948) considered the Pimenteira Formation as composed of a lower shale member and an upper sandy member. Kegel (1953) named the shale section the Picos Member and a lower section not considered by Plummer (1948) the Itaím Member. Blankennagel and Kremer (1954) designated only the predominant shale section as Pimenteira (Kegel´s Picos Member) and this procedure has been followed by most geologists and by me. The lower Passagem Member of the overlying Cabeças Formation of Kegel (1953) is here considered as the upper part of the Pimenteira Formation following Beurlen (1965) and Campanha and Mabesoone, 1976).

Its maximum thickness is estimated to be over 500 m in a new depocenter of the basin. Figure 66 shows the isopach contour map of the Pimenteira Formation with a different depocenter of the Serra Grande Group.

At the outcrop area, the Pimenteira Formation consists of variegated, but mainly gray and black siltstone, mudstone, shale and ironstone beds with yellow and red fine- to medium-grained sandstone interbeds. In the northeastern portion of the basin the unit is made up of gray silty shale beds and very fine- to fine-grained sandstone beds. In the western part, sandstone and ironstone beds interlayered with shale and siltstone beds are present with conglomerates at the base.

In the subsurface, the formation comprises gray, black, dark gray and greenish dark gray, micaceous, commonly sideritic shale,

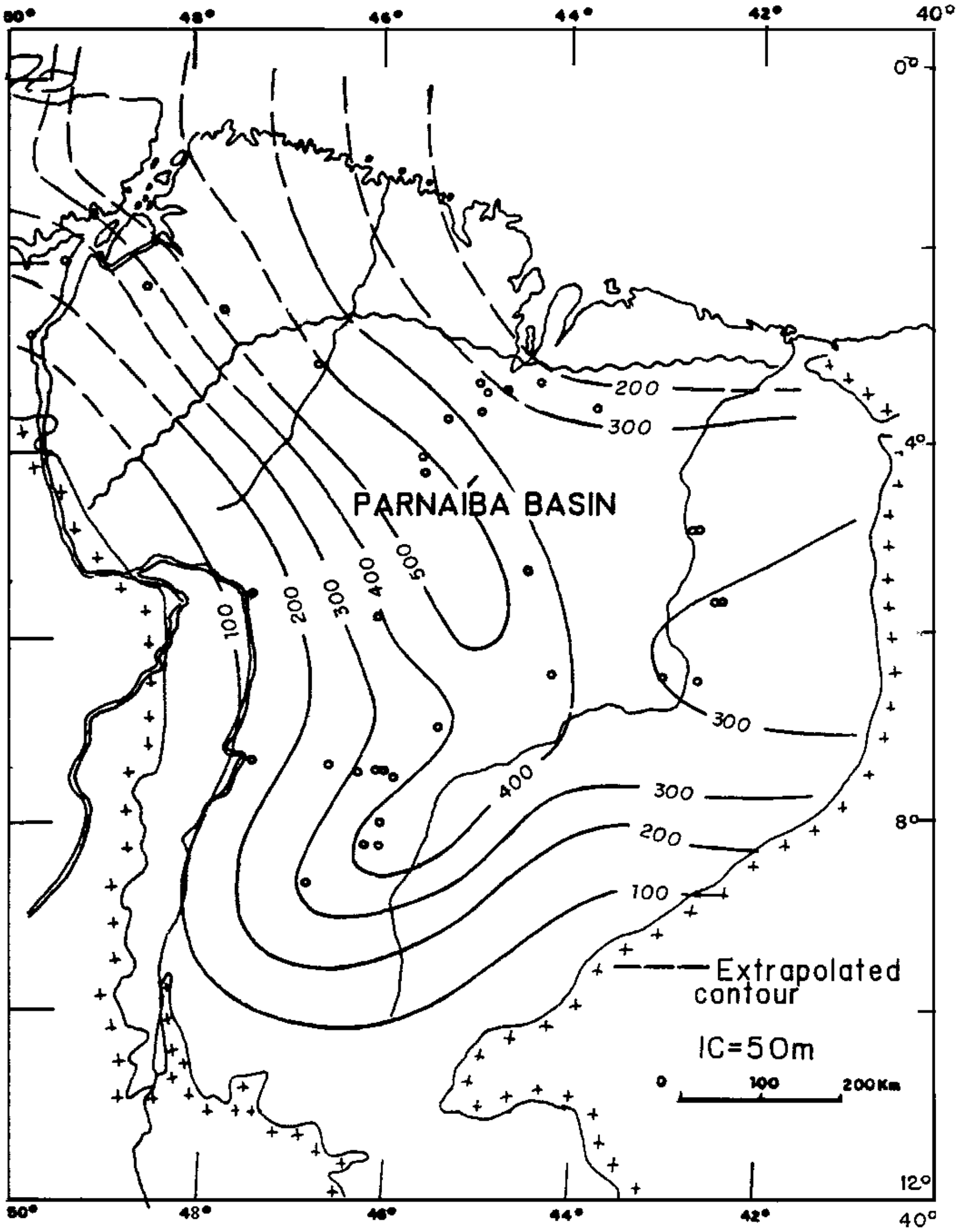


FIGURE 66- Isopach map of the Pimenteira Formation

silty shale and siltstone beds. Thin fine-grained sandstone beds, sometimes with small shale fragments, are also present in the section as observed in well cores. Several shale and siltstone horizons are bioturbated and rich in siderite. Where these beds are exposed they acquire red and purple oxidation colors.

The Pimenteira Formation rests conformably on the Itaím Formation or unconformably on the Serra Grande Group and on older basement sedimentary, metamorphic and igneous rocks. It overlies conformably as well as grading laterally into the Cabeças Formation. The Pimenteira Formation is also overlain by the Cretaceous Urucuia Formation. The isopach contour map show that the Pimenteira Formation may have been developed beyond the northern Brazilian coast.

Most of the Devonian fossils found in the basin come from the Pimenteira Formation. The richest horizon is located at its base. The unit also contains large plant remains (Kräuse and Dolianiti, 1957).

The mid-part of the Pimenteira Formation yielded *Burmeisteri* sp., *Metacryphaeus* sp., and *Tropidoleptus* sp. which suggests a Middle Devonian age. *Metacryphaeus* sp. (*Asteropyge*) is similar to one found in the Ererê Formation of the Amazonas Basin. *Tropidoleptus* sp and *Chonetes* cf. *systalis* are also found in North America in Middle Devonian rocks (Kegel, 1953).

In the western side and in the subsurface, the upper part of the unit is younger due to facies changes with the Cabeças Formation and its age is based on palynological content and on plant fossils. Because the Devonian-Carboniferous section in the western part is chiefly composed of siltstone, shale and diamic- tite, an arbitrary upper boundary is placed at the base of the diamictite beds because the diamictites are intimately connected with the overlying Cabeças Formation. The uppermost beds of the Pimenteira Formation contain the trace fossil *Spirophyton*, and algae or land plant *Protosalvinia* sp. (Carozzi and others, 1975) in some places, indicating an Early Fammenian age. Considering this boundary, the age of the Pimenteira Formation ranges from the Eifelian to Early Fammenian (early Mid-Devonian to mid-Late Devonian). This corresponds to the biostratigraphic intervals V, VI and VII.

In the eastern portion of the Parnaíba Basin, the Pimenteira Formation correlates with the Ererê Formation and part of the Barreirinha Formation of the Amazonas Basin. In the western part of the Parnaíba Basin, the Pimenteira Formation correlates with the Ererê and Barreirinha formations and the lower part of the Curiri Formation (*Protosalvinia* zone) of the Amazonas Basin.

The unit also correlates with the Accra shale of Ghana and with the lowermost part of the Takoradi beds of Ghana where the

Devonian palynomorphs are similar to those occurring in the Pimenteira Formation (Bär and Riegel, 1974). According to these investigators the Devonian beds of Ghana should be interpreted as a direct continuation of the northeastern Brazilian Devonian beds (Pimenteira beds) because of their striking similarity.

Campanha and Mabesoone (1976), and Mabesoone (1977) interpreted the formation as a regressive unit deposited mainly in tidal flats and shallow lagoons separated by barrier bars under cold conditions. Ribeiro and Dardenne (1978) stressed that the genesis of the Pimenteira Formation as well as the ironstones corresponds to that of a tidal flat zone under tropical conditions.

Here the Pimenteira Formation is interpreted as a record of a worldwide Late Devonian transgression with sea-level oscillations. In the Frasnian Stage, laminated shale beds with high-level radioactivity indicate the maximum sea-level high-stand.

In the northeastern side of the Parnaíba Basin a silty shale section overlain by a fine- to very fine-grained sandstone section occurs twice. The shale sections, up to 70 m in thickness each represent pro-fan deposits and the sandy section represent fan delta front deposits. According to Fuzikawa and Souza Filho (1970) these deposits are good examples of large cyclothems. In the southeastern outcrop belt, coarse sandstone beds may record minor delta fan lobes. Probably the inland Pimenteira sea was non-tidal as can be inferred by extensive bioturbation, which indicates poor current activity. In early stages of transgression over the craton, as was the case along the western Parnaíba Basin margins, the area under tidal action would be narrow and the dominant mode of deposition would be shoreface and storm-dominated (klein, 1982).

Extensive siderite precipitation and bioturbations suggests a starved basin where the clastic supply was very low in the inter-fan and shoreface areas. Although no deltaic bodies occur in the western part of the basin, where ironstones are thicker, Carozzi and others (1975) pointed out that in the Parnaíba Basin a chamosite matrix, partially replaced by siderite, occurs in delta front situations at times of progradation. In addition, it also occurs at an equivalent depth at times of destruction of deltas during transgressions under tropical conditions as reported by Porrenga (1967) in the Niger delta.

The "chemostte" name given by Porrenga (1967) for poorly organized clay is somewhat misleading according to Kimberley (1979) because it has a typically greater magnesium percentage than true chamosite, the abundant iron mineral in iron formations. The "chamosite" has formed diagenetically within tropical and also non-tropical sediments rich in decaying fecal matter (Porrenga, 1967).

It is interesting to notice that the Amazonas River yields 0.03 ppm Fe, while the European rivers yield about 0.8 ppm Fe (Kimberley, 1979), that is, more than 25 times, indicating that cold climate rivers deliver more iron to the sea than tropical ones. Iron minerals are forming today in tropical shelves as well as in basins with restricted circulation such as fjords, the Black Sea, poorly drained and recently glaciated regions or areas of older iron-rich rocks (Blatt and others, 1980), so the genesis of ironstones is not restricted to tropical environments because Fe solubility is to a large extent a function of high pH values. The pH of water is about 4 and 5 in tropical areas. The presence of vast economic Precambrian iron-formation areas to provide sources, which drain toward the Parnaíba Basin with cold climate and low clastic supply, may explain the higher iron content in Paleozoic sediments in western than in eastern Parnaíba Basin.

There is a gradual facies change basinward from limonite and goethite at the basin margins to siderite-clay ironstone with inclusions of oolitic goethite ironstone and from this to pure siderite-clay ironstone. Oolite chamosite ironstones increase basinward but they are present in all ironstone facies (Kimberley, 1979).

The energy of the environment may have been provided by wind-driven waves and occasional storms where nearshore oolites are generated. During minor sea-level lowstands, large areas covered by iron-rich sediments may have been exposed, and siderite and chamosite oolites may have been subjected to oxidation in place, which resulted in iron crusts. The erosion of these iron deposits produced cross-bedded goethite oolite, sandstone, conglomerate beds with broken fossils (Ribeiro and Dardenne, 1978) which may have accumulated in new depositional sites. Sandstone beds with clay chips are suggestive of resedimentation processes.

The climate during the deposition of the lower part of the Pimenteira Formation may have been arctic or subarctic as interpreted from the nature of the fauna (Copper, 1977), scarcity of limestone and absence of red beds, reefs and evaporites. The presence of plant remains, calcareous cement and very thin lenticular clayey limestone beds suggest warmer conditions in Middle Devonian time than in early Late Devonian time (Frasnian Stage). During the Frasnian Stage, the area may have been under periglacial (subarctic) conditions as inferred from the presence of overlying tillites of the Cabeças Formation.

Meyerhoff and Meyerhoff (1972) used the faunas found in Ghana as an argument against the new global tectonics. They stressed that the similar Early Devonian faunas in North America, Bolivia, South Africa, Amazonas, North Africa and Ghana indicate widespread open-marine conditions throughout the length and breadth of the present Atlantic region. However, the distribution of the Devonian related faunas may be explained without the presence of the supposed Atlantic Ocean. Considering Gondwana separated from Laurasia by a small ocean (Phoibic Ocean), a marine communication from North America across Venezuela, Colombia, Peru, Bolivia, Argentina, South Africa and Paraná Basin is feasible and the communication of North America across Taoudeni, Amazonas, and Accra basins through the Parnaíba Basin is also possible, without invoking the presence of the Atlantic Ocean. According to Caster (1952), Ghanaian Devonian fossils have more affinities with the Amazonas Basin ones than with those of other North African basins. The Devonian outcrops of Ghana are probably marginal deposits of the Brazilian Parnaíba Basin (Petters, 1979).

CABEÇAS FORMATION

The term Cabeças Formation was introduced by Plummer (1948) to designate a Devonian section composed of medium- to coarse-grained sandstone beds. Plummer (1948) divided the formation into three members, but according to Beurlen (1965) the lowermost part of the lower Passagem Member has more affinities with the Pimenteira Formation and its upper part has the same characteristics as the middle Oeiras Member. The upper Ipiranga Member was included in the overlying Longá Formation by Aguiar (1971).

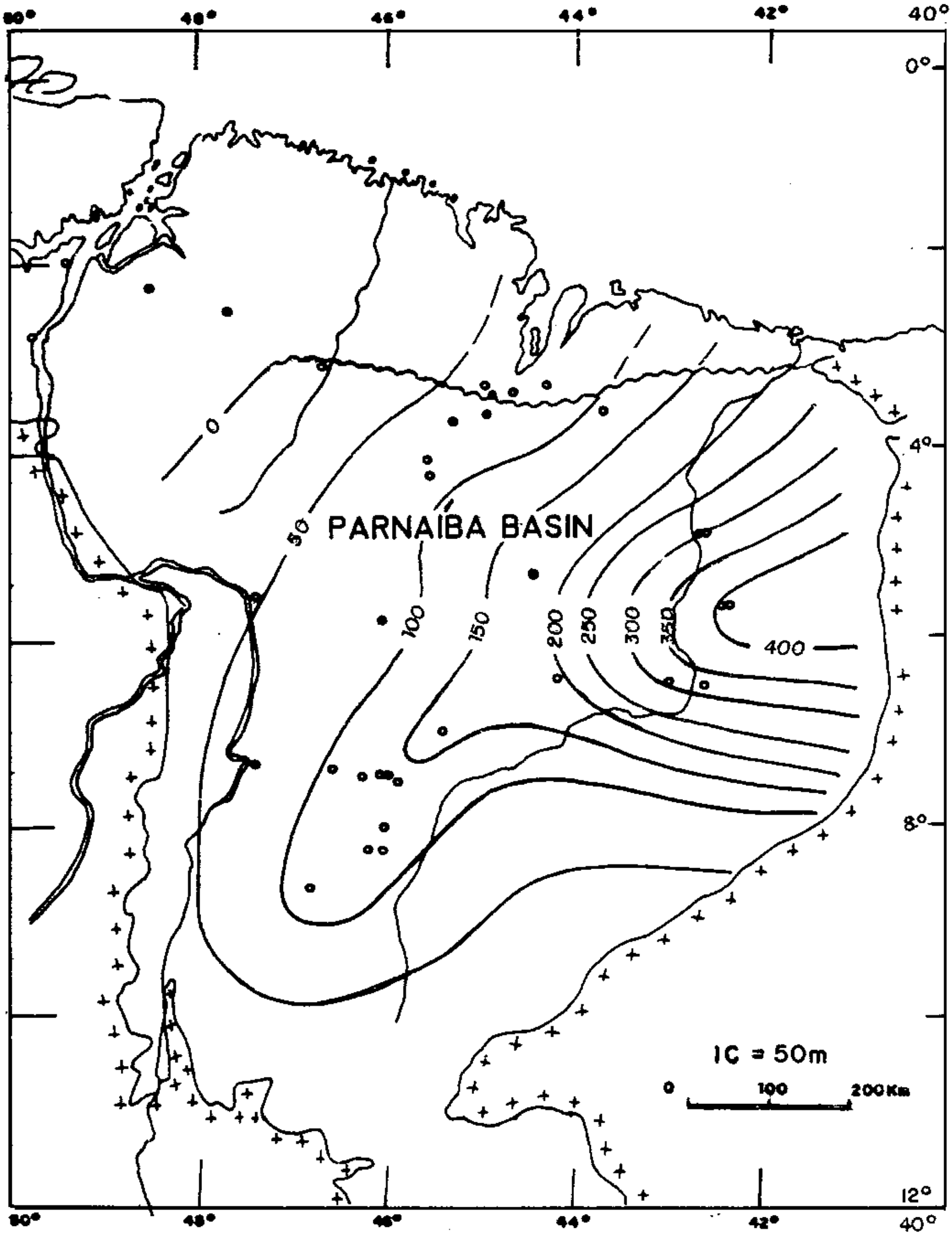


FIGURE 67- Isopach map of the Cabeças Formation

Here the subdivisions of Beurlen and Aguiar are also followed so that the Cabeças Formation is made up of the upper sandstone beds of the Passagem Member and the whole Oeiras Member of Kegel (1953). Figure 67 shows the isopach contour map of the Cabeças Formation, with a depocenter in the eastern region.

The Cabeças Formation consists of about 100 to 400 m of medium- to coarse-grained, hard, cross-bedded to massive, light gray to white sandstone beds with some conglomerate and pebbly sandstone interbeds. In the cross-bedded strata, coarser and finer grains are commonly alternated along the stratification planes. The main characteristic of the sandstone is that in many horizons it is massive and in some places it displays slump structures. In addition, its surface may be broken into polygonal blocks by a joint system. This polygonal pattern is a characteristic of the unit in the entire basin where it is exposed. The stratified horizons are commonly made up of giant sand waves bedforms 10 to 15 m high and 100-200 m long. The unit is also typified by ruinform erosional features. The Lower Cabeças Formation is rich in trace fossils (skolithos). At the upper part of the unit, some striated surfaces occur below diamictite horizons (Photo 10).

In the southwestern region, of the basin the Cabeças consists of lower and upper sandstone beds separated in some areas by a zone of diamictites. The lower sandstone unit is composed of sandstone beds, light-gray to white, fine grained, well sorted, subangular to fairly rounded, micaceous, with sucrosic texture and medium-bedding. The diamictite zone consists of polygenic unsorted clasts included in an unstratified clayey and sandy matrix, and is highly polygonally jointed and exfoliated. The material is easily weathered.

The clasts, from sand-grain size up to 1 m in diameter comprise gneiss, schist, phyllite, siltstone, and iron oolite fragments from the underlying Pimenteira Formation. The upper part consists of medium- to coarse-grained poor- to fairly well-sorted, angular grains, kaolinitic matrix, micromicaceous cross-bedded, and medium bedding. Each unit varies in thickness, but the general trend is a thinning northwards. In the southeastern margin, the total thickness is about 80 m.

South of the town of Carolina, in the western side of the basin, the most part of the unit, up to 70 m in thickness, is composed of diamictite and mudstone. At the base (15 m in thickness) cross-bedded, fine- to medium-grained sandstone with scattered granules (0.5 cm across) change laterally to diamictite. This section is followed by 5 m in thickness of diamictite with soft sediment deformation structures containing clasts, up to 5 cm across, "floating" in an argillaceous matrix. Upwards, up to 15 m in thickness, fine- to very fine-grained, clayey, micaceous sandstone beds with scattered coarse-sized quartz grains are present.

The upper 40 m of the section consist of diamictite containing clasts of siltstone, sandstone, shale, oolitic ironstone and micaschist, up to 50 cm across, some striated, and supported within a micaceous silty and clayey matrix. This section shows many soft sediment structures. Thus, in the western portion of the basin glacial deposits are predominant and in the eastern side of the basin glaciofluvial sediments are most abundant.

In the subsurface, the Cabeças Formation divides into half a dozen sandstone bodies, separated by shale, siltstone, varve-like sediments (Photo 11), and diamictite beds, and it pinches out northward as it interfingers with the Pimenteira and Longá formations.

No macrofossils were identified in the Cabeças Formation up to date. The fossils listed by Kegel (1953) as belonging to his Ipiranga Member are actually located in the upper part of the Itaím Formation in the Carolina well (1-Cl-1-MA). When this well was drilled, Kegel (1953) did not know that the Cabeças sandstone pinched out or was eroded in that well. He considered the first thick sandstone beds found in the well to be the upper member of the Cabeças Formation which yielded many fossils of Early to Middle Devonian age. This sandstone belongs to the Itaím

Formation. Since then, this mistake has been perpetuated in the literature. Figure 68 shows the lithologic log of the Carolina bore hole. The ages are based on palynological studies (Andrade and Daemon, 1974; Lima, 1983, written communication).

It is interesting to note that the diamictite beds drilled in that well were placed by Kegel (1953) in the Longá Formation, because this formation should be located above the supposed Cabeças Formation, but the Longá is above the Cabeças Formation diamictites. Field geologists who have mapped the western outcrop belt considered the diamictite beds as pertaining to the Pimenteira Formation. Thus, diamictite beds were supposed to be present in three different formations (Pimenteira, Cabeças and Longá formations) also with distinct ages. Palynological data and rock corrélation suggest, however, the presence of an unconformity below the diamictite beds which truncates different older formations.

The Cabeças Formation is located in the biostratigraphic Interval VII and in the lower part of the VIII (Famennian Stage) on the basis of palynomorphs corresponding to mid-Late Devonian to late Late Devonian. The Cabeças Formation started its deposition in the uppermost part of the biostratigraphic interval VI in the eastern outcrop area.

The Cabeças Formation is correlated with the middle and upper Curiri Formation and lowermost Oriximiná Formation of the Amazonas Basin. The *Protosalvinia* zone is located in the lower part of the

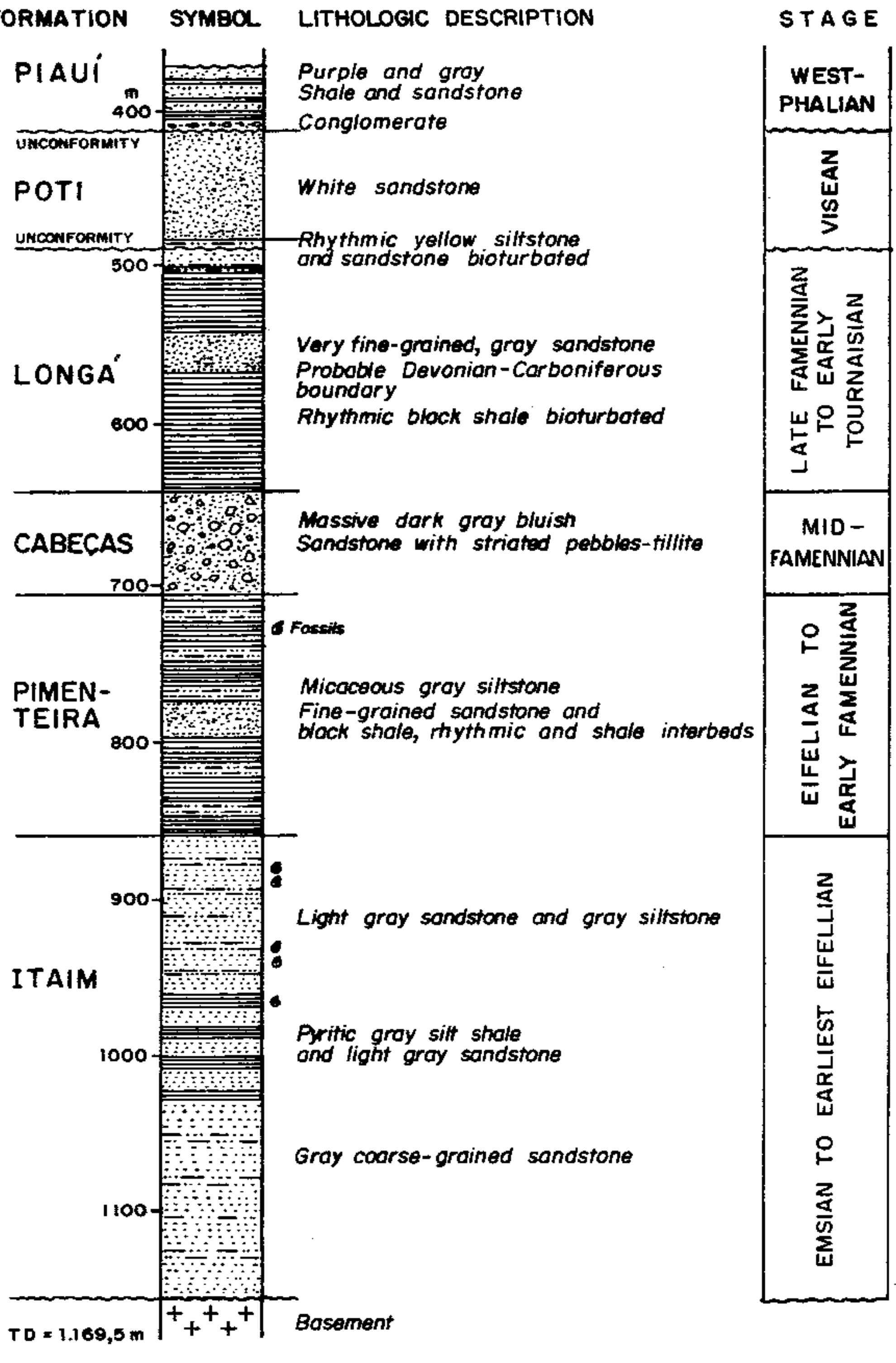


Figure 6 8 Modified partial lithologic log of the Carolina well (1-CL-1-MA) showing the tillites identified by Kegel (1953)

Curiri Formation (Amazonas Basin) and in the upper part of the Pimenteira Formation.

The Cabeças Formation also correlates with the Ibimirim Formation of the rifted Jatobá basin in northeast Brazil. It correlates as well with the upper sandstone beds of the Accraian Series and with the Takoradi sandstone beds of the Sekondian Series of the Accra Basin of Ghana (Crow, 1952; Bär and Riegel, 1974).

Some geologists have interpreted the Cabeças diamictites as a result of mass flow or turbidity currents associated with synchronous faulting or volcanism. As far as can be determined to date, however, no Devonian volcanic rocks or volcanogenic sediments exist in the Parnaíba Basin. In addition, it seems that no flysch-like sediments exists in this intracratonic basin.

The Cabeças Formation represents a regressive progradation and fast retrogradation of alluvial fan, fan delta and fan delta front systems and glacial lobes. At the base of the unit in the eastern outcrop area it contains sandstone beds with skolithos characteristic of very shallow marine environment, while in some western areas the basin was probably covered by ice sheets. Upwards, in the eastern area, the formation is composed of very coarse to conglomerate cross-bedded and massive sandstone beds with scattered quartzite pebbles and boulders (Photo 14).

The top of the sandstone is capped by diamictites, which in turn are covered by shales of the overlying Longá Formation. Shoreface, pro-fan and basinal environments correspond to the Pimenteira and Longá formations composed of shales and siltstones. At places between the uppermost diamictite and the Longá shales, sandstone beds of possible fluvial origin are present.

Most of the massive and cross-bedded sandstone beds developed in the eastern part of the basin may have been laid down under periglacial conditions. Slumped structures so common in the unit may have been produced by ice undermelting, and possibly fossil patterned grounds may have been generated in association with very low arctic temperatures. Diamictite horizons represent true tillites because Kegel (1953) described striated pebbles in wells, Barbosa (1966) also documented striated pebbles in outcrops and Malzahn (1957) and Bigarella (1973a,b) described a striated pavement overlain by diamictite interpreted as tillite and rhythmites with dropstones interpreted as varves. Carozzi and others (1975) reported dropstones in varves in cores from oil wells (Photo 12). I have revised these sections in the field and bore hole cores and I confirm their glacial evidences and nature. The presence of glacial and periglacial features in the sediments indicates an arctic and subarctic climate during the deposition of the Cabeças Formation in the Parnaíba Basin.

In many places marine shales are covering either the tillites or the uppermost sandstone beds of the Cabeças Formation, suggesting a very fast transgression after ice melting. In the eastern and southeastern margins of the basin, during most of Ordovician and Silurian times normal sedimentation was composed of coarse clastics; only when major sea transgressions took place the sedimentation of finer clastic predominated. During the sedimentation of the Cabeças Formation, the predominant coarse-grained diamictite source-areas were located at the south and west sides of the basin. The distribution of the upper part of the Cabeças Formation diamictites and striated pavements suggests that the entire basin was once covered by ice-caps.

LONGÁ FORMATION

The Longá Formation was established by Albuquerque and Dequech (1950) to designate a section essentially composed of shale. This practice was followed by many subsequent investigators, but Rodrigues (1967), Aguiar (1971) and Lima and Leite (1978) and many other geologists included the Ipiranga Member of the Cabeças Formation in the Longá Formation (Plummer, 1948; Kegel, 1953). Figure 69 shows the isopach contour map of the Longá Formation. The formation may be subdivided into three units and I describe

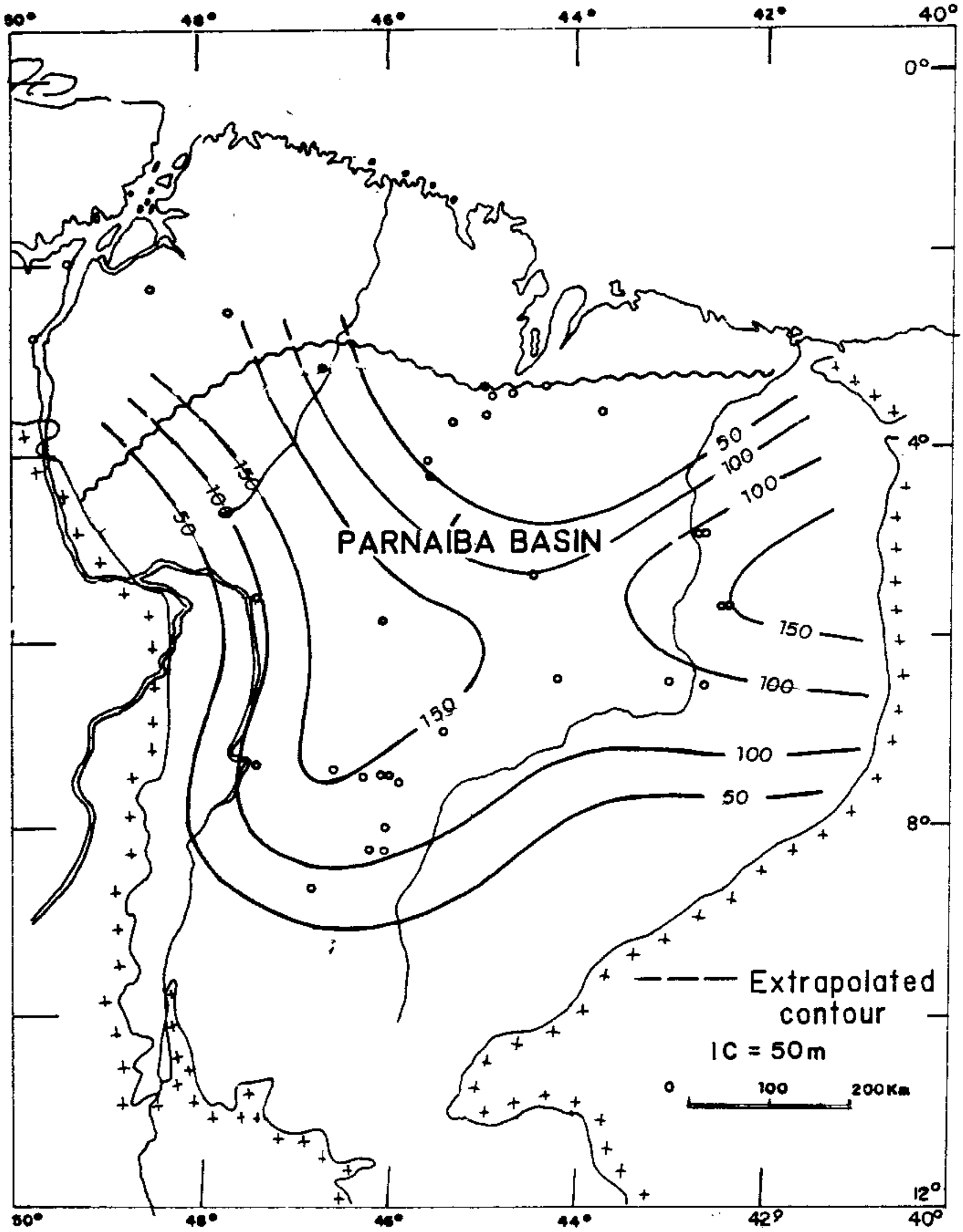


FIGURE 69- Isopach map of the Longá Formation

as follows: The lower unit comprises micaceous, sideritic, well laminated greenish gray shale, that is also bioturbated, dark gray, micaceous, and in part contains sideritic shale and siltstone interbeds with thin conglomerate lenses and conglomeratic sandstone beds. The middle unit consists of white or yellow ferruginous, sideritic in part, micaceous, argillaceous, cross-laminated and cross-stratified very fine- to medium-grained sandstone beds. Coarse-grained to conglomeratic sandstone beds are also present. The upper unit consists of greenish gray, dark, bituminous, sideritic, pyritic, micaceous, fissil, parallel and cross-laminated shale and, at places, bioturbated shale, silty shale and siltstone beds. Some thin clayey and sandy carbonate layers are present in the upper part of the unit. At the upper part of the section a conglormerate bed 30 cm thick may be present composed of quartz clasts up to 15 cm across with a sandy matrix. It is like a diamictite. Some clasts seem to be carried by ice. Upwards thin fine- to medium-grained sandstone interbeds become thicker and more frequent. Some thin fairly coarse sandstone interbeds are present throughout the entire formation.

The thickness of each member varies and the middle unit may pinch out completely. The total thickness of the formation is about 150 m. It overlies conformably the Cabeças Formation, but a small hiatus may exist at the basin edges where iron stained surfaces of the Cabeças Formation, interpreted here as palesoils or hardgrounds, are overlain by the Longá Formation. The upper contact is conformable in the central parts of the basin, but in the outcrop and flank areas it is disconformable with the overlying Poti Formation. This unconformity represents a period of subareal exposure and erosion along the margins and flanks of the basin as occurred at the top of the Oriximiná Formation of the Curuá Group of the Amazonas Basin.

Kegel (1953) assigned a Late Devonian age to the Longá Formation based on the presence of marine Devonian fossils and its position below Early Carboniferous sediments of the Poti Formation. Later Müller (1962) based on palynological data considered the Longá Formation as laid down in Late Devonian and Early Carboniferous times. The unit is located in the upper part of the biostratigraphic interval VIII, IX and in the lower part of the biostratigraphic interval X, corresponding to the Late Famennian to Late Tournaisian stages.

It is interesting to note that the boundary between Devonian and Carboniferous sediments is probably located within the lower part of the biostratigraphic interval IX, as deduced by Daemon and Contreiras (1971a,b).

The Longá Formation carrelates with the Oriximiná Formation of the Curuá Group of the Amazonas Basin. Probably the uppermost

Oriximiná Formation correlates with the overlying lower Poti Formation.

The unit also correlates with the stratigraphic section between about 2,585-2,677 m depth interval of the well Ibimirim (2-IM-1-Pe) drilled in the Jatobá Basin, northeastern Brazil. This section composed of a black shale with fine sandstone interbeds is provisionally considered here as Jaçu Formation. A Late Famennian-Early Tournaisian (Latest Devonian-Earliest Carboniferous) age was assigned to the Jaçu beds by De Quadros (1980) on the basis of palynological studies.

The isopach contour map shows that the Longá Formation was developed beyond the present northern Brazilian coast. In Ghana, in the Accra Basin, the Takoradi Shale of the Sekondian Series shows a striking similarity and a good age correlation with the Longá Formation (Mensah and Chaloner, 1971; Bär and Riegel, 1974; Crow, 1952). The upper part of the Takoradi Shale contains plant remains and sporomorphs of Early Carboniferous age. Taking into account the correlatable sediments in the Jatobá and Accra basins (Figure 70), the maximum water depth of the basin during deposition of the Longá Formation may have been limited in view of its enormous area and small thickness of about 50 to 150 m. Probably the original basinward slope was much less than one meter per kilometer.

The basal unit of the Longá Formation represents a fast

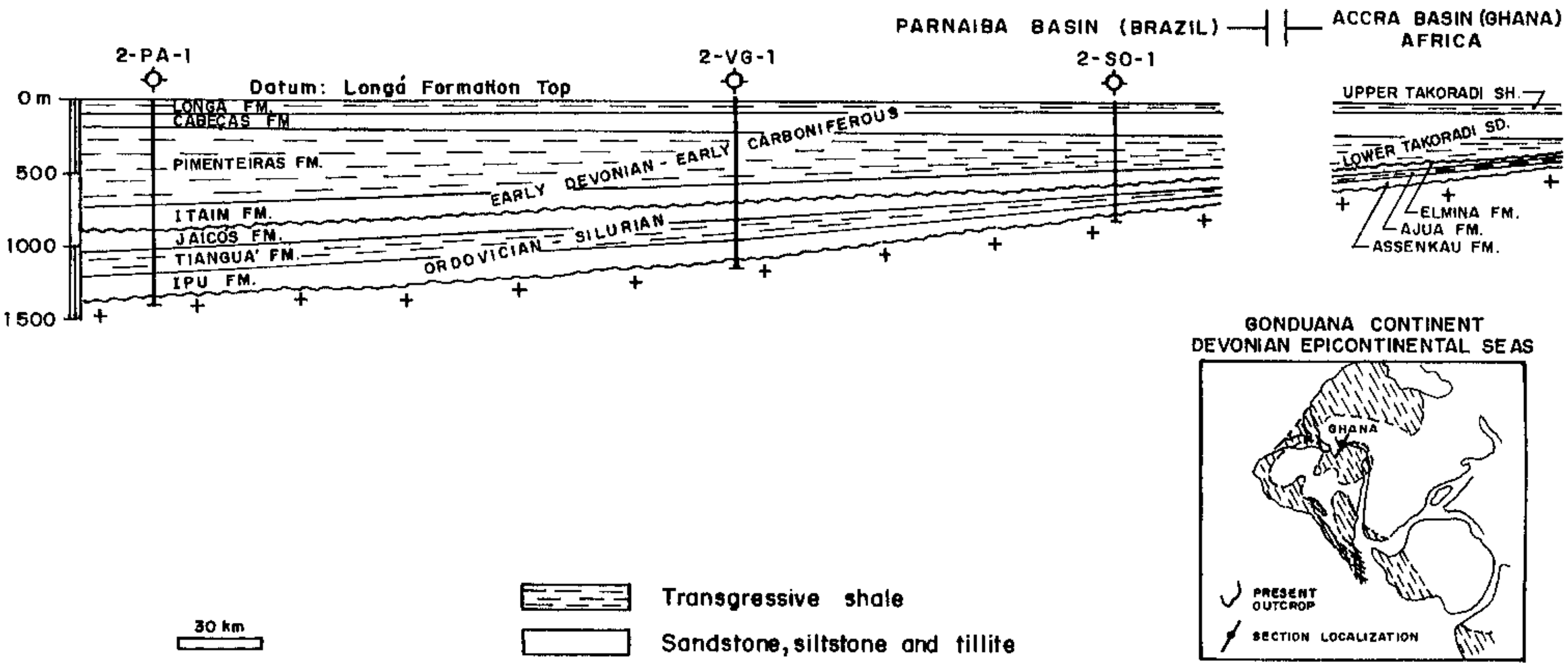


Figure - 70 Stratigraphic section showing past continuity between Accra (Ghana) and Parnaíba (Brazil) basins from Ordovician to Early Carboniferous time. The ghanaian section is referred in the text according to Crow (1952), Bär and Riegel (1974) and Talbot (1981) transgression, perhaps resulting from ice-melting, where shore-face and basinal deposits are present above fluvial sediments of the Cabeças Formation.

The shorelines of the unit may have fluctuated over a wide belt during sea-level oscillations due to the flatness of the depositional basin. This can be deduced by the presence of correlative basinal sediments in the Jatobá and Accra basins and proximal ferruginous conglomerate and coarse sandstone interbeds with indications of exposure in the Parnaíba outcrop area. Also, thin coarse- to fine-grained, at places, graded sandstone interbeds suggest storm-derived deposits.

The lower unit grades laterally into the Cabeças Formation. The middle unit suggests delta front and shallow marine sandstone deposits which may have resulted from an episodic regressive interval, but the cause of this fast regressive interval is not known. It may be glacio-eustatic in origin. In areas where the middle unit is not present, lateral interdeltaic bay and shoreface sedimentation may have occurred. The upper unit indicates a new transgressive shale deposition with more restricted circulation than that occurring in the lower unit. The presence of some thin clayey and sandy limestone layers may indicate a slight climatic amelioration during the deposition of the upper unit, and the climate might have been subarctic.

POTI FORMATION

The name Poti was first mentioned by Lisboa (1914) to describe sandstone beds cropping out in the Poti River Valley. In 1937, Paiva and Miranda established the Poti Formation to describe a section occurring between 219 and 566 m depth in water well No. 125, in the city of Terezina, but Campbell (1949) limited the Poti Formation to the depth interval of 219-423 m of the same well, excluding 143 m of section which belongs to the Longá Formation.

Later, palynological studies made by Müller (1962), Ludwig and Müller (1964, 1968) and Andrade and Daemon (1974) revealed an unconformity between the Poti and the underlying Longá Formation in the edges and flanks of the basin. They also observed that the Poti Formation incorporates more section in its central parts where no breaks exists. Ludwig and Müller (1964, 1968) proposed the Imperatriz Formation to designate the section of about 10 m thick in the Imperatriz well (2-IZ-1-Ma) below the typical Poti sandstone. However, in this marginal well, that section is also incomplete. Only in the central parts of the basin is the section complete. Note that the rock interval defined by Campbell (1949) as Poti Formation includes a more complete basal section, so the Imperatriz Formation is here included in the Poti Formation.

Figure 71 shows the isopach contour map of the Poti Forma-

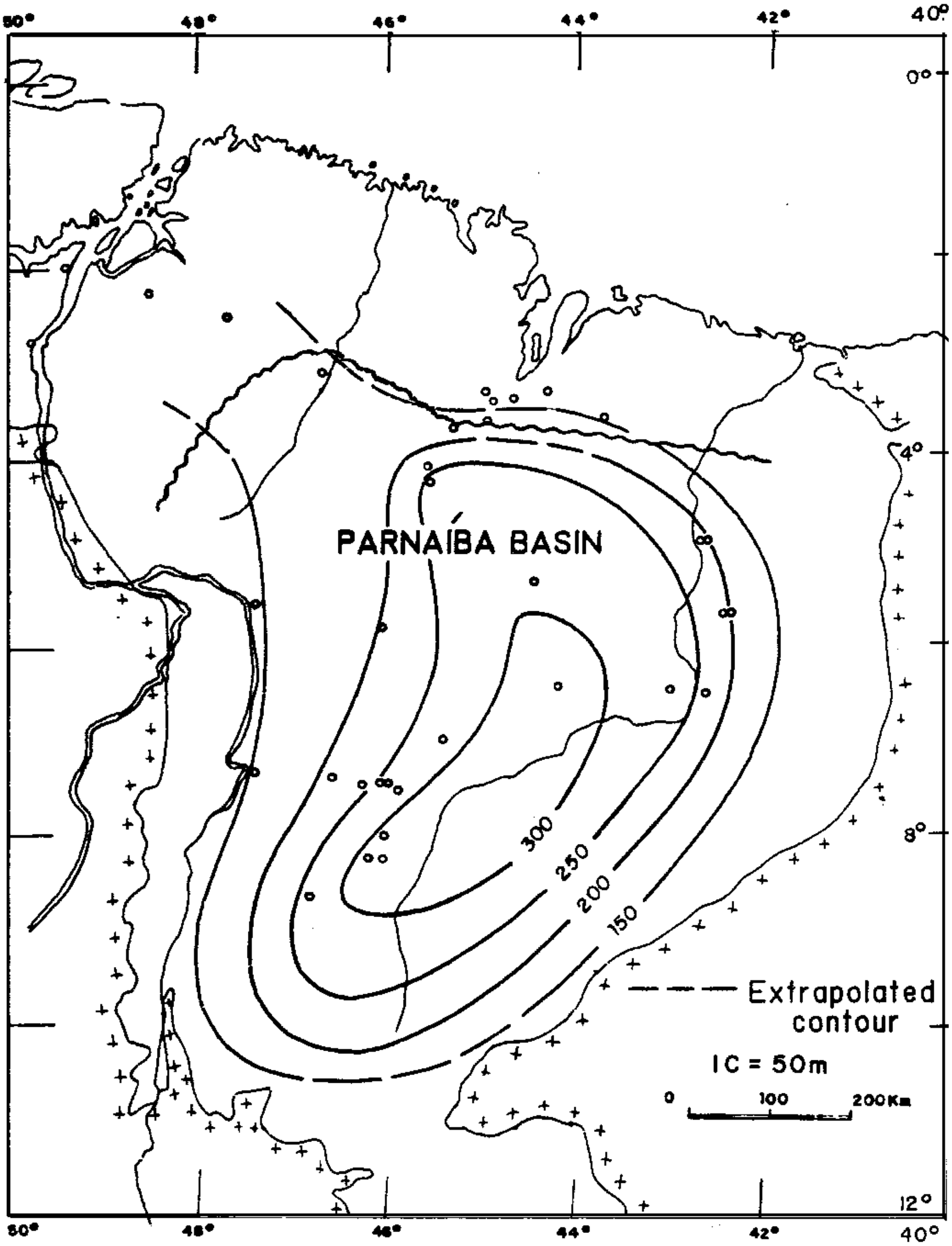


FIGURE 71 - Isopach map of the Poti Formation

tion. The Poti Formation, up to 300 m thick, may be subdivided into four units as I have observed in bore holes and the surface. The lower unit is composed of pink, friable, kaolinic fine- to medium-grained, in part, massive or with low angle cross-bedding, sandstone beds with scattered pebbles and boulders consisting of siltstone, quartz, sandstone, gneiss, and pegmatite rocks up to 30 cm across. In some horizons polymictic conglomerates occur with subrounded to rounded clasts. Siltstone interbeds are common close to its lower boundary. The second unit is composed of siltstone or silty shale, pink, micaceous parallel laminated, soft, and ferruginous. The third unit consists of pink fine- to medium-grained, massive sandstone beds with dispersed and unoriented quartzite, quartz, claystone, gneiss, granite and sandstone clasts up to 60 cm across. The upper and last shale unit is composed of variegated, mainly greenish light gray micaceous siltstone and shale beds which interfinger with pinkish purple massive diamictite with dispersed sand- and pebble-sized rockk fragments, within a massive silty and clayey matrix. The diamictite shows a peculiar ferruginous black alteration. Some very fine calcareous sandstone interbeds with plant remains are also present. The uppermost part of this unit is composed of some calcareous sandstone beds with siltstone and shale interbeds, with plant remains and very thin coal beds (1 mm to a few centimeters thick). The lower shale unit is not persistent throughout the basin, but it can be seen in the western and southwestern outcrop belt (Moore, 1963; Andrade, 1968) and in bore hole cores.

In the eastern part of the basin, Mabesoone (1977) pointed out that the entire Poti Formation consists only of sandstone beds. However, in some eastern regions the Poti Formation shows a thin shaly unit with sandy limestone interbeds at its upper part (Aguiar, 1971). Massive coarse- to very coarse-grained conglo-meratic sandstone beds with abundant sandy and silty matrix occur throughout the lower section. The sandstone bodies contain subangular to subrounded quartz pebbles. In many sections, the formation shows soft-sediment deformation (Kegel, 1955) and in the upper unit Ojeda and Bembom (1966) recordered the presence of sandstone dikes and veins 3 cm thick and 20 m long. These dikes and veins I interpret as fossil ice-wedges filled with sand after the ice-melting in a periglacial environment.

The Poti Formation overlies conformably the Longá Formation in the central parts of the basin, but in the basin flanks and outcrop belt the upper part of the Longá Formation may be absent.

The Poti Formation is unconformably overlain by the Piauí Formation. In many places, basement boulders up to 1 m across are observed in the upper contact zone in the western part of the basin. This means that quite large basement boulders were transported to the basin, more than 100 km away from the present basin edge. It seems that the boulders are residual from the Poti Formation rather than transported by Piauí paleorivers. The Poti Formation yielded some marine fossils in its lowermost section and continental plant remains in its uppermost unit.

The lower part of the formation contains marine strata with *Edmondia* and *Lingulidiscina* (Mesner and Wooldridge, 1964; Duarte, 1936) which led Kegel (1953; 1954) to consider it as deposited in Early Carboniferous time. The upper shaly unit yielded a rich flora of *Adiantites, Sphenopteris, Triphylopteris, Rhodeopteridium* (Rhodea), *Lepidodendropsis, Cyclostigma,* etc. (Dolianiti, 1972 in Rocha-Campos, 1972). The plants, which compose the Poti Adiantites flora, are found in the Lower Mississippian beds of North America. They show indisputable boreal characteristics with Euramerican affinities that are quite different from the Gondwana Glossopteris flora (Dolianiti, 1972, in Rocha-Campos, 1972).

On the basis of palynological studies, Andrade and Daemon (1974) placed the Poti formation in local biostratigraphic intervals X, XI and XII, which corresponds from Late Tournaisian to Visean stages (Early Carboniferous). The Poti Formation contains the same palynomorphs as those of the Faro Formation of the Amazonas Basin, but its deposition may have begun earlier.

The Poti Formation correlates with the Faro Formation of the

Amazonas Basin, based on palynological and lithological similarities. Both units record a similar depositional history. The formation also corrrelates with the Moxotó Formation in Jatobá and Tucano basins. Correlation was found with the uppermost beds of Takoradi beds of the Accra Basin.

At the end of the Longá Formation deposition (Latest Tournaisian Stage) a very sudden regression took place in the basin. The sea was then restricted to the central part of the basin where sedimentation continued with no apparent gap while the basin flanks were being exposed. The regression was so quick that no regressive coarse sediments were deposited in the top of the Longá Formation in the basin flanks and edges, and if they were deposited, they were readily removed before the new sedimentary cover was laid down.

The lower portion of the Poti section, in the central part of the basin, records a general coarsening upwards texture. The lowermost section may represent fan delta front and shoreface environments followed by alluvial fan deposits under periglacial conditions. It seems that much glacial sediment was reworked in periglacial outwash streams in source areas near the edge of ice-caps. Massive sandstone with scattered pebbles, soft sediment deformation structures and ruinform topography are the only indications of a periglacial environment in the lowermost unit of the

Poti Formation. Figure 72 shows the stratigraphic framework of the uppermost Pimenteira, Cabeças, Longá and lowermost Poti formations under glacial and periglacial conditions.

The second unit composed of shale and siltstone interbeds may represent isolated lakes, because in many sections the shales disappear giving way to a continuous sandy section. The third unit may represent outwash deposits and sandy diamictites. The fourth unit, composed of siltstone and diamictite, may record glacial and lacustrine environments. The presence of sand dikes suggests fossil polygonal patterned grounds and ice wedges. The uppermost part of the Poti Formation, with some coal thin beds and films, may record a fluvial environment with flood plain deposits. The deposition of diamictites, lacustrine and fluvial sediments in the Poti Formation, as well as in the Faro Formation of the Amazonas Basin, at the same time that tillites were being laid down in the Pimenta Bueno graben and Subandean belt basins, reinforces the idea that all these sediments were deposited under glacial and periglacial conditions.

The amount of sandstone deposited in the Chorro (Bolivia), Sernambi (Solimões Basin), Faro (Amazonas Basin}, Poti (Parnaíba Basin) and Moxotó (Jatobá Basin) formations under glacial and mainly periglacial conditions is very impressive. The climate may have been arctic and subarctic. Red diamictite beds occur in the

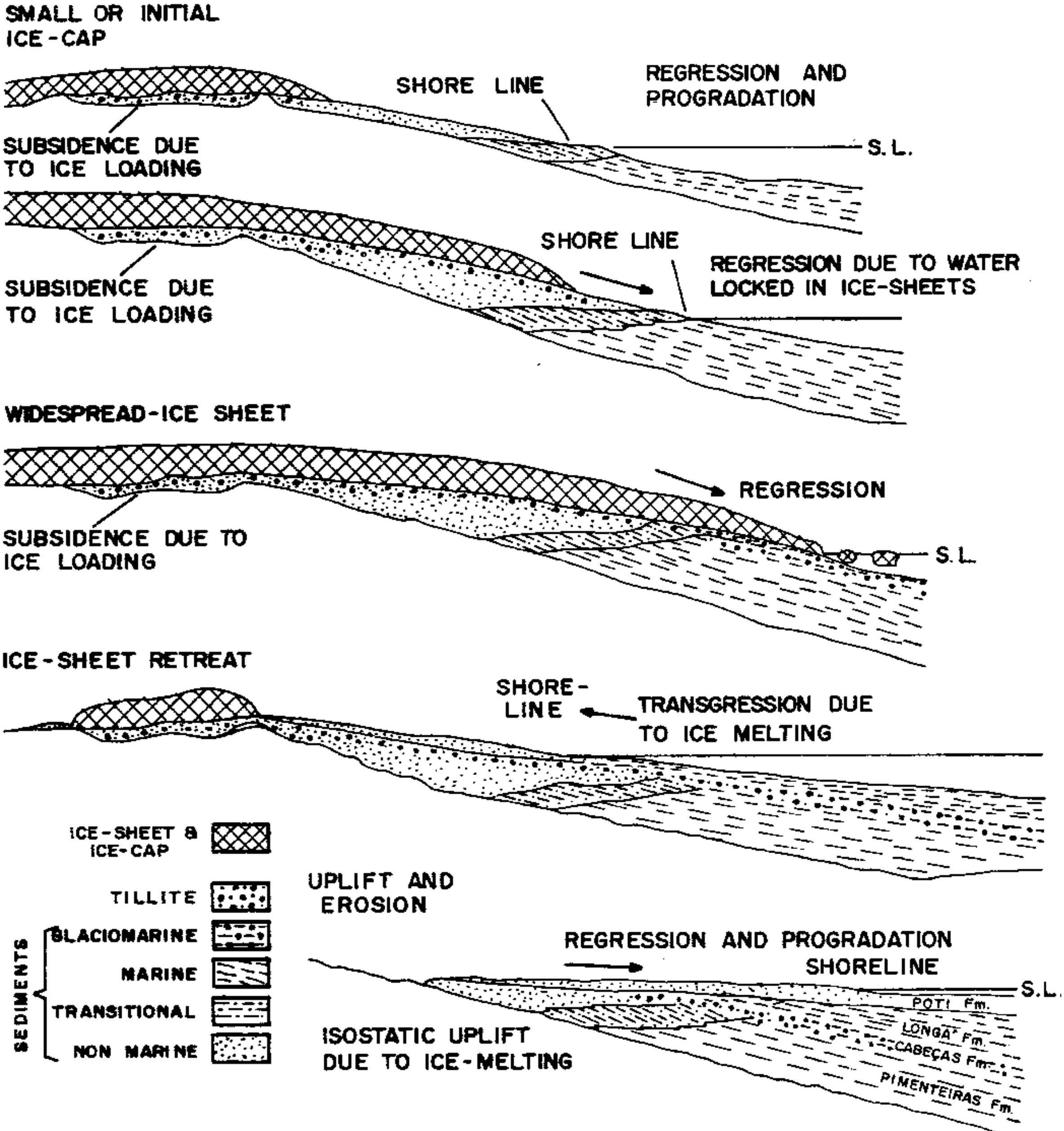


Figure- 72 The figure shows the stratigraphic framework of the Pimenteira, Cabeças, Longá and lowermost Poti formations under glacial and periglacial conditions in the Parnaíba Basin.

Poti, Pimenta Bueno and Early Carboniferous glacial formations of the Andes. Some plant remains, calcareous sandstone, sandy limestone and red beds may indicate a climate amelioration during the deposition of the uppermost beds of the Poti Formation. The deposition of the Cabeças Formation occurred under colder climate than that of the Longá and Poti formations. The relatively poor development of coal beds (films) may indicate a somewhat dry and warmer climate at the end of the deposition of the Poti Formation.

The regression that preceded the Poti deposition may have been related to water removal of the oceans due to the growth of ice caps in the Gondwana continent. After the deposition of the Poti Formation in Namurian time worldwide regression provoked erosion in the Parnaíba Basin.

A fundamental change in basin climate and tectonics started to occur in the Middle Carboniferous time. The Marajó arch started to uplift, cutting off communication with the Phoïbic Ocean. The Ferrer arch, which was a continuation of the Marajó arch into the Parnaíba Basin and the coastal Brazilian shield started to rise while the basin axis of greater subsidence shifted 500 km toward the west (Mesner and Wooldridge, 1964). In the western and northwestern parts of the basin, where argillaceous sediments had been deposited since the Early Silurian, a greater compaction may have occurred contributing also to the westward shift of the depositional axis.

PIAUÍ FORMATION

Small (1914) first used the name Piauí Series to include the whole Paleozoic roks of the Parnaíba Basin. Oliveira and Leonardos (1943) redefined the term Piauí Formation as corresponding to the rock interval located above 219 m depth in the well no. 125 drilled in the city of Terezina.

Mesner and Wooldridge (1964) and Aguiar (1971) considered the unit as made up of two members. The lower member is composed mainly of sandstone bodies and the upper member of sandstone bodies separated by shale, siltstone and chert intercalations. The chert beds originated from replacement of limestone beds. Mabesoone (1977) recognized three members in the unity, where the lower member consists of sandstone beds, the middle limestone beds, and the upper, sandstone beds. Figure 73 shows the isopach contour map of the Piauí Formation.

The Piauí Formation, up to 330 m in thickness, is made up of sandstone, brick-red or pink and white, with well- sorted and rounded grains, frequently frosted, slightly micaceous, kaolinic, and commonly highly cross-bedded. It is polygonally jointed at the base.

Fine- to coarse-grained sandstone beds with conglomeratic

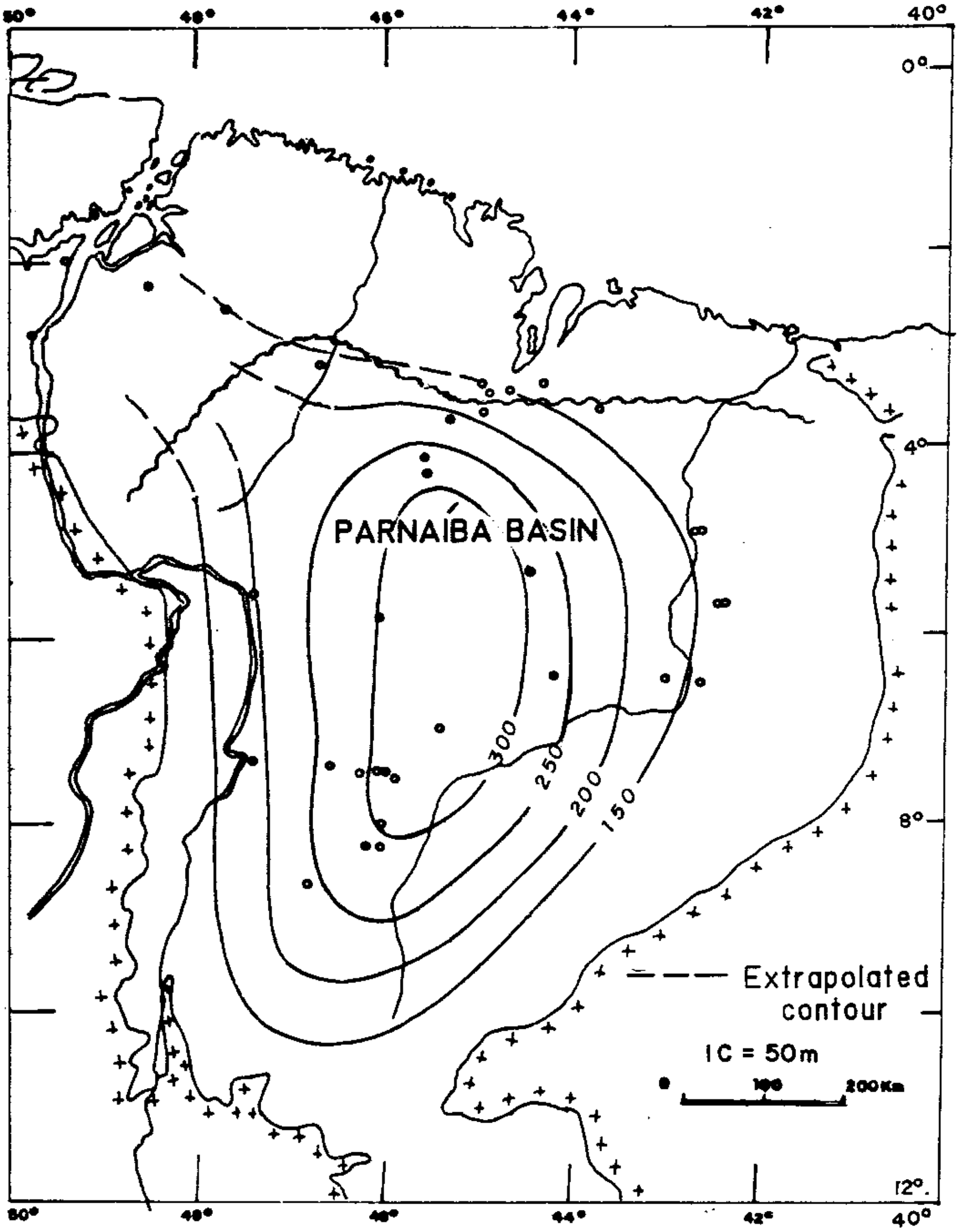


FIGURE 73 - Isopach map of the Piauí Formation

lenses are also present. Red maroon, hard, at places silicified shale and siltstone beds with limestone or chert interbeds are intercalated mainly in the upper part of the section. Large-scale cross-bedding is a characteristic of the unit.

In the subsurface, toward the northwest part of the Parnaíba Basin, shale, limestone and anhydrite beds dominate, however, in the lowermost and uppermost parts of the section, sandstone beds predominate. In the northwest part of the basin the unit presents similar cyclic characteristics as those of the Itaituba Formation of the Amazonas Basin. Scattered sand grains are included in rhythmic lacustrine shales and ventifacts are dispersed in cross-bedded sandstone beds in some sections.

The Piauí Formation rests unconformably on the Poti Formation and is conformably overlain by the Pedra de Fogo Formation. It is also unconformably overlain by the Mesozoic Sambaíba, Pastos Bons, Urucuia and Areado formations (Lima and Leite, 1978). The Piauí Formation thins toward the southeastern and eastern regions, indicating a reduction in the size of the basin in those directions. In the northern area deposition may have been restricted to isolated basins.

In the limestone intercalations a fairly rich marine fauna and flora are present, indicating a Late Carboniferous age. The fossil content suggests Amazonian and Andean affinities. Mesner and Wooldridge (1964) based on palynological studies made by Müller (1962) stressed that the unit was deposited from Westphalian to Stephanian stages.

The Piauí Formation correlates on the basis of macrofossils, palynomorphs and lithology with the Monte Alegre and Itaituba formations, although its lower beds could be somewhat older than the Monte Alegre Formation. It correlates with the lower Efia Nkwanta beds (Crow, 1952) of the Accra Basin of Ghana. The lower Efia Nkwanta beds are composed of aeolian and fluvial sandstones beds.

In most parts of the outcrop area, the Piauí Formation was deposited in continental aeolian and fluvial environments. Interbedded wind- and water-transported sediments are evidenced in the field by fine-grained sandstone with large-scale cross-bedding and reworked ventifacts included in coarse-grained to conglomeratic sandstone beds. The sea ingressed many times in the area during the deposition of the unit. The marine, lagoonal and lacustrine intercalations are represented by very fine-grained sandstone, shale, limestone and anhydrite beds. These marine intercalations characterize cyclic deposition. The sparse sand grains included in rhythmic shale beds are interpreted as air-transported into lakes isolated from the sea during drier seasons.

It is striking that thin marine limestone tongues were accumulated close to the eastern tip of South America, more than 4,000 km from the Paleopacific Ocean. The Late Carboniferous sea ingressed the northern Brazilian basins throughout the Andean, Solimões and Amazonas basins. Only the highest sea-level stands were recorded in the eastern outcrop area of the Parnaíba Basin, but in the subsurface in the northwest side, the marine intercalations are more frequent while the continental intercalations are less common. This marine environment was certainly non-tidal and very shallow. Anhydrite beds mark the regressional phase of each cycle. The presence of anhydrite, limestone, aeolian sandstone and red beds suggests that the area was localized in the zone of trade winds of high evaporation during the deposition of the Piauí Formation.

Here it is appropriate to discuss briefly a paper written by Meyerhoff and Meyerhoff (1972) in which in their Figure 8 a large area appears in the Parnaíba Basin with coal beds. Considering the area covered with coal beds it seems that more coal exists in the Parnaíba Basin than in the easternmost part of North America !

The Late Carboniferous and Permian coal present in the Parnaíba Basin occurs beneath gypsum beds with a thickness measured in millimeters (Barbosa and Gomes, 1957). This may indicate a short-lived algal mat development before dry seasons. In fact the climatic conditions may have been adverse for production and accumulation of significant coal. Although a great number of surveys were made in the basin, with the objective of finding coal deposits, the results were discouraging. Only coal films and coal beds, less than 5 cm thick, are present in the eastern side of the basin in the Early Carboniferous Poti Formation (De Paiva and Miranda, 1937), thus, the existence of thick coal deposits in the Parnaíba Basin is a myth.

PEDRA DE FOGO FORMATION

The name Pedra de Fogo was first introduced by Plummer (1948) to define a sequence of shale and chert beds containing petrified wood (*Psaronious*) located between the towns of Pastos Bons and Nova Iorque. Figure 74 shows the isopach contour map of the Pedra de Fogo Formation.

The Pedra de Fogo Formation, up to 240 m in thickness, is characterized by a great variety of sedimentary rocks. It is composed of shale, siltstone, sandstone, anhydrite, chert and limestone beds. According to Aguiar (1971) these rocks form cyclothems. He was able to recognize four major cycles, on the field, each with 20 to 30 m in thickness, but small cycles also exist. In a composite cycle the following rocks are present, from the bottom upwards: maroon calcareous shale; light green siltstone; yellow to green, calciferous sandstone with wood remains; yellow to green siltstone; green shale or whitish-pink limestone or chert with concretions; greenish gray shale with plant remains; oolitic limestone or limestone with dark gray to green shale interbeds; pink to green shale with wood remains; and purple, blue or green laminated shales.

In the upper cycle, Aguiar (1971) described thin coal beds (a few millimeters thick) below the sandstone beds. In the subsurface anhydrite beds up to 20 m thick (volurne 2,150 km^3) and thicker limestone beds are present. In the outcrop area, the anhydrite beds may have been later dissolved so that post-sedimentation slump and collapse structures are present. In the western and northwestern portions of the basin some marine intercalations were observed in bore holes.

The Pedra de Fogo Formation overlies conformably the Piauí Formation, although in the outcrop area some diastemic contacts were observed by Aguiar (1971) and by me during detailed field work for Petrobras. The Pedra de Fogo Formation is conformably overlain by the Motuca Formation and unconformably by the Sambaíba and Pastos Bons formations. Many isolated lacustrine basins may have developed during the deposition of the formation.

Price (1948) assigned an Early Permian age to the Pedra de Fogo Formation based on the fossil amphibian *Prionosuchus* sp. The presence of plant stems Psaronius also indicate a Permian age for

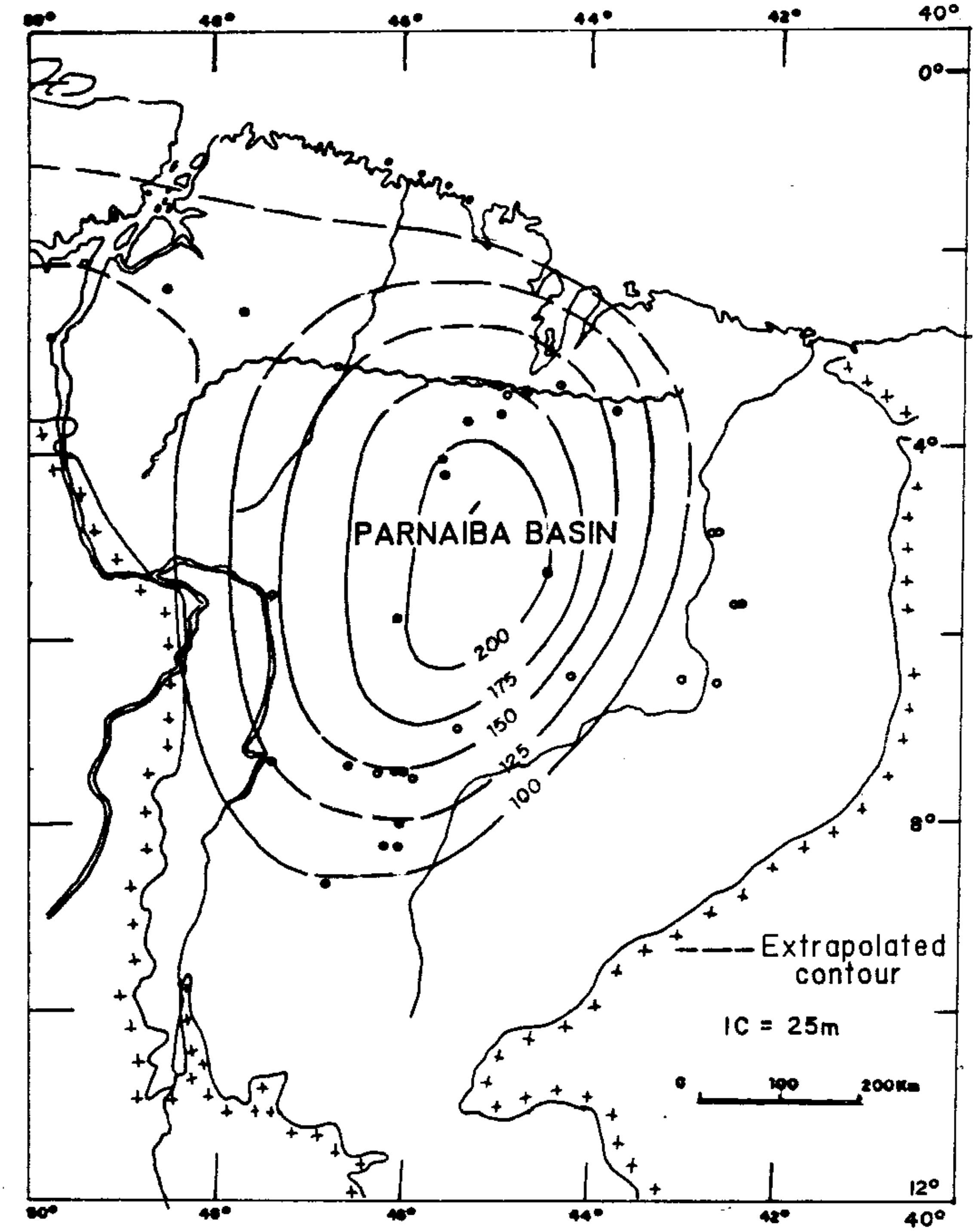


FIGURE 74 - Isopach map of the Pedra de Fogo Formation

the formation. Sporomorphs found in the unit suggest an Early to Middle Permian age.

The Pedra de Fogo Formation correlates with the Nova Olinda Formation and perhaps with the uppermost part of the Itaituba Formation of the Amazonas Basin. In the northwestern portion of the basin, in the subsurface, the unit shows a similarity with the Nova Olinda Formation but without halite beds. The unit also correlates with the Batinga (Boacica Member) and Aracaré formations of the Sergipe-Alagoas Basin, Middle and Upper Efia Nkwanta beds of the Accra Basin (Crow, 1952) and with the Agoula Series of Gabon (Africa). All these formations are composed of heterogeneous sediments with chert interbeds. The sediments may represent restricted basins where the deposition was controlled by similar climatic conditions.

Some investigators have interpreted most of the outcrops of the Pedra de Fogo Formation as deposited in tidal flats or in marine environments, but it is unlikely that tides could develop in such a geographic setting. At that time, few short-lived marine ingressions reached the northwestern area of the Parnaíba Basin. In the outcrop area the unit is envisaged as deposited mainly in lacustrine, fluvial and aeolian environments under arid conditions.

The unit is mainly pelitic associated with continental fossils and with interbeds of limestone, dolomite, pisolitic dolomite, porcelanite (chert) and anhydrite. Pisolitic and oolitic dolomite beds may indicate lacustrine beaches. Flat circular concretions known by the name "bolachas" (flat biscuit) similar to those found in arid playa deposits of the Tertiary of the western United States, are present in the unit. Chert nodules with surface cracks (Magadi lake-type nodules) and tepee deposits are also found in the formation (Della Favera and Uliana, 1979). In central areas of high water table vegetation of *Psaronius* developed, where trunks as much as 50 cm in diameter were found.

Large-scale cross-stratified sandstone beds as well as trough and festoon cross-bedding suggest fluvio-aeolian activity during times of lake regression.

Collapsed and brecciated beds are regarded as a result of solution of anhydrite beds which are common in the subsurface but almost absent in outcrops. Under arid conditions, siliceous epigenetic replacement was common and intense, so fossils were silicified and limestone beds were changed to chert beds. Petrified *Psaronius* trunks are characteristics of the unit. The climate during the deposition of the Pedra de Fogo Formation was arid, suggesting that the area was situated in the belt of trade winds of high evaporation. At the beginning of its deposition, glaciation was taking place in southern Paraná Basin and other areas of the Gondwana continent (Crowell and Frakes, 1972). The subarctic belt may have been located in the northern Paraná Basin.

A problem raised by Meyerhoff and Meyerhoff (1972) was that, in the supposed Gondwana supercontinent huge ice-caps could not have existed because they would have been too far inland to have been reached by adequate amount of moisture. The presence of a large amount of evaporites in the Andean, Solimões, Amazonas, Parnaíba and north African basins suggests that interior seas may have contributed much moisture for building up the Permo-Carboniferous Gondwana ice-caps. They discussed the Permo-Carboniferous glaciation as a synchronous event, although King (1958) and Crowell and Frakes (1975) have shown that its distribution changes in time and space. Crowell (1978) concluded that as the Gondwana supercontinent moved across the south rotational pole, an intermittent glacial imprint followed its course in Permo-Carboniferous time. This behavior can be extended into the past as shown by the presence of glacigenic rocks of Devonian, Silurian and Ordovician age in different places of Gondwana (Caputo and Crowell, in press).

MOTUCA FORMATION

The name Motuca Formation was introduced by Plummer (1948) to designate sandstone and shale beds with limestone and anhydrite interbeds, occurring in the central parts of the basin. According to Aguiar (1971) the formation may be grossly subdivided into three units: the lower member consists of sandstone beds, the middle member mainly of shale, limestone and anhydrite beds and the upper member of sandstone beds. The maximum thickness of the unit is estimated to be over 200 m in the central part of the basin. Figure 75 shows the isopach contour map of the Motuca Formation.

The lower unit consists of friable, subrounded- to rounded-grains, pitted and frosted, in fine- to medium-grained sandstone beds. In some places, the basal unit is replaced by calcareous brick-red siltstone and shale beds, poorly laminated. The middle unit is composed of red siltstone and shale beds with calcite and gypsum lenses and tepee structures. In this unit two anhydrite beds as thick as 30 m are present. The upper part of the unit is more persistent. It is composed of pink, red fine- to medium-grained, cross-stratified sandstone beds with rounded and frosted grains displaying bimodal distribution. Continental sabkha, tepees, adhesion ripples, convex-upward cross-bedding and large-scale cross-bedding, cyclic sedimentation, collapsed structures due to evaporite solution are characteristic features of the formation.

The Motuca Formation rests conformably on the Pedra de Fogo Formation and is conformably overlain by the Sambaíba Formation.

Some conglomerates with chert clasts, from the Pedra do Formation, were found locally at the base of the formation (Lima and Leite, 1978).

The gastropod *Pleurotomaria* sp. and fish forms similar to the Permian fish *Paleoniscus* sp. and *Elonichty* sp. make up the only fossil record known to date from the unit (Mesner and Wooldridge, 1964). *Pleurotomaria sp.* is also found in the Permian of Peru. Based on its stratigraphic position and this poor fossil evidence, a Late Permian age has been assigned to the Motuca Formation, but it probably spans from Late Permian to Earliest Triassic. The unit correlates with Andirá and Fonte Boa formations of the Amazonas and Solimões basins and with the Lower Sekondi Sandstone of Accra. The Lower Sekondi Sandstone is composed of a lower sandy unit, a middle shaly unit and an upper sandy unit with sparse chert fragments.

The Motuca Formation is interpreted as a result of fluviatile, aeolian and lacustrine deposition under arid conditions, with accumulation of continental evaporites. The paleotopography of most of the basin may have been very flat and during wetter periods, huge lakes may have existed, and during dryer periods, fluvial and aeolian progradation towards the center of the lake may have taken place. Some geologists have regarded the presence of two very persistent anhydrite beds as a result of marine deposition; however, the erosion and solution of older gypsum beds from the Pedra de Fogo Formation, in the northwestern side of the basin may have furnished sulphate

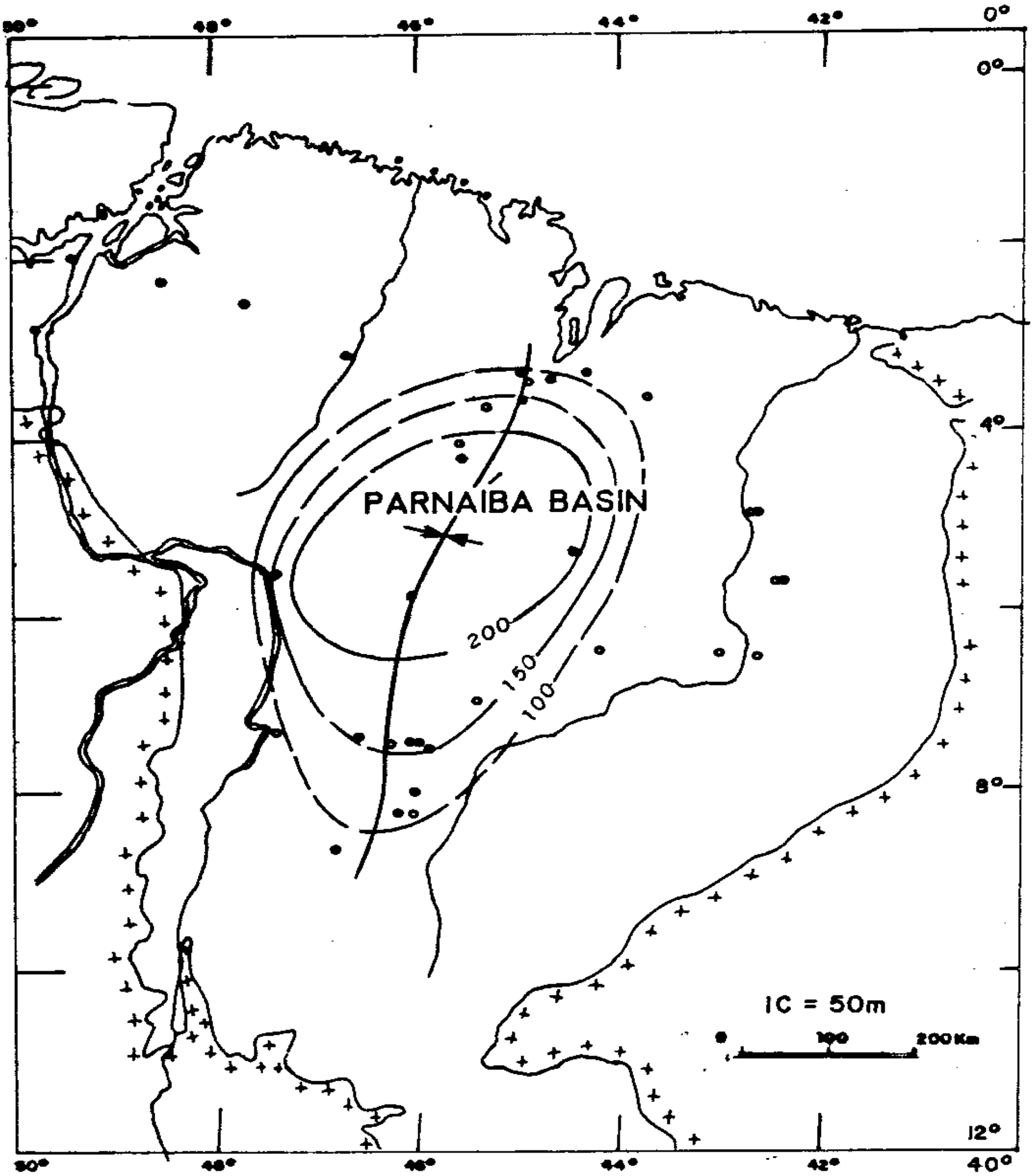


FIGURE 75 - Isopach map of the Motuca Formation

which was redeposited in the central parts of the Motuca Basin in a continental environment.

A similar present-day tectonic and climatic model may be Lake Chad which shrinks three times during the present dryer seasons, and in the past, during dryer epochs, dunes encroached a great part of it and now the sand dunes are covered with lacustrine sediments and water. The Chad area is presently undergoing subsidence related to the uplift of the eastern African rift valley (Burke, 1976).

In Late Permian and Early Triassic times, a major worldwide regression took place on earth, when huge areas emerged. At these times, no significant glacial rocks have been recorded anywhere, so the regression should perhaps be related to tectonic causes, as for example, the final collision of Gondwana with Laurasia making up the supercontinent Pangea II. At this time, the equatorial and tropical climatic belts may have expanded considerably in comparison with times of huge polar ice-caps. The combination of broadly exposed areas, the formation of Pangea, and wide warm climatic belts, may have resulted in the generation of widespread deserts in many areas of this supercontinent.

SAMBAÍBA FORMATION

The name Sambaíba Formation was first applied in 1948 by Plummer to describe a sandstone interval which forms tableland near

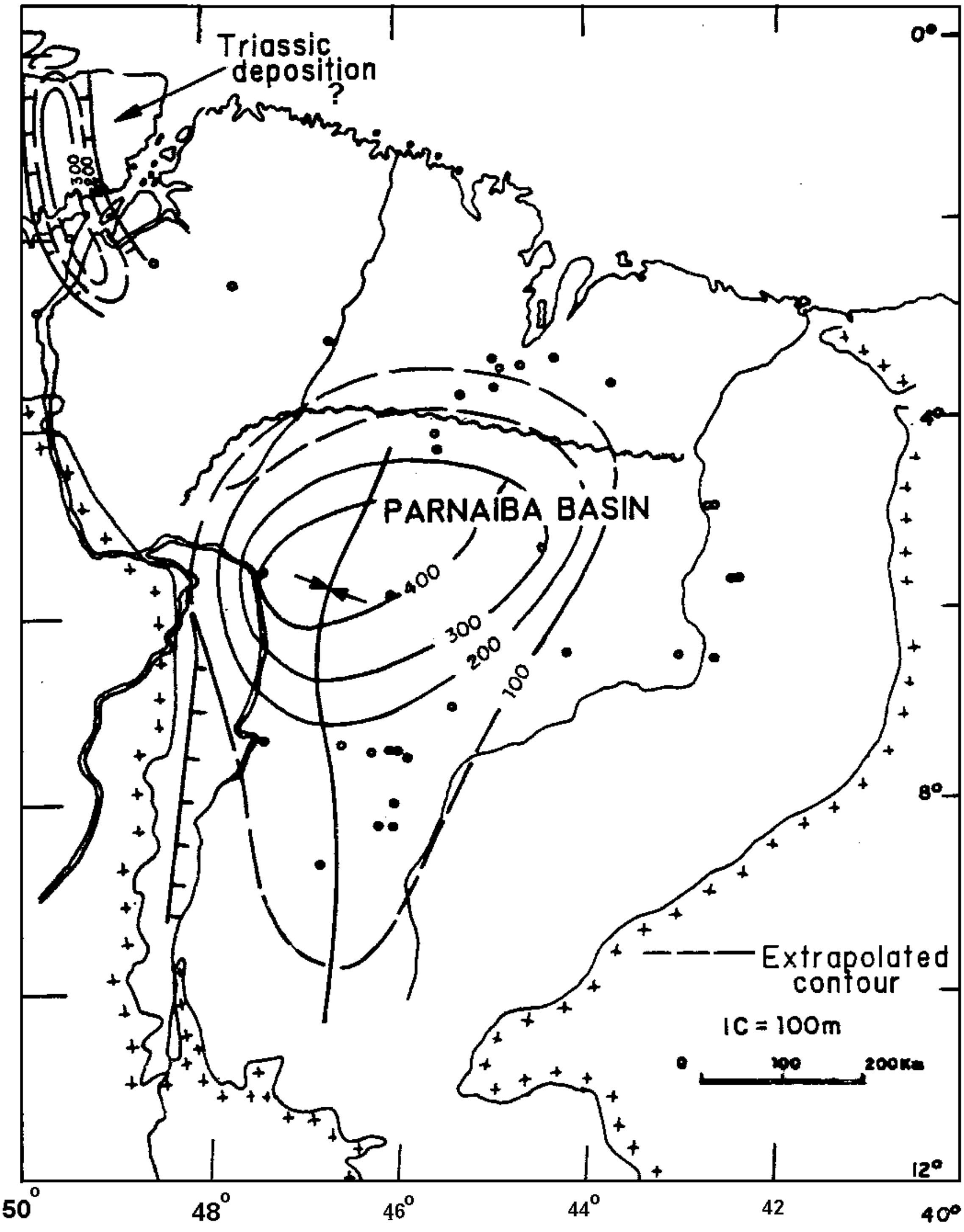


FIGURE 76 - Isopach map of the Sambaíba Formation

the locality of the same name. Figure 76 shows the isopach contour map of the Sambaíba Formation. The section, as much as 400 m thick, consists of yellow, pink, cream-colored, red and white, cross-stratified fine-to medium grained and bimodal sandstone beds with large number of scattered frosted subrounded, rounded and spherical grains. Large-scale cross-bedding is common throughout the unit.

In the central parts of the basin, in bore holes, the Sambaíba Formation rests conformably on the Motuca Formation where brick red siltstone, shale and very fine-grained sandstone beds of the Motuca Formation change to white and pink fine- to medium-grained sandstone beds of the overlying Sambaíba Formation.

The formation also rests unconformably on the Pedra de Fogo and Piauí formations, suggesting a previous period of erosion before deposition and a larger area of deposition than that of the Motuca Formation in the southern part of the basin (Kegel, 1956).

The upper contact of the Sambaíba Formation is conformably overlain by basalts and intertrap sediments of the Mosquito Formation. In some areas the unit is also unconformably overlain by Cretaceous Corda, Urucuia and Itapecuru formations.

The Sambaíba Formation up to date has not yielded fossils, so that its age is based on stratigraphic position, since it is located between the Earliest Triassic (Upper Motuca Formation) and Middle Triassic (age of Lower Mosquito Basalt Formation). The unit has no correlation with units of the Amazonas Basin. The Sambaíba Formation is partly correlative with the Rosário do Sul Formation (Middle to Upper Triassic age) of the Paraná Basin and it may correlate with the upper Secondi beds of the Accra Basin in Ghana.

The Sambaíba Formation represents deposition under aeolic and fluvial conditions. It is characterized by repetition of cosets of large-scale cross-strata and thin horizontal deposits representing both dune and interdune environments. Wadi (river) deposits consist of medium-grained sandstone beds which interfinger with the aeolian deposits. The climate was dry, hot, and desertic and probably located in the zone of trade winds belt of high evaporation.

MOSQUITO FORMATION

The Mosquito Formation was introduced by Aguiar (1971) to define a section composed of intercalated basalts and sediments. This name was first used by Northfleet and Neves (1967) in a Petrobras internal report. It was described and subdivided into five members from bottom to top by Northfleet and Neves (1966): Lower Basalt, Macapá, Middle Basalt, Tingui and Upper Basalt.

Its maximum thickness is estimated to be over 200 m in the western part of the basin. Figure 77 shows the isopach contour map of the Mosquito Formation.

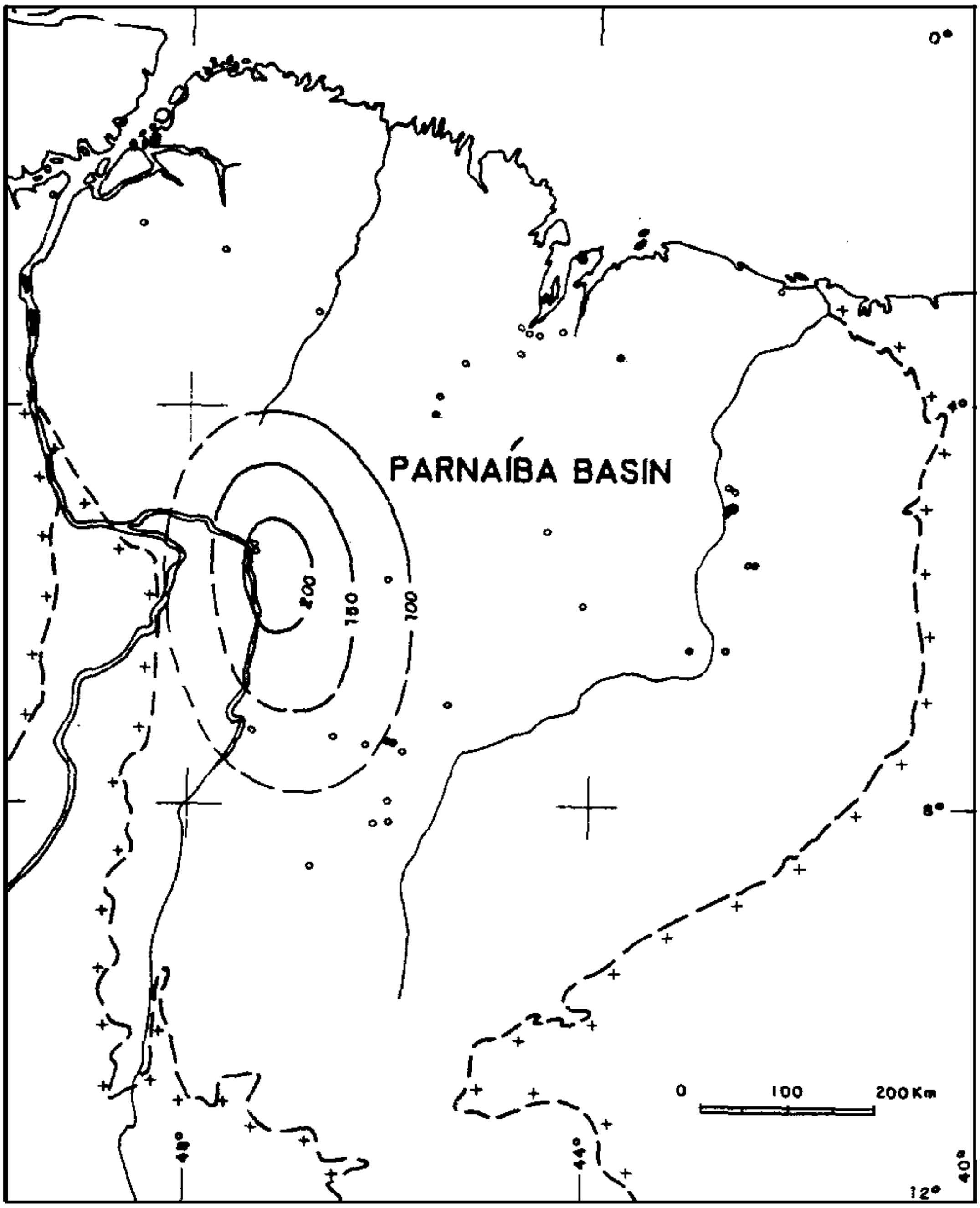


Figure 77 Isopach map of the Mosquito Basalt - Parnaíba Basin.

The basalts are dark gray to red brown (weathered), massive aphanitic, amygdaloidal and composed of mainly labradorite or andesine, augite and opaque minerals. The Macapá Member, up to 30 m thick, consists of pink to white, fine- to medium-grained, cross-laminated or parallel-stratified, well sorted sandstone beds. The upper part is composed of interbeds of fine-grained, pink silicified sandstone and pink and red siltstone with pisolitic chert lenses. The Tingui Member, up to 15 m thick, is composed of pink siltstone beds with chert interbeds. No fossils were found in the formation.

The intertrap sediments indicate aeolian and lacustrine environments of sedimentation during the accumulation of the Mosquito Formation. This formation is restricted to the central parts of the basin where subsidence and extrusion took place. Arid and hot climate was present during the deposition of the Mosquito Formation.

Igneous basic rocks were formed at different times called stages. The first stage corresponds to basalts as old as 215 m.y. (Middle Triassic); the second or main stage corresponds to basalts extruded from ± 180 to ± 150 m.y. ago and the last stage is characterized by diabase rocks intruded between ± 150 to ± 120 ± 10 m.y. ago (Caldasso and Hana, 1978). The oldest basalts correlate with diabase dikes occurring in the Amapá State area. The main stage basalts correlate with the Penatecaua Diabase of the Amazonas Basin and to the beginning of the basic igneous activity at the Paraná Basin. The late stage, consisting only of diabase, is correlated with the main flows of the Paraná Basin (Serra Geral Formation).

Thomaz Filho and others (1974) recognized two magmatic cycles in the Amazonas Basin, while Caldasso and Hana (1978) recognized three stages in the Parnaíba Basin. According to Caldasso and Hana (1978) the first (early) stage would be related to the opening of the North Atlantic Ocean, the main stage would be related to a transition between the opening of the North and South Atlantic Oceans (Equatorial Atlantic Ocean) and the last stage would be related to the opening of the South Atlantic Ocean. The early stage is not registered in the Paraná Basin and the late stage is not known in the Amazonas Basin. Therefore, the basic igneous activity lasted longer in the Parnaíba Basin than in other Brazilian intracratonic basins.

PASTOS BONS FORMATION

The name Pastos Bons was first applied in 1914 by Lisboa to define a variegated shale and sandstone section which is present at the town of Pastos Bons. A great deal of re-definition and re-description of the Pastos Bons Formation has taken place since then. Melo and Prade (1968), Aguiar (1971) and Lima and Leite (1978) provided comprehensive studies of the unit. Figure 78 shows the isopach contour map of the Pastos Bons Formation.

The sequence consists of about 80 m of white and greenish, yellowish and whitish, fine- to medium-grained, subrounded argillaceous, generally parallel stratified sandstone beds with local limestone lenses. Above the sandstone beds, gray to green sandy mudstone beds with scattered medium- sized sand grains are present. The mudstones are interbedded with green and white fine- to medium-grained sandstone beds. The upper section consists of red to pink fine-grained sandstone to siltstone beds with shale interbeds; black and fissile shales occur locally.

The Pastos Bons Formation overlies unconf'ormably the Poti, Piauí, Pedra de Fogo and Motuca formations (Aguiar, 1971) and is conformably overlain by the Corda Formation. The unconformity surface beneath the Pastos Bons Formation is important because it indicates that a great part of the basin margins were exposed when its deposition began. This stratigraphic relationship also shows that the depositional site changed in relation to the underlying unit. The regions flooded with basalts may have become high for some time, and younger Pastos Bons deposition started to take place in the basin flanks which were free from basaltic lavas.

The unit yielded fish remains and ostracodes. The fossil fish genus

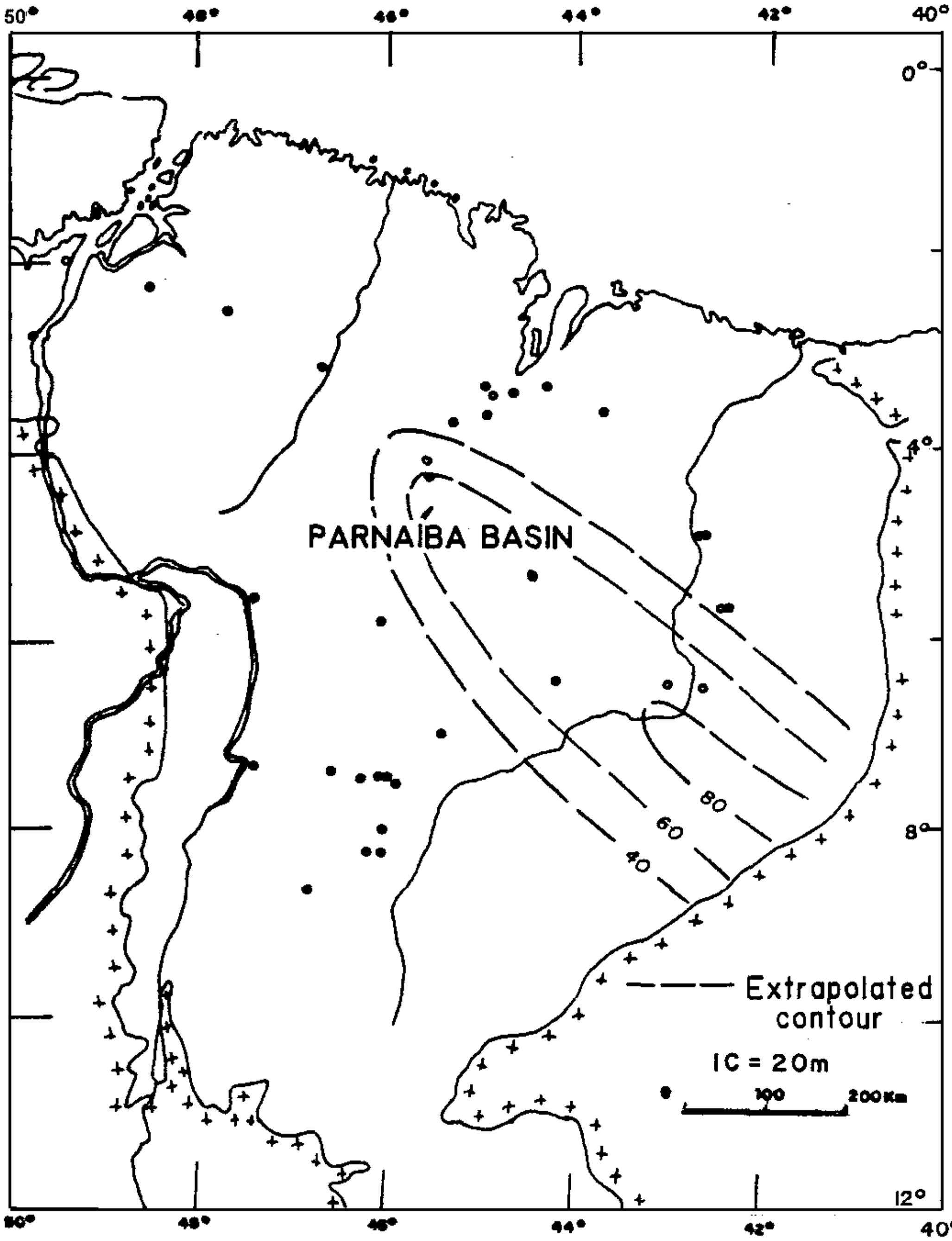


FIGURE 78 - Isopach map of the Pastos Bons Formation

Semionotus .and the species *Lepidotus piauhiensis* were identified by Roxo and Lofgren (1936, in Aguiar, 1971). The fish *Macrossemidae* and *Pleuropholidae* suggest a Middle to Late Jurassic age for the unit (Santos, 1974). The ostracode *Macrolimnadiopsis* sp. studied by Pinto and Purper (1974) indicates a Late Jurassic to Early Cretaceous age. Based on this poor fossil record a Late Jurassic age is attributed to it.

The Pastos Bons Formation was deposited in lacustrine and fluvial environments as a result of a drainage reorganization in northeast Brazil. The western, northwestern (Tocantins arch) and northern (Ferrer arch) sides of the basin were being uplifted, so that the drainage system was toward the east and southeast areas, outside the Parnaíba Basin. The drainage system was dammed toward the Amazonas Basin as can be observed in the isopach contour map (Figure 78) of the Pastos Bons Formation.

In depressions, inside and outside the Parnaíba Basin, small to large lakes may have originated along the rivers. River and lacustrine sediments were deposited in small basins over crystalline rocks as well as over Paleozoic sediment remnants occurring between the reduced Parnaíba Basin and the present coastal area. The drainage system may have continued toward Africa following some downfaulted areas somewhere in states of Espírito Santo or Rio de Janeiro. The presence of limestone beds in the Pastos Bons Formation suggests a warm climate during its deposition.

CORDA FORMATION

The term Corda Formation was originally used by Lisboa (1914) to describe red sandstone beds intercalated between basalts. Its maximum thickness is estimated to be over 70 m in the southeast side of the basin. Figure 79 shows the isopach contour map of Pastos Bons and Corda formations. This unit has occupied several stratigraphic positions, but Melo and Prade (1968) were able to place it in a proper position conformably over the Pastos Bons Formation. This stratigraphic relationship was followed by Aguiar (1971), Lima and Leite (1978) and by me.

The unit overlies conformably the Pastos Bons Formation and disconformably the Mosquito, Sambaíba, Motuca, Pedra de Fogo, Piauí, Poti, Longá and Cabeças formations. The depositional site was shifted toward the southeast during its deposition. According to Aguiar (1971) the unit is unconformably overlain by the Sardinha Basalt and younger formations.

Lima and Leite (1978) reported the presence of the conchostracans, *Lioestheria* sp. and *Macrolimnadiopsis* sp. and ostracods of the *Candona* genus. *Macrolimnadiopsis* was also found in the underlying Pastos Bons Formation (Pinto and Pupper, 1974). Lima and Leite (1978) dated as Late Jurassic on the basis of fish and conchostracans

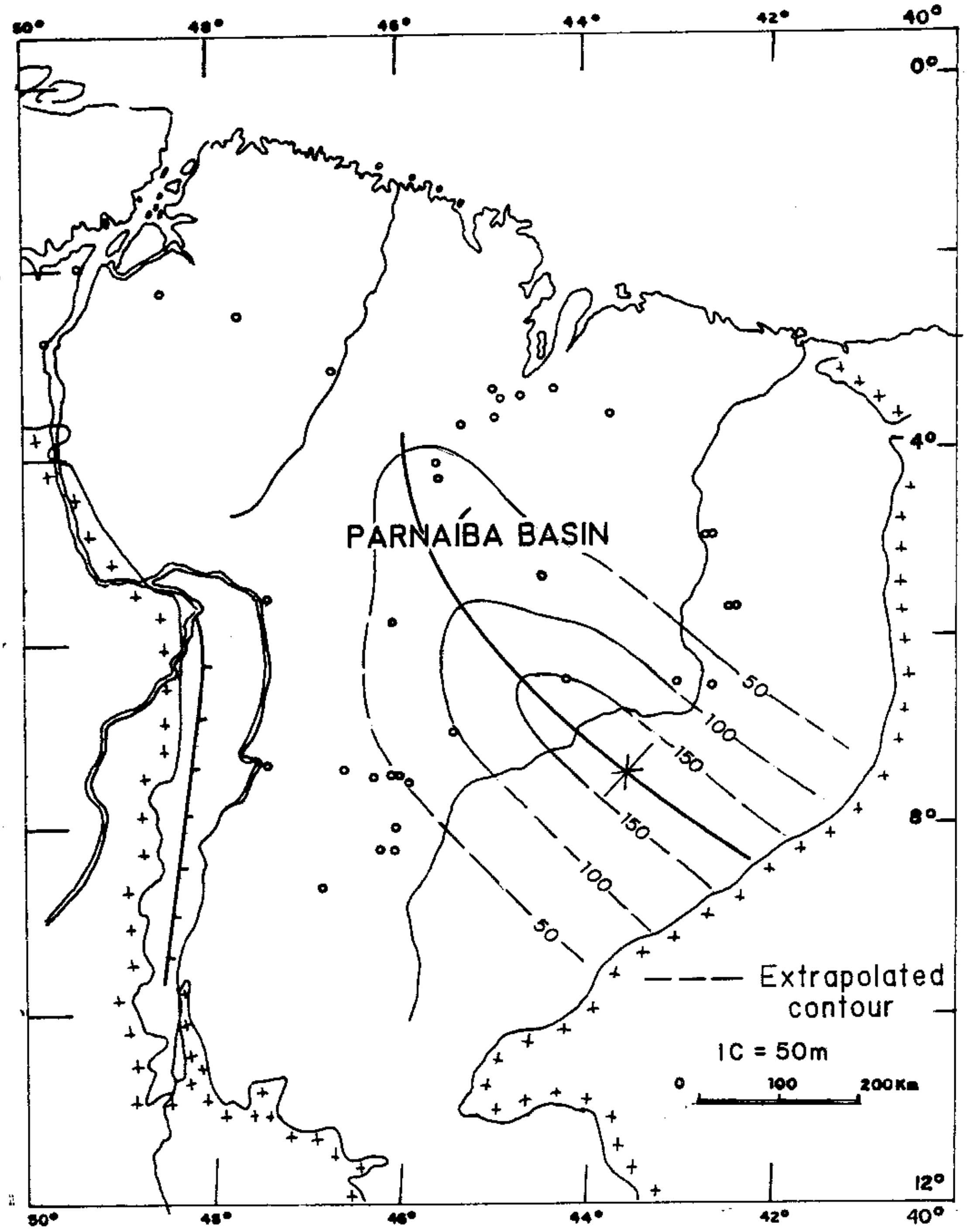


FIGURE 79 - Isopach map of Jurassic beds (Corda and Pastos Bons Formations)

fossils. Based on this poor evidence, a Late Jurassic age is attributed to the unit.

The Corda Formation may correlate with the Sergi Formation of the Recôncavo Basins as well as with the Serraria Formation of the Sergipe-Alagoas Basin. Its environment of deposition may have been alluvial fan, alluvial plain and desertic. Before the deposition of the Corda Formation a large part of Paleozoic strata were exposed. At this time, in the Parnaíba Basin and in the northeastern part of Brazil, a drainage system was established directed east and southeastward, and reaching the Araripe, Tucano and Recôncavo basins.

The climate may have been semi-arid as interpreted by the presence of red beds, chert beds, aeolian and alluvial fan deposits. Although the climate was arid to semi-arid, the environment of deposition of the Corda Formation was not conducive to the deposition and preservation of evaporites.

SARDINHA FORMATION

Aguiar (1971) formalized the name Sardinha Formation to describe Cretaceous basalts occurring at the village of Sardinha. The formation is about 20 m thick and its areal extent is very limited. He correlated the Sardinha Basalt with other basalts occurring southeast of the town of Natal and the town of Lizarda. This last basalt is as thick as 50 m.

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According to Aguiar (1971) the basalt is black and amygdaloidal but it is normally altered to purplish gray and brown colors. Due to its high degree of alteration, no geochronologic dating has been possible up to the present, but a nearly fresh basalt from Lizarda was tentatively dated as Early Cretaceous (Cordani, 1967), that is, the same age as that of the youngest intrusive diabase rocks found in the Paraná Basin. The stratigraphic position of the Sardinha Formation is controversial. According to Caldasso and Hana (1978) it may be the same as the Mosquito Basalt, but they recognized the presence of diabase dikes in the Corda Formation which overlies the Mosquito basalts. These diabase dikes could produce lava over the Corda Formation.

Lima and Leite (1978) considered that the Sardinha Basalt is underlying the Grajaú Formation but they also admitted that the Sardinha Formation is overlying (!) the Grajau Formation. Here, the Sardinha Formation is considered as overlying the Corda Formation and underlying the Grajaú and Codó formations, because basalts or diabase were never found affecting these two formations. After the Sardinha magmatism the tectonism in the basin changed. The northern and eastern coastal areas began to subside, then a pre-Late Aptian unconformity was developed so that the beds are now gently dipping northward. The dips of the units beneath the unconformity surface are still directed toward the basin center (Figure 56), but the dips of the units above the unconformity are directed toward the coast.

This stratigraphic-structural relationship is important because it reflects the same behavior as that which occurred in the Amazonas Basin, where the uplifted Marajó arch and coastal areas subsided in the Late Cretaceous. In the Amazonas Basin the subsidence of the coast may have begun in Late Cretaceous time. Rezende and Pamplona (1970) concluded that the Ferrer arch started to rise after the Cretaceous magmatic activity (120 ± 10 m.y. ago) but the inference has no support. After the magmatic phase, the Ferrer arch area, together with the coastal region, started to subside when the sea ingressed from the Equatorial Atlantic Ocean and encroached downfaulted coastal areas in Late Aptian to Late Albian times.

AREADO FORMATION

The term Areado sandstone was first applied by Rimann (1917 in Lima and Leite, 1978) to describe red sandstones occurring in the State of Minas Gerais. The unit was described by Lima and Leite (1978) and is discussed here due to its paleotectonic importance.

In the Parnaíba Basin, the unit is about 70 m thick and is restricted to its southern corner. Its major development occurs southward in the São Francisco Basin with a distribution of about 1,000 km in length in a N-S direction and 500 km in an E-W direction, with a maximum thickness of about 230 m.

In the Parnaíba Basin, the Areado Formation at the base, is made up of conglomerate and red conglomeratic sandstone beds with a red clayey matrix. The clasts are composed of quartz (up to 6 cm across), quartzite (up to 35 cm across) chert (up to 20 cm across). Many quartz pebbles are true ventifacts with several truncated surfaces.

The middle part of the formation consists of calcareous, micaceous, laminated, reddish brown siltstone beds with variegated fine- to medium-grained calciferous sandstone beds. In the upper section micaceous, pink, fine- to medium-grained, parallel- stratified and cross-laminated sandstone beds predominate.

Basalts over 1,500 m in thickness laid down in the Paraná Basin, and extensive volcanic activity in the Rio de Janeiro and Espirito Santo state regions, may have changed the previous southeastward direction of the drainage system, partly damming the drainage system for an interval, so that deposition took place up-river.

The formation overlies disconformably the Piauí and older formations and is conformably overlain by the Urucuia Formation. Outside the basin boundaries, in the southern side of the São

Francisco Basin, the unit is overlain by volcanoclastic rocks.

Lima and Leite (1978) listed the conchostracan fossils *Liostheria cardoense* Cardoso, *Pseudograpta brauni* Cardoso, *Bairdestheria* and *Paleolimnadiopsis* and sporomorphs *Inaperturo-pollenites* and *Verrutriletes*, indicating an Aptian-Albian age. According to Barcelos and Suguio (1980) the lower age limit of the unit is Barremian in the São Francisco Basin, although those investigators did not present any substantial evidence for the age assignment.

The Areado Formation is partly the time-equivalent of the Bahía Series of the Recôncavo Basin. The formation may have been deposited in alluvial fans, alluvial plain, lacustrine and desertic environments under semi-arid conditions. It may include deposits of the old São Francisco river system which drained towards the Parnaíba Basin and northeast Brazil.

The deposition of the Areado Formation, over older basement rocks of the São Francisco Basin, indicates that subsidence was taking place in that area and coastal regions while the central Parnaíba area was being uplifted. Regions of maximum upwarping, as the Recôncavo basin collapsed in Jurassic time, producing large interior lakes such as those in eastern Africa. Sedimentation took place in the rift zone and in back basins of the cratonic interior in the São Francisco Basin. The São Francisco Basin is an intracratonic basin active in Cretaceous time as a result of subsidence marginal to a long-lasted coastal uplift.

The same structural behavior also took place in the Amazonas Basin where the coastal area was being raised while the western interior basin was being downwarped. In the Amazonas and Parnaíba basins, the subsidence was observed far inland in flat regions which had been almost at sea level since the Carboniferous time, while in the São Francisco Basin, where some ancient relief still existed, the area was able to retain sediments, only in the Cretaceous time.

Until the pre-Aptian time, northeastern Brazil and central western Africa, had land continuity as is indicated by similar rock and fossil content between the Recôncavo and Sergipe-Alagoas Basins and all African rifted marginal basins from Senegal to Angola.

Meyerhoff and Meyerhoff (1972) contested the existence of land continuity between South America and Africa up to the Early Cretaceous time. They argued that the presence of identical species of ostracodes in Brazil and Gabon could be due to ostracodes transportation across both sides of the Atlantic Ocean in migrating birds' feet. It is interesting to mention that the fresh water Early Cretaceous Congo Basin, 150 km apart, the ostracodes are completely different from those of the Gabon and Reconcavo basins (Franks and Nair, 1973).

For a long time, the Brazilian Early Cretaceous stratigraphic zoning is based on ostracodes, consequently, these fossils are much better known in Brazil than in the Gabon Basin. Over 2,000 wells were drilled in the Recôncavo and Sergipe-Alagoas basins, so that up to 1967 Petrobras paleontologists were able to determine over 150 species in Brazil, whereas only over 30 species were determined (Grekoff and Krommelbein, 1967) in the Gabon Basin. Further studies certainly will increase the number of common ostracodes species in both sides of the Atlantic Ocean.

On the other hand, the presence of the non-marine Early Cretaceous crocodilian *Sarchosuchus* in both Brazil (Bahia State) and Niger provides additional evidence of the land continuity between South America and Africa (Buffetaut and Taquet, 1977). It seems that transportation of giant crocodilian *Sarchosuchus* eggs in migrating birds' feet is quite unlikely. The arguments presented by Meyerhoff and Meyerhoff (1972) are untenable.

GRAJAÚ FORMATION

The name Grajaú sandstone was first used by Lisboa (1914) to describe a Cretaceous section supposed to occur under the Codó Formation. The stratigraphic position of the unit, up to 100 m thick, was not well known, when Carneiro (1974), based on photointerpretation and field work, demonstrated that the Grajaú and Codó formations interfinger with each other and are synchronous.

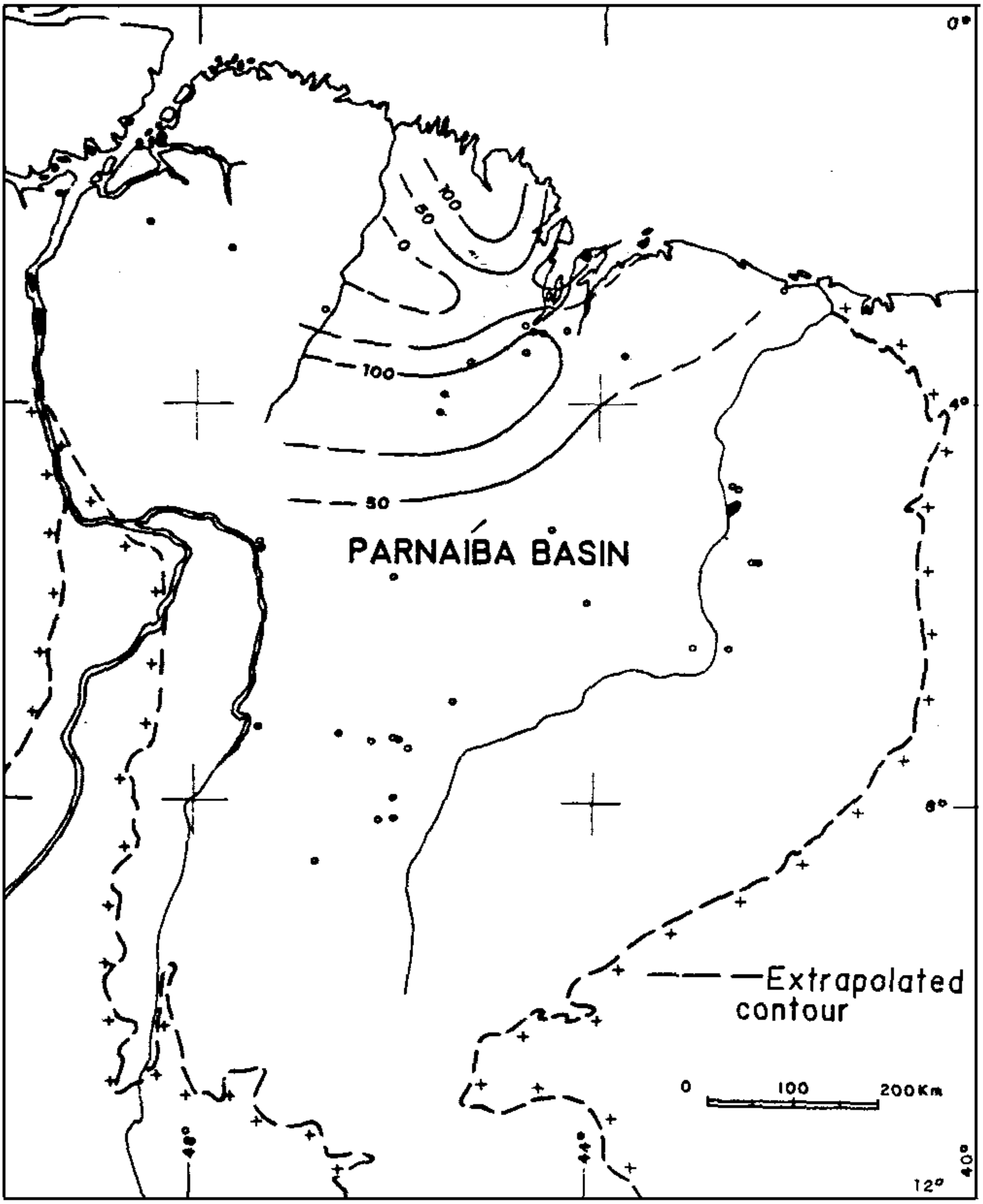

Figure 80 Isopach map of the Grajaú Formation - Parnaíba Basin.

Lima and Leite (1978) agreed with Carneiro's (1974) observations. Figure 80 shows the isopach contour map of the Grajaú Formation.

According to Lima and Leite (1978) in the lower part, the Grajaú Formation normally consists of a polymictic conglomerate composed of quartz, chert, basalt, and sandstone pebbles and cobbles of subjacent rocks. The bulk of the unit consists of white, cream-colored, pink and rarely purple, feldspathic, fine-grained to conglomeratic sandstone beds that contain low- to high-angle cross-stratified or parallel-stratified units. The grains are frosted or bright, clean, and subangular to rounded. Some interbeds up to 2 m thick of dark red, brown, or purple, partly silicified mudstone beds are present. The top of the section is silicified in many places (Lima and Leite, 1978).

The Grajaú Formation overlies unconformably the Corda, Mosquito and Sambaíba formations. Its upper contact is unconformable with the Itapicuru Formation.

No fossils were found in the Grajaú beds. However, due to its stratigraphic relationship with the Codó Formation, a Late Aptian to Late Albian age is attributed to it. The Grajaú Formation represents fan delta and fan-delta front environments developed around lagoons and lakes. The climatic conditions are the same as those of the Codó Formation.

CODÓ FORMATION

The term Codó was first used by Lisboa (1914) to describe shale and limestone interbeds occurring in the Itapecuru Valley. Carneiro (1974) demonstrated that the Grajaú and Codó formations interfinger with each other and that they have the same age. Figure 81 shows the isopach contour map of the Codó Formation. The Codó Formation, up to 400 m thick, was subdivided into 3 units in subsurface work (Rezende and Pamplona, 1974). They described the section as follows:

The lower unit consists of a basal conglomerate followed by black and greenish-gray laminated bituminous shale beds with thin limestone interbeds overlain by a gypsum bed as thick as 10 m. The middle section is composed of a polymictic conglomerate overlain by ostracoidal shales at the base and marls with ostracodes, gastropods and lamellibranchs at the top. The upper unit comprises calcareous, micaceous, gray sandstone and siltstone interbeds with plant remains, ostracodes and gastropods. At the surface, the lower or the middle units may pinch out towards the west and south sides of the basin (Lima and Leite, 1978).

The Codó Formation overlies disconformably older sedimentary rocks and the basement in the northern part of the basin where

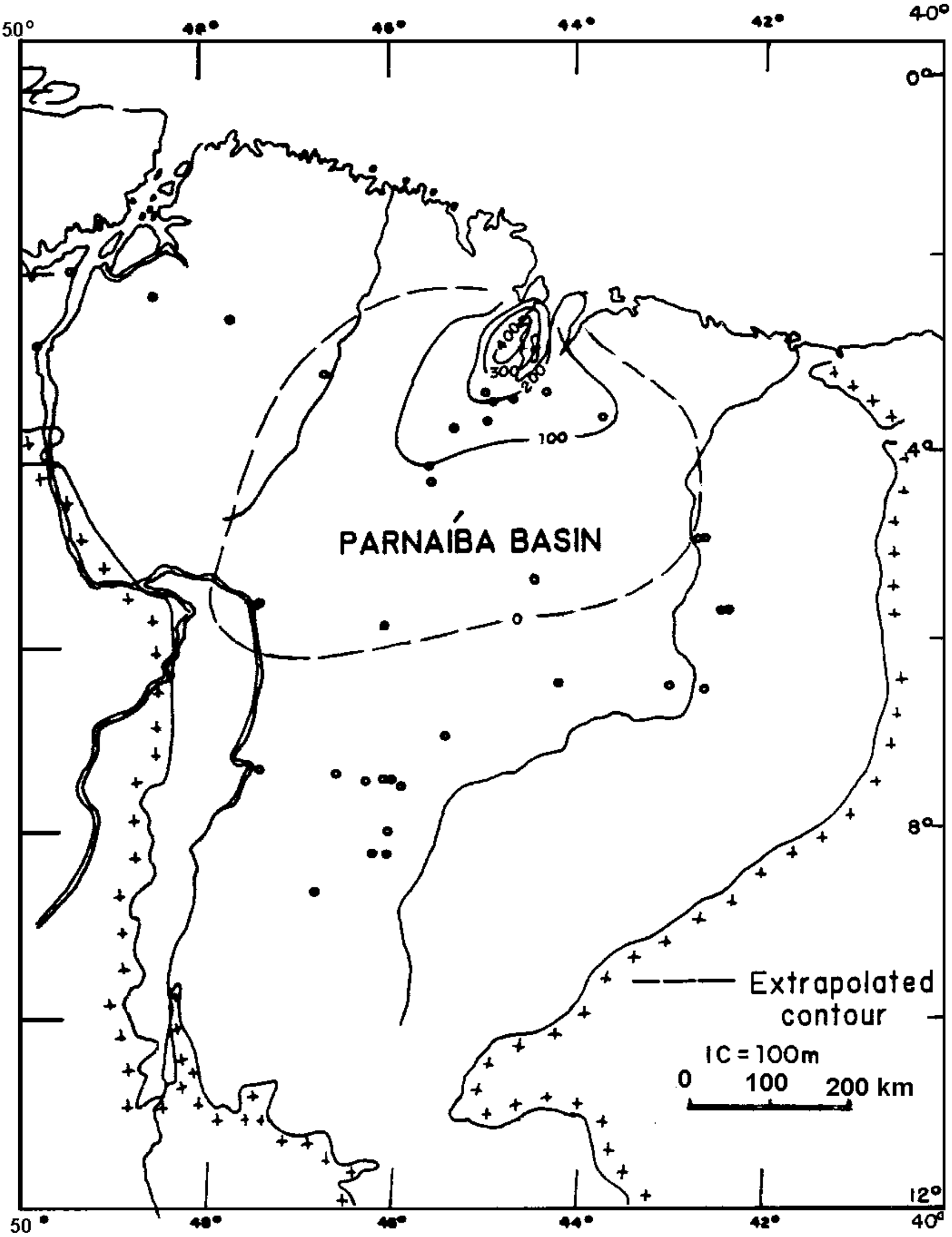

FIGURE 81 - Isopach map of the Codó Formation

basement (Ferrer arch) was exposed. The formation dips towards the coast, making a gentle angular unconformity with the pre-Codó rocks which are dipping in the opposite direction. The Codó Formation interfingers with the Grajaú Formation (Carneiro, 1974) and both units are conformably overlain by the Itapecuru Formation.

Sporomorphs and fresh water ostracodes *Cypridea* sp.; *Darwin-ulidae* sp.; and *Paraschileridea* sp. and marine gastropods *Turritella* sp. and *Nerinea* sp. are mentioned by Lima and others (1978) and *Anomia* sp., *Arca* sp., *Corbula* sp. and *Turritella* sp. by Mesner and Wooldridge (1964).

On the basis of the fossil content Lima (1981, written communication), assigned to the Codó Formation a Late Aptian to Late Albian age. The unit correlates on the basis of palynomorphs with the Barro Duro and Arpoador formations of the rifted marginal Barreirinha Basin and with the Santana Formation of the Araripe plateau, located as far as 400 km eastward from the eastern Codó outcrop edge. It also correlates with the Riachuelo Formation of the rifted marginal Sergipe-Alagoas Basin, located on the eastern Brazilian coast.

The lower part of the Codó Formation was laid down in a lagoonal shallow water environment. A narrow connection with the sea ended with a regression which caused the precipitation of evaporite beds. The middle part records a new transgression ending with lagoonal and lacustrine brackish water environment rich in ostracodes. The upper part of the section may have been deposited in deltaic lobes in a lacustrine-lagoonal environment.

During the deposition of the Codó Formation, a thick sequence of evaporites was also being deposited in rifted marginal Atlantic basins. At that time, the region of the Parnaíba Basin continued to be located in the trade winds belt of high evaporation.
The presence of the first marine sediments along the eastern South American coast indicates that the Proto-Atlantic Ocean had developed between South America and Africa before Late Aptian times.

ITAPECURU FORMATION

The name Itapecuru Formation was first applied by Campbell (1949) to describe variegated sandstone, siltstone and shale beds of Cretaceous age. Its maximum thickness is estimated to be over 2,000 m in the São Luís Basin. Figure 82 shows the isopach contour map of the Itapecuru Formation.

The formation, described by Lima and Leite (1978), consists of variegated, mainly red, pink and white cross- and parallel-stratified, fine- to coarse-grained sandstone beds with conglomerate, red and green siltstone and mudstone interbeds. Lateritic and bauxitic crusts are present at the top of the unit in the

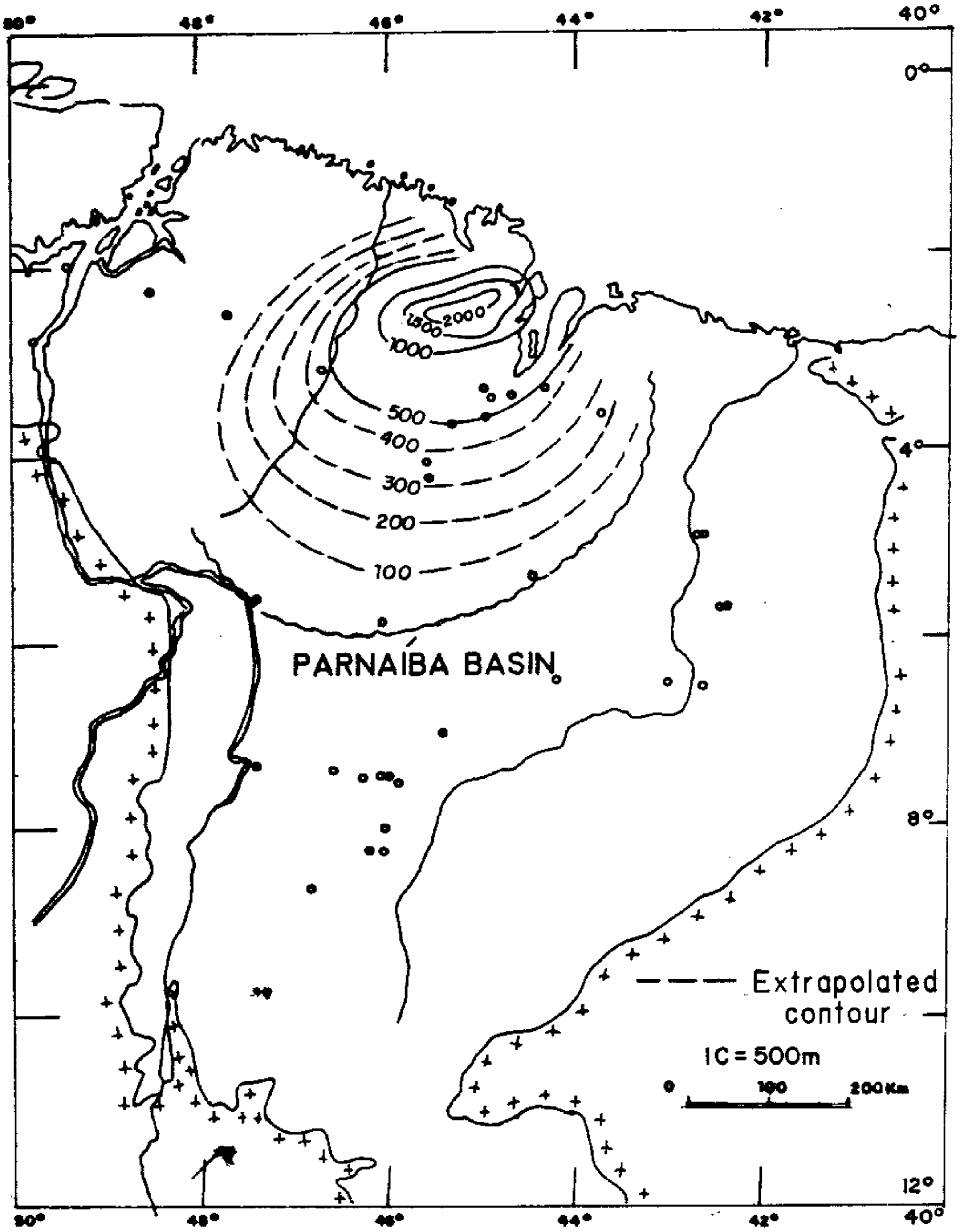

FIGURE 82 - Isopach map of the Itapecuru Formation

northwestern part of the basin. Its overall coarseness increases westward and easternward while its thickness increases gradually northward from the middle of the basin to the coast.

The isopach map shows that the formation overlies unconformably the Codó Formation, but beyond the limits of the Codó Formation it overlies unconformably Mesozoic, Paleozoic and Precambrian rocks. It pinches out toward the Tocantins arch where it is overlain by Tertiary rocks. In the coastal area, the Itapecuru is overlain by the Cretaceous Alcântara Formation and also by unnamed Tertiary sediments.

In most parts of the basin, the unit is located at the top of of the sedimentary pile. No fossils were found in the Itapecuru strata, but in the coastal area, the overlying Alcântara Formation limestone yielded fossils as old as Coniacian to Maestrichtian (Aguiar, 1971). In this area the Itapecuru Formation may have been laid down from latest Albian to Turonian times. It perhaps interfingers with part of the Alcântara Formation, so that in some areas it may be somewhat younger than Turonian age.

I found that the Itapecuru Formation correlates with part of the Preguiças and Bonfim Formations of the rifted marginal Barreirinha Basin and with part of the Alter do Chão Formation of the Amazonas Basin. It is equivalent to the Urucuia Formation occurring in the south part of the basin.

I consider that the Itapecuru Formation may have been deposited in an alluvial plain system in which the fine and medium sandstone beds represent channel environments and the shale and siltstone represent flood plain environments. The absence of evaporite minerals in the red bed shale may indicate a predominant moist tropical climate.

During the deposition of the Itapecuru Formation, the coastal area of the São Luís Basin was downfaulted. Almost all sedimentary fill is made up of the Itapecuru Formation in the rifted marginal São Luis Basin which was controlled by major normal fault zones bounding it. The bounding faults of the rifted Marajó Basin may have extended towards the Tocantins River, forming narrow grabens in the western part of the Parnaíba Basin, along the old Araguaia metamorphic belt weak zones. Along the Tocantins River, fault displacements range between 200 and 300 m and are affecting the Codó and Itapecuru formations, close to the Marabá region.

URUCUIA FORMATION

The Urucuia Formation was first described by Hartt (1870) and Dodt (1872), but it was first named by De Oliveira (in De Oliveira and Leonardos, 1943) to describe cliff-forming sandstone beds occurring along the watershed between the Tocantins and Parnaíba rivers. Figure 83 shows the isopach contour map of Cretaceous and

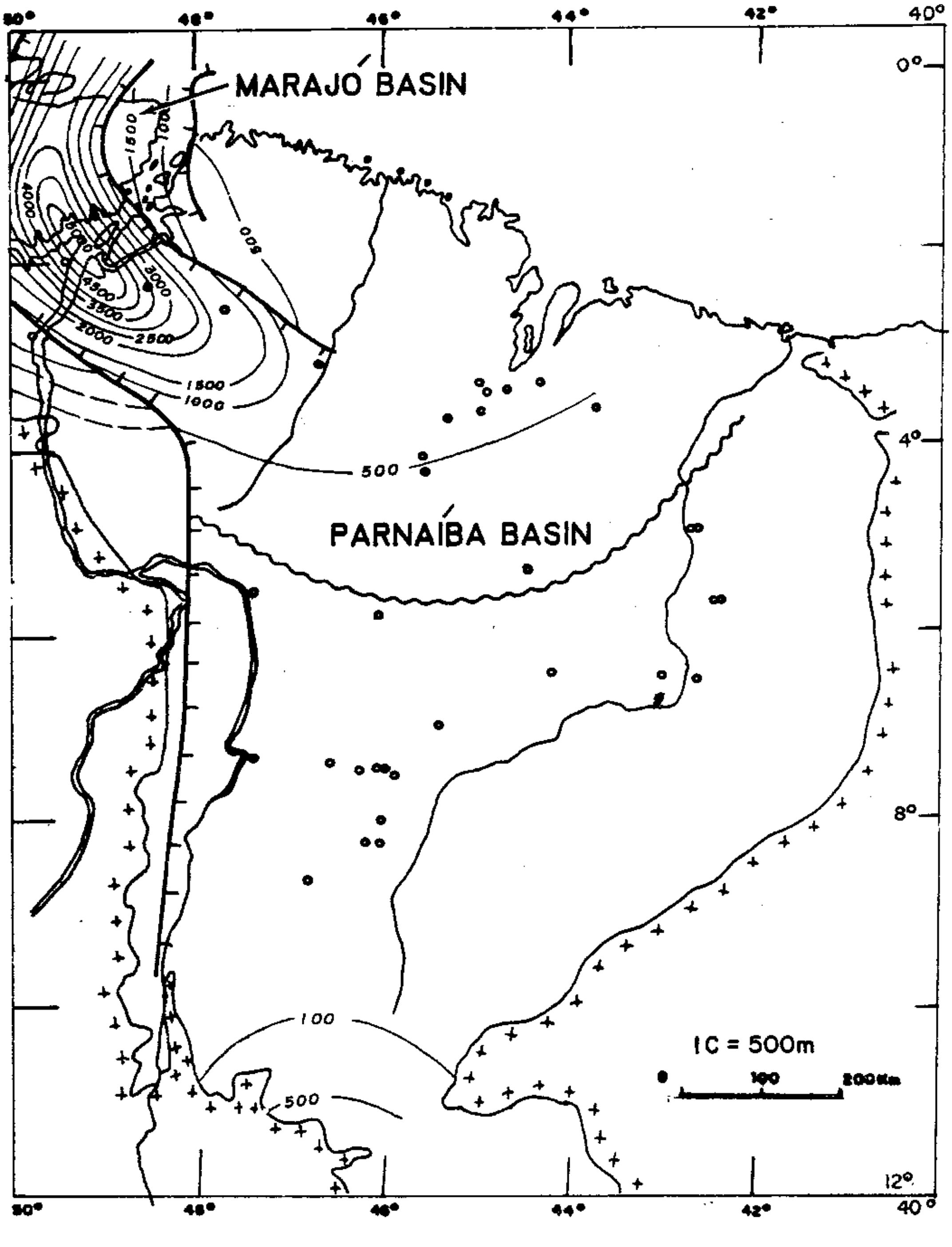

FIGURE 83 - Isopach map of Cretaceous and Tertiary beds (Grajaú, Codó, Itapecuru, Alcântara, Urucuia, Limoeiro and Marajó Formations)

Tertiary beds (Grajaú, Codó, Itapecuru, Alcântara, Urucuia, Limoeiro and Marajó formations). The Limoeiro and Marajó formations were laid down in the Marajó graben. The Urucuia Formation was described by Lima and Leite (1978).

It consists of light gray, reddish brown, fine- to medium-grained, cross-stratified, medium- to thick-bedded sandstone beds with well rounded grains showing, fair sorting, frosted grain surfaces, slight1y argillaceous, kaolinitic. At the top of the unit silicified sandstone beds and chert bands are also present. The unit is almost free of shale and siltstone beds.

The Urucuia Formation, as much as 500 m thick in the south corner of the basin, covers conformably the Areado Formation. Most of it is present in the São Francisco Basin overlying the Areado Formation or old basement rocks. Northward from the São Francisco Basin, the Urucuia Formation oversteps on to progressively the Serra Grande, Pimenteira, Cabeças, Longá, Poti, Piauí and Pedra de Fogo formations. No Urucuia outcrops are present north of the town of Lizarda where the unit pinches out. The Urucuia Formation is located at the top of the Parnaíba Basin stratigraphic column in the southern portion of the basin.

Only Dodt (1872) mentioned the presence of Dicotyledonous wood in the formation. The Urucuia Formation was considered as equivalent to the Itapecuru Formation (Northfleet and Neves, 1967) because both units are overlying Late Albian formations (Areado and Codó) and are located at the top of the sedimentary column in different areas of the Parnaíba Basin.

In the State of Minas Gerais, the Urucuia Formation overlies lavas and volcanoclastic sediments dated as old as 80 m.y. (Santonian). An age as old as Campanian is inferred for its upper part.

The formation occurs between the Paraná and Parnaíba basins and its paleodrainage system was directed towards the northeastern Brazilian coast. It may represent deposits of the old São Francisco river system reworked by wind activity.

The absence of flood plain shales and the presence of large scale cross-bedding suggest a braided river environment of deposition and strong wind reworking. The Urucuia Formation may have been deposited in an area where the climate was warm and moist.

NOVA IORQUE FORMATION

The name Nova Iorque beds was first used by Plummer (1946) to describe gray siltstone and shale beds occurring in a limited area at the Parnaíba river valley. No detailed studies have been made in the formation to date.

Its stratigraphic relationship is not known, although it is inferred that the Nova Iorque strata overlies disconformably the

Itapecuru Formation. The unit has yielded some plant and fish remains dated by Aguiar (1971) as Eocene. These beds correlate with the basal sediments of the Almeirim Formation of the Amazonas Basin.

The Nova Iorque Formation may have been deposited in a coastal plain environment. Since the Late Cretaceous along the coast, subsidence was occurring while in the central part of the basin uplift was taking place. This was a process opposite to that which occurred during the coastal uplift before the break up of the Gondwana continent.

PIRABAS FORMATION

The name Pirabas was first used by Maury (1924) to describe a Miocene rich fossiliferous limestone section occurring in the coastal area of the State of Pará. Later it was known that the formation is also present at the north part of the Parnaíba Basin in the State of Maranhão. The formation was described by Ferreira (1964).

The unit, over 20 m thick, consists of argillaceous limestones with shale interbeds. It lies unconformably over the igneous and metamorphic basement in some places. Although its stratigraphic relationship with other sedimentary units is not known, it probably overlies several Paleozoic and Mesozoic sedimentary rocks. It is overlain by red beds of the Barreira Formation and it is found only on a fringe close to the coast. The unit is very fossiliferous. Gastropods and pelecypods are the most common fossils, but echinoids, crustaceans, bryozoans and fish also are present in a large number. An Early Miocene age was assigned to it by Maury (1924) and Ferreira (1964) based on its abundant fauna.

The unit may have been deposited in a tidal and shoreface environment and its fauna indicates a deposition under warm climatic conditions.

BARREIRAS FORMATION

The Barreiras Formation overlies the Pirabas Formation along the coast of the states of Pará and Maranhão. It consists of variegated claystone, mudstone, sandstone and conglomerate beds up to 80 m in thickness.

This unit is known since the discovery of Brazil (Pedro Vaz de Caminha´s letter to the king of Portugal). It is assigned a Plio-Pleistocene age based on its stratigraphic position.

CHAPTER 11. TECTONICS OF NORTHERN BRAZIL FROM ORDOVICIAN TO RECENT TIMES

The sedimentary record shows that basins grew by intermittent onlap. Younger beds covered older ones on the margins as basins broadened.

The initial subsidence may have been related to thermal decay (Kinsman, 1975) and the subsequent subsidence may have been related to sediment, volcanic lava, ice and water loading.

The Guyana, Guaporé, Brazilian and North African shields were probably covered by wide-spread ice sheets during Late Ordovician-Early Silurian times, causing subsidence and uplift (glacial and interglacial stages) in the area as well as world-wide rapid regressions and transgressions with corresponding facies changes in the geological record.

During the glacial event in the Late Ordovician-Earliest Silurian, only the area of the Solimões and other northern Andean cratonic basins were located outside the ice-caps. Only in the Late Llandovery (late Early Silurian) did glacial sedimentation take place in the southern Andean area (Crowell and others, 1980).

The unconformity between Silurian and Devonian rocks has not been seen at the outcrop level but it was seen on regional evidence. Ludwig (1964) did not recognize the unconformity; he placed Devonian and Silurian strata together in the same "Trombetas

Group". Carozzi and others (1975) working on cores of wells of the Parnaíba Basin, also admitted the absence of this sedimentation break in both the Amazonas and Parnaíba basins. However, with the aid of data from drilled oil wells and paleontological studies this unconformity (Figure 13) can be traced across the basin (Lange, 1967, p. 240; Daemon and Contreiras, 1971a,b). This disconformity represents a gap of 40-50 m.y. in sedimentation, possibly indicating great tectonic stability in the basin during that time because angular unconformity was not observed in the entire basin. Only minor faults along basin axis may have formed. It is striking that exactly the same stratigraphic relationship exists in the Sahara region basins (Tassili peri-Hoggar, Beuf and others, 1971) where Emsian rocks rest on Llandoverian or Wenlockian deposits.

Evidence presented in Chapter 12, indicates the presence of icecaps on the Guyana, Guaporé and Brazilian shields in Late Devonian and Early Carboniferous times. It is speculated here that one of the remarkable effects of the Late Devonian (Famennian Stage) glaciation was the lack of deposition of Famennian (late Late Devonian) rocks over all ice-free South American pericratonic basins. This is interpreted as due to withdrawal of the sea and growth of a considerable forebulge where erosion took place. In early Early Carboniferous time (Tournaisian Stage) some interruptions in sedimentation took place, but in the late Early Carboniferous (Namurian Stage) a major break in sedimentation occurred. This interruption in sedimentation was probably related to tectonism in the basins and ice build up in western Gondwana.

The Parnaíba and Amazonas basins lost communication with the sea during the Visean time, causing deposition of fluvial and lacustrine sediments in both basins. These deposits suggest that either the east side of the Amazonas Basin was being uplifted or that a collision between Gondwana and Laurasia closed the Phoibic Ocean (ocean between these mentioned continents before the Hercynian orogeny – McKerrow and Ziegler, 1972).

A hot spot developed in the north coast of the Amapá territory (Figure 84) that uplifted the coastal area in Permo-Triassic time. A swarm of dikes is present in the area suggesting the existence of a hot spot. It corresponds to the hot spot identified by Smith (1982) in the State of Florida, U.S.A. An examination of several basins in eastern South America and Western Africa has shown that the present coastal area began to be uplifted since the Carboniferous time. Such an almost simultaneous uplift along the entire coast of Western Africa and eastern South America strongly suggests that this major epeirogenetic movement may have been caused by the initial stages of opening of the present Atlantic Ocean. In the Parnaíba Basin, as Mid-Pennsylvanian

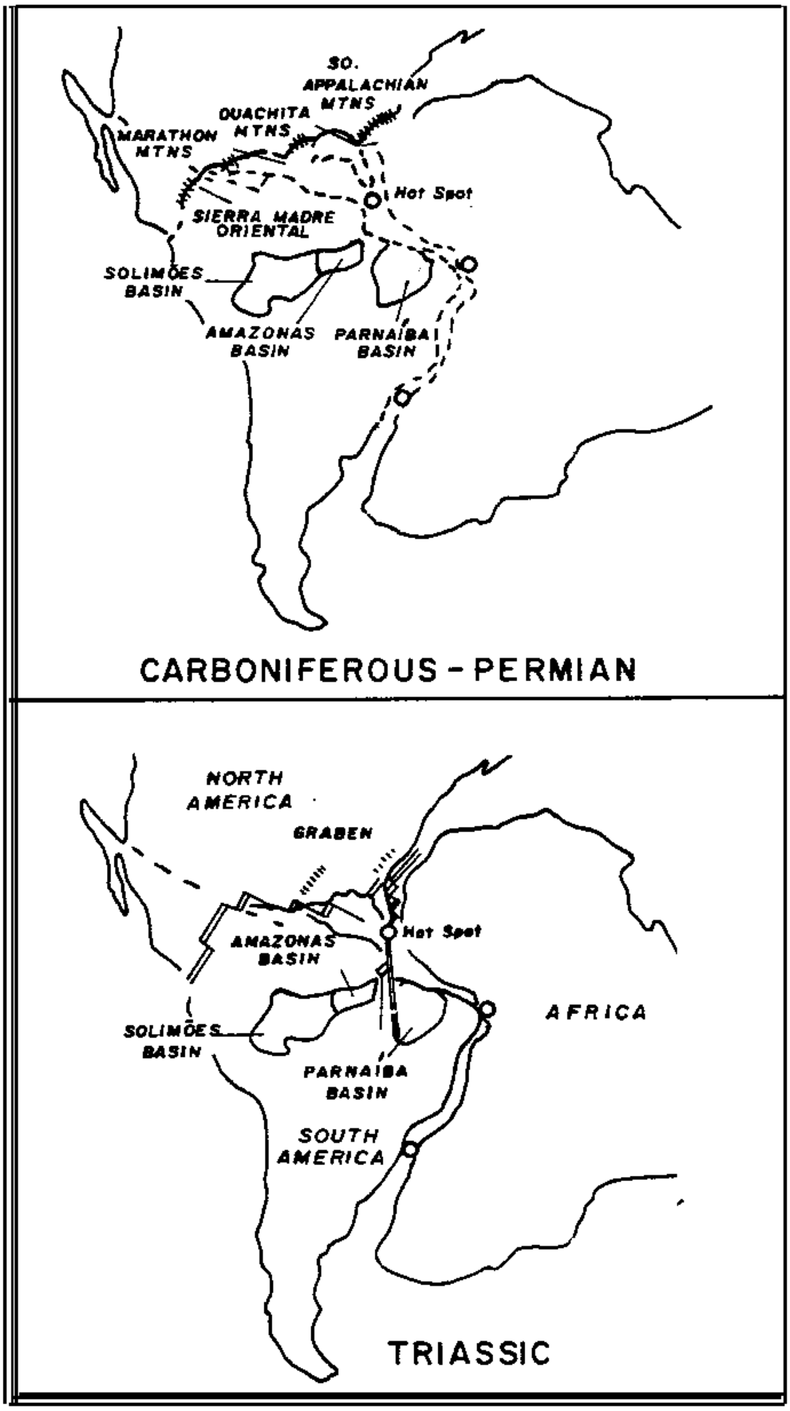

FIGURE 84 Development of the Triassic Marajó rift due to the presence of a hot spot in southern Florida or in northeast Amapá Territory (Modified from Smith, 1982)

deposition commenced, a fundamental change in basin tectonics occurred. The eastern Brazilian coastal shield and northern Ferrer Arch had risen, shifting the axis of greatest basin subsidence 500 km towards the west (Mesner and Wooldridge, 1964, p. 1505).

In the study of marginal Atlantic basins of eastern Brazil, Estrela (1972, p. 30) concluded that during the stage of pre-rift development, upwarping occurred in Paleozoic times. He observed that some Precambrian limestones were eroded from areas of major upwarp before deposition of Juro-Triassic continental sediments. In the Paraná Basin, the present eastern limit of Devonian outcrops does not correspond with the original margin of the basin. It has attained its present shape by erosion. The Devonian sea probably covered the present coastal shield; only the later uplift of the Ponta Grossa arch and the successive rises of the eastern Brazilian shield, account for the fast and final erosion of Devonian sediments, which originally covered that eastern part of the Paraná Basin (Sanford and Lange, 1960, p. 1352). When the ice-sheets came from South Africa (Crowell and Frakes, 1972), they may have found the shield uplifted, but they could override the elevated area. The Permo-Carboniferous Itararé tillites show a large number of reworked Devonian palynomorphs as mentioned by Eglemar C. Lima (1983, written communication) indicating erosion of Devonian terranes due to the glacial activity.

In the Karoo Basin, on the southern tip of Africa, depocenters show a shift from west to the east from Devonian to Permo-Carboniferous time (Theron and Blignault, 1973, p. 354). On the southernmost tip of South Africa, the Cape orogeny occurred in Permian time, but De Villiers (1944) has suggested that a western tectonic episode predates the main southern event.

Furthermore, the Congo Basin, in central western Africa, where only continental sediments have been deposited since the Permian (Frakes and Crowell, 1970a), shows thinning in the sedimentary section on its western margin and thickening on its eastern margin, indicating uplift on the coastal side of the basin and larger subsidence in its eastern side.

In the Taoudeni, Iullemmeden and Chad basins of northwest Africa, more than 1,000 m of Permian-Jurassic continental beds, called the "Continental Intercalaire" were laid down. The "Continental intercalaire" Group is thickest in the Chad and Iullemmeden basins in the east and thinnest in the Taoudeni in the west where a thin veneer of Carboniferous rocks is present (Petters, 1979). The structural geometry of these intracratonic basins was substantially modified by this major event from Carboniferous time onward. I conclude that the general uplift of the Atlantic coastal areas of South America and Africa from

Carboniferous to Jura-Cretaceous times indicates that tectonism associated with the break-up of Gondwana started long before rifting and sea floor spreading.

Tilting of the east side of the South American continent occurred up to approximately 800 km away from the present coastal areas as can be observed in Carboniferous isopach maps. During this uplift, from about 800 km to 1,500-2,000 km away from coastal areas, general downwarping occurred not only in South America but also in Africa. The bulk of the sediments in the Congo Basin, entirely continental, were deposited during this period of coastal uplift and interior subsidence. This process of downwarping of the craton could be responsible for the genesis of that intracratonic Basin, as well as the intensification of subsidence in pre-existent ones as the Amazonas Basin. The subsidence may make room not only for marine but also for continental sediments.

Gentle upwarping on the north flank of the Amazonas Basin may be attributed to doming that preceded the genesis of the Takutu rift (Rio Branco arch of Amaral, 1974). On the northern flank of the Amazonas Basin, the exposed and preserved sediments are thicker, older, and deposited in a deeper water environment than along the southern flank as discussed in the stratigraphy of the basin. This suggests that shallower water sediments, comparable to those on the southern flank, were removed by erosion which indicates deeper erosion on the northern flank due to uplift of the area. Carozzi and others (1972) noticed a 70 km shift of the depositional axis from the north to the south in the western part of the Amazonas Basin, during Permo-Carboniferous time, which possibly reflects the uplift formed before the collapse of the Takutu rift. Isopach countour maps show the truncation of many formations in northern part of the basin. Figure 85 shows the tectonic framework of the area between Amazonas and Parnaíba basins.

The Purus arch, the divide between the Solimões and Amazonas basins, was also downwarped during Permo-Carboniferous time. The subcrop belt of pre-Pennsylvanian formations in the Purus arch area became wider because subsidence reduced the dips as observed in bore holes correlation. In the Purus arch area, in the western side of the Amazonas Basin, although the formation thicknesses are small, the subcrop belt area is nearly 8 times wider than that on the northern or southern flanks. Subsidence of the western Amazonas Basin is related to the uplift of the eastern Amazonas Basin. Isopach contour maps are parallel to the western margin of the basin. In the eastern basin margin, isopach maps show contour truncated due to an uplift and erosion.

General uplift of the east side of the Amazonas Basin from Permian time on directed drainage to the west toward the Paleopacific Ocean. The eastern side of the basin, where the Atlantic Ocean is today, was a source area for continental clastics. Red sandstone beds are intercalated in the Itaituba and Nova Olinda formations indicating a source of continental sands in the eastern side of the Amazonas Basin. Hot and dry desert conditions prevailed resulting in a thick section of evaporite and limestones in the Solimões and Amazonas basins when a new sea transgressed from the Paleopacific Ocean in Mid-Pennsylvanian time as evidenced by the Andean fauna (Mendes, 1959, 1961). During the eastward sea advance, the Coari high, due to a large sedimentary loading in the Solimões Basin, migrated eastward allowing the sea to reach the Amazonas and Parnaíba basins intermittently. The new position of the high corresponds to the present position of the Purus arch. A large amount of salt concentrated in the deeper parts of the Solimões and Amazonas basíns, resulted in large differential subsidence between the two basins and across the shallow Purus high. The raw material for carbonate and salt came in solution, adding to the normal detrital load derived from the watersheds. Only in the eastern half of the Amazonas Basin salt deposits are absent, reflecting continuous upwarping of that area. The evaporitic sedimentation was cyclic as a result of sea level changes which may have been caused by fluctuations of ice volume on ice-covered areas of Gondwana. Despite the disintegration of the ice-caps in Permian times, one of the largest regressions on the earth took place (Hallam, 1977), carrying the sea entirely out of the

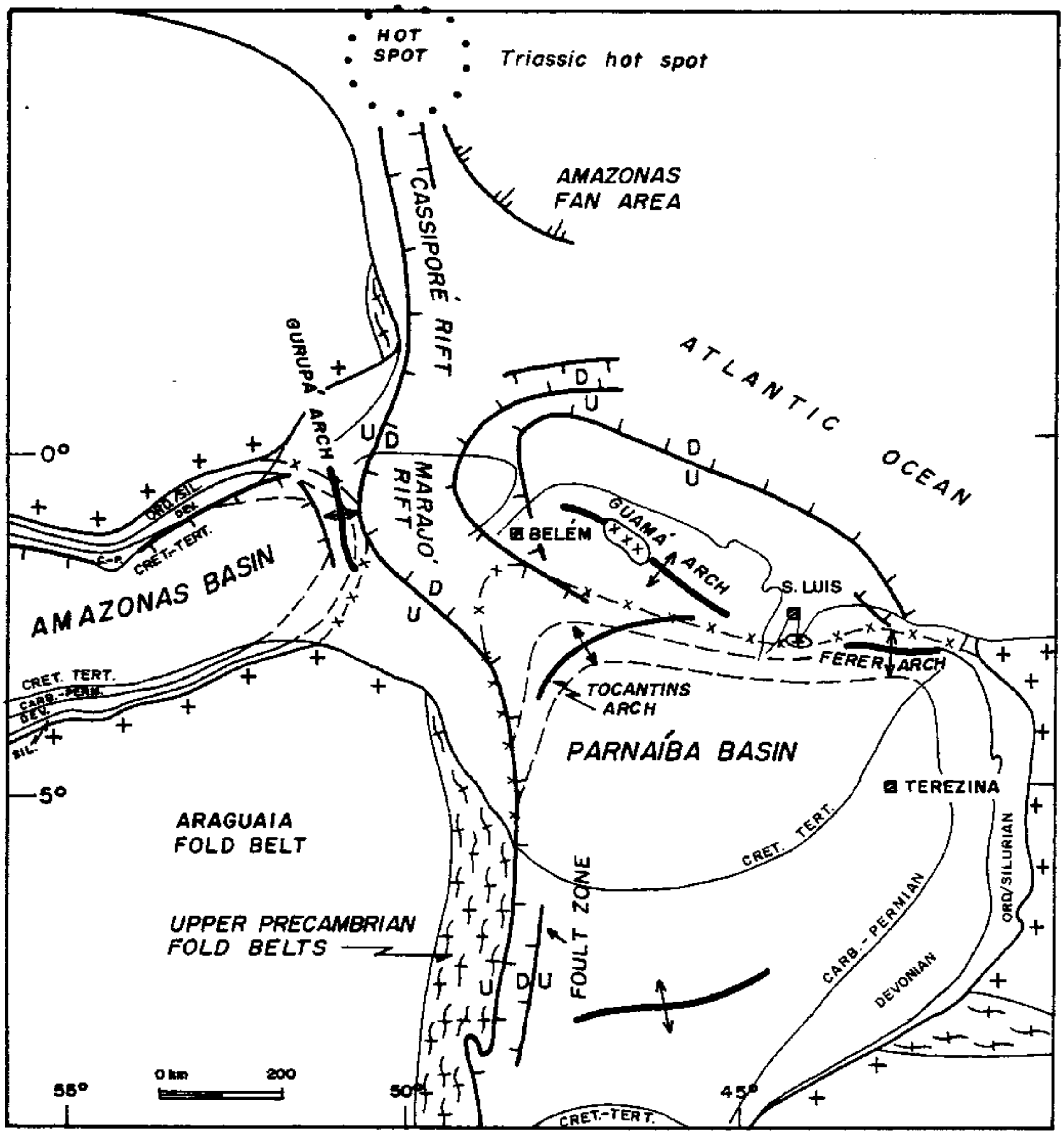

Figure- 85 Tectonic framework of the area between Amazonas and Parnaíba basins.

northern basins. I postulate that this long-lasting regression may have been linked to subduction of oceanic ridges, and to the orogeny that made up Pangea, and built mountain chains in many places in the world, and so increased the volumetric capacity of the oceans.

From Permian to Jurassic times (Thomaz Filho and others (1974), much basic igneous activity took place in the Solimões and Amazonas basins in the form of diabase dikes and sills. In the Amazonas Basin, the main dike directions seen on surface and radar imageries are N and N25ºE to N45ºE, and the main magmatic activity occurred in sedimentary areas while minor dike intrusions also occurred on shield areas.

I have calculated the volume of diabase in the Solimões (6,604 km^3) and Amazonas (7,489 km^3) basins with the aid of Petrobras computers. Some Solimões Basin wells present a total diabase sill thickness of 925 m and Amazonas Basin wells 750 m. These figures are very impressive; diabase rocks may constitute one-third or one-fourth of the entire stratigraphic section.

In some central areas, dikes, as interpreted from seismic surveys made by Petrobras, are a few kilometers thick making approximately 5% or more of the surficial rocks. The crust probably extended during the tholeiitic activity. When the magmatism ended, the emplacement of the diabase dikes produced an enlargement of the continental crust. The successive diabase sills may also have elevated the land above the level of intrusions, but afterwards this igneous load may have depressed the crust. The greatest thickness of diabase was located at the basin axis, so this considerable load on central parts of the basin may have contributed to uplift of the basin flanks.

In areas of compressive stress, as in the Andean basins, stratigraphic sections are free of diabase, but in areas of tensile stress, as in the Solimões Basin, much diabase was intruded into the section from Permian to Jurassic times.

During the magmatic activity, the area was under erosion; streams of the Amazonas and Solimões basins probably drained toward the Andean foreland basins where a thick section of red beds and evaporites were deposited from Permian to Jurassic times (Benavides, 1968). In contrast, streams of the Parnaíba Basin may have drained toward eastern Brazil, from Triassic to Jurassic times.

The eastern Amazonas Basin area, originally most elevated, was rifted in Jurassic times (Caputo and others, 1983) forming the Cassiporé and Marajó rifts related to the opening of the North Atlantic Ocean. Away from the coastal area, many faults were reactivated, forming new rifts in the Amazonian platform along weakness zones. The shoulders around the coastal rifts remained uplifted and fed the rifts with coarse sediments which were in part mixed with basalt flows. In Early Cretaceous time compressive stress predominated in the Amazonian platform preventing basic magmatic activity. In Late Cretaceous time, while the marginal Marajó rift was being filled with fanglomerates and coarse clastics, the eastern part of the Amazonas Basin (Marajó Shoulder) was being eroded, but the western Amazonas and Solimões basins were being filled with the Alter do Chão continental red beds. This is observed in the cross-sections along the Solimões and Amazonas basins. Towards the Andean basins, in Peru, Cretaceous continental sediments changed laterally to marine strata (Caputo, 1974).

During the Late Cretaceous, the coastal uplifted Brazilian areas together with rifted areas started to subside due to general cooling of the crust and loading of sediments. The coastal thermal uplift started to decay; the crust, thinned by erosion or extension, subsided and started to receive sediments. This mechanism was invoked by Kinsman (1975) for subsidence of rifted basins. The sediment load continued to cause more subsidence in the Marajó Basin and offshore, while the Marajó shoulder remained uplifted.

In the Parnaíba Basin, coastal subsidence started in the Late Cretaceous when the basin tilted northward due to cooling similar to that which takes place in areas away from the hot spreading centers of middle ocean ridges as explained by Kinsman (1975).

Figure 86 shows the tectonic evolution of eastern Amazonas and Parnaíba basins since Late-Ordovician to Recent time.

Strong compressive episodes related to the Andean orogeny reached the Solimões Basin in Early Cretaceous and Tertiary times (Caputo, 1984; Caputo and others, 1983). The high Andean chain was being built up and the load of the mountain belt may have depressed the foreland basin areas which were filled with marine and continental beds. Beyond the peripheral sink the Jutaí arch was uplifted. Close to the eastern Andes, molassic sedimentation was almost continuous with many conglomerate layers. Areas near the Solimões Basin and Brazilian shield regions received fine elastics (Mason and Caputo, 1964, Caputo and others, 1979). All areas around the Andes were slowly depressed, so that part of the Jutaí arch itself became a portion of the peripheral depression. The Jutaí arch apex migrated eastward. In Miocene times, the Andean belt probably blocked all drainage of the foredeep toward the Pacific Ocean, causing reorientation of the drainage toward the Atlantic Ocean. The last marine deposition in the Subandean belt occurred in Oligocene time (Caputo and others, 1979). The high clastic supply disturbed the thick carbonate shelf deposition in the Atlantic Ocean at the Amazonas River mouth as observed in bore holes drilled by Petrobras. Finally, Late Tertiary and Quaternary sedimentation continued in the Solimões and Amazonas river valleys, and coastal areas

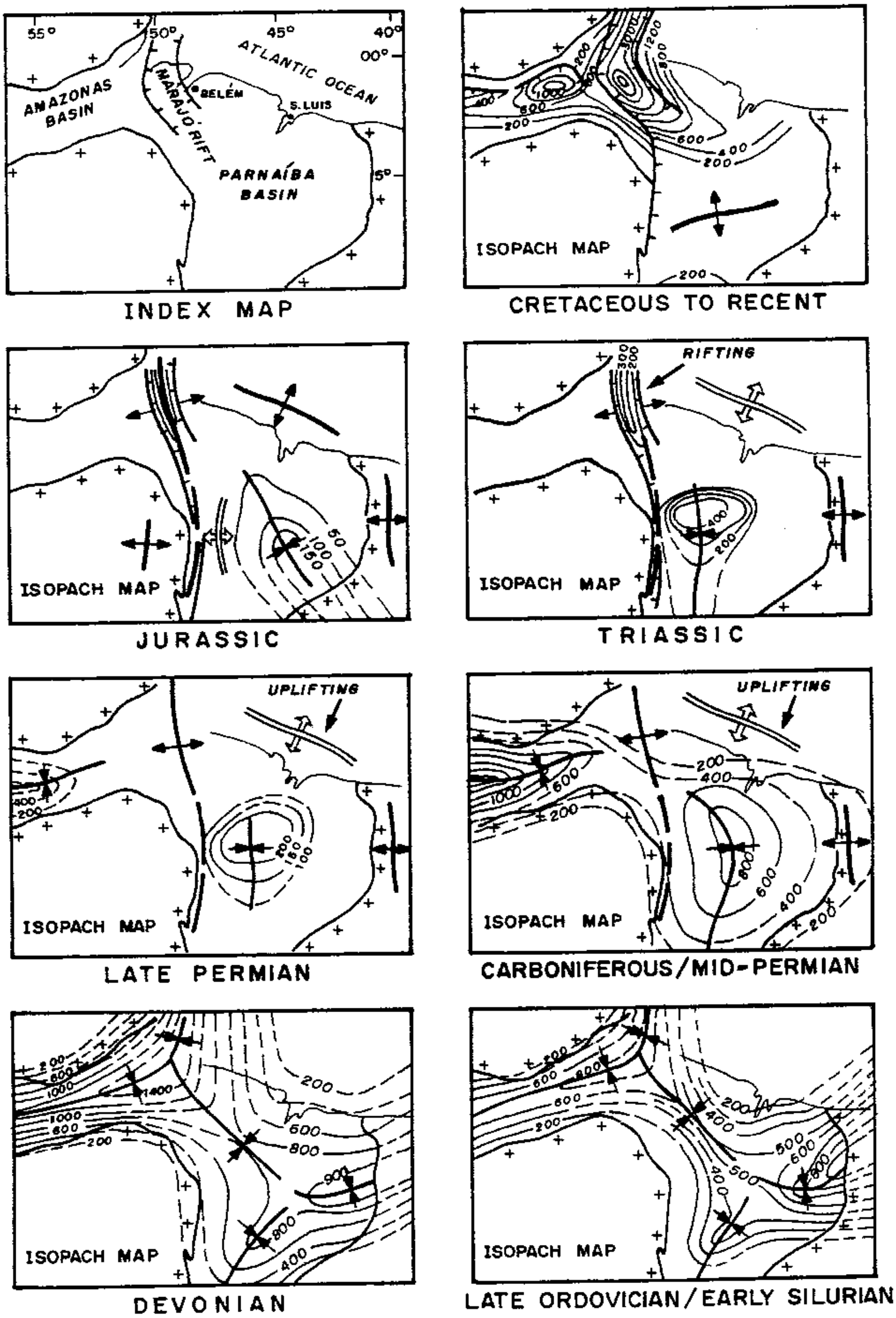

Figure- 86 Tectonic evolution of eastern Amazonas and Parnaíba Basins .

This study has shown that during Devonian time tectonism was not active, so diamictites deposited in north Brazilian basins are not related to tectonic disruptions.

CHAPTER 12. LATE DEVONIAN GLACIATION CONTROVERSY

BACKGROUND

Following the description and correlation of stratigraphic units of northern Brazilian basins it is appropriate to discuss the genesis of the Late Devonian diamictites observed in many places. Due largely to the sedimentary cover, as well as to the inaccessibility of the Solimões Basin (previously named Upper Amazonas Basin), Late Devonian diamictites were not detected there until 1978, when Caputo and Vasconcelos recognized them as resulting from glacial activity based on rock texture and correlation with similar glacigenic beds of the Amazonas and Parnaíba basins.

In the Amazonas Basin, in the Curuá Formation (now Curuá Group), De Moura (1938) first interpreted Devonian diamictites as deposited under glacial conditions, based only on rock characteristics observed in borehole cores. Later, Bouman and others (1960) mentioned that some clasts from cores of the unit were striated and called the section containing them as the "rafted pebble member" of the Curuá Formation. Ludwig (1964) refuted the glacial nature of these sediments based on the supposed absence of glacial features and considered them as turbidites.

Caputo and Vasconcelos (1971), as well as Rodrigues and others (1971), Carrozi and others (1972, 1973), reinterpreted the diamictites as a result of glacial activity based on rock texture, pebble striae and wide extent. Later on Macambira and others (1977) described outcrops with diamictites bearing some striated clasts and laminated shales with sparse clasts in the Curiri Formation of the Curuá Group. Esteves and Carneiro (1977) concluded that the Curiri glacial deposits were related to high altitude rather than to high latitude environment. Rocha-Campos (1981b) pointed out that the glacial nature of the Curiri sediments could not be definitively established on the basis of available published data. Consequently, in Hambrey and Harland book (1981) the Curiri sediments were classified as diamictites of unknown origin.

In the Parnaíba Basin Kegel (1953) recognized diamictites with striated clasts in examining cores from the Carolina well (1-Cl-1-Ma). He supposed that the diamictites were included in the Longá Formation, interpreting them as a result of glacial deposition. Malzahn (1957) mapped tillite-like and varve-like sediments overlying a striated surface at the top of the supposed Serra Grande Formation (now Serra Grande Group) but later it was shown by Aguiar (1971) that the glacial beds lay at the top of the Cabeças Formation of Late Devonian age (Andrade and Daemon, 1974) instead of at the top of Serra Grande Group of Ordovician-Silurian age. The Ipu Formation has some diamictite beds seen in the Cocalino well (1-CI-1-MA) and regional mapping (Mabesoone, 1978; and Lima and Leite, 1977), but not in the upper Jaicos Formation of the Serra Grande Group.

Ludwig (1964) implicitly also contested the glacial origin for the Parnaíba Basin diamictites in Petrobras unpublished reports. Bigarella and others (1965) studying the true Serra Grande Group denied the presence of glacial features at its top.

Barbosa and others (1966) mapped diamictites with striated clasts in the supposed Pimenteira Formation in the western part of the basin considering them as of glacial origin.

Rodrigues (1967) argued that the diamictites could be a result of either mud flows or glacial activity.

Later on, Bigarella (1973a) re-examined striated pavements, tillites and varve-like sediments, described by Malzahn (1957) in the State of Piauí, confirming their glacial nature, but again misplacing them at the top of the Serra Grande Group, as Malzahn (1957) had previously done. Andrade and Daemon (1974) contested both the glacial origin and the turbidity currents deposition for the Parnaíba Basin Devonian diamictites. They argued that the fauna of the Cabeças Formation was incompatible with a glacial environment, and, in addition, that the depositional environment was too shallow for turbidity currents to develop. However, Copper (1977) stressed that differences in faunal composition of contemporary Devonian marine communities in Brazilian Paleozoic sedimentary basins suggested variation because of climatic gradients, and that a drastic cooling in South America might have been responsible for the Frasnian-Famennian world massive biotic extinction. Carozzi and others (1975) documented the presence of diamictite and varve-like sediments with dropstones in cores of the Cabeças Formation of the Tem Medo well (1-TM-1-Ma), and Lima (1978) envisaged the Devonian diamictites as resulting from syn-sedimentary faulting along the western border of the Parnaíba Basin. Frakes (1979), based on Bigarella's (1973) studies, cited glacial activity at the top of the Serra Grande Group (Silurian-Ordovician age) instead of at the top of the Cabeças Formation (Late Devonian age).

Heckel and Witzke (1979) pointed out that during Devonian time no glaciation developed on the Gondwana continent. Rocha-Campos (1981c) favored a glacial genesis for diamictites of the Cabeças Formation, but the evidence presented did not convince Hambrey and Harland (1981) who classified the Cabeças tillites as diamictites of unknown origin.

Many investigators do not recognize glaciation during Devonian times because they have not seen evidence anywhere in the world for such glaciation. They argued, had these rocks been deposited under glacial conditions there whould be two possibilities: a) either their age are not Devonian; or b) they could have been deposited under alpine conditions (Esteves and Carneiro, 1977). Had the rocks been deposited under alpine conditions, they would not imply the existence of a true ice age nor a worldwide climatic refrigeration.

This brief review shows that the genesis of the Devonian diamictites of northern Brazil has been controversial. The main reason for this controversy is that several different geological environments may develop many features in common, and rocks having a superficial resemblance to glacial sediments may have been deposited by non-glacial processes and may, on detailed investigation, contain uncertain inprint even of distant glaciation (Crowell, 1964). These unusual strata have been examined here, in outcrop and in cores, in order to determine their mode of origin.

EVIDENCE FOR THE LATE DEVONIAN GLACIATION

The age of the diamictites of northern Brazil has been contested by some investigators, because the macrofossils found in beds above the diamictites are not index-fossils. However, in the Amazonas and Parnaíba basins the diamictites overlie directly several species of *Protosalvinia*, a plant or algae dated as Early Famennian by Niklas and others (1976). This same stratigraphic relationship in two separated huge basins indicates that there is

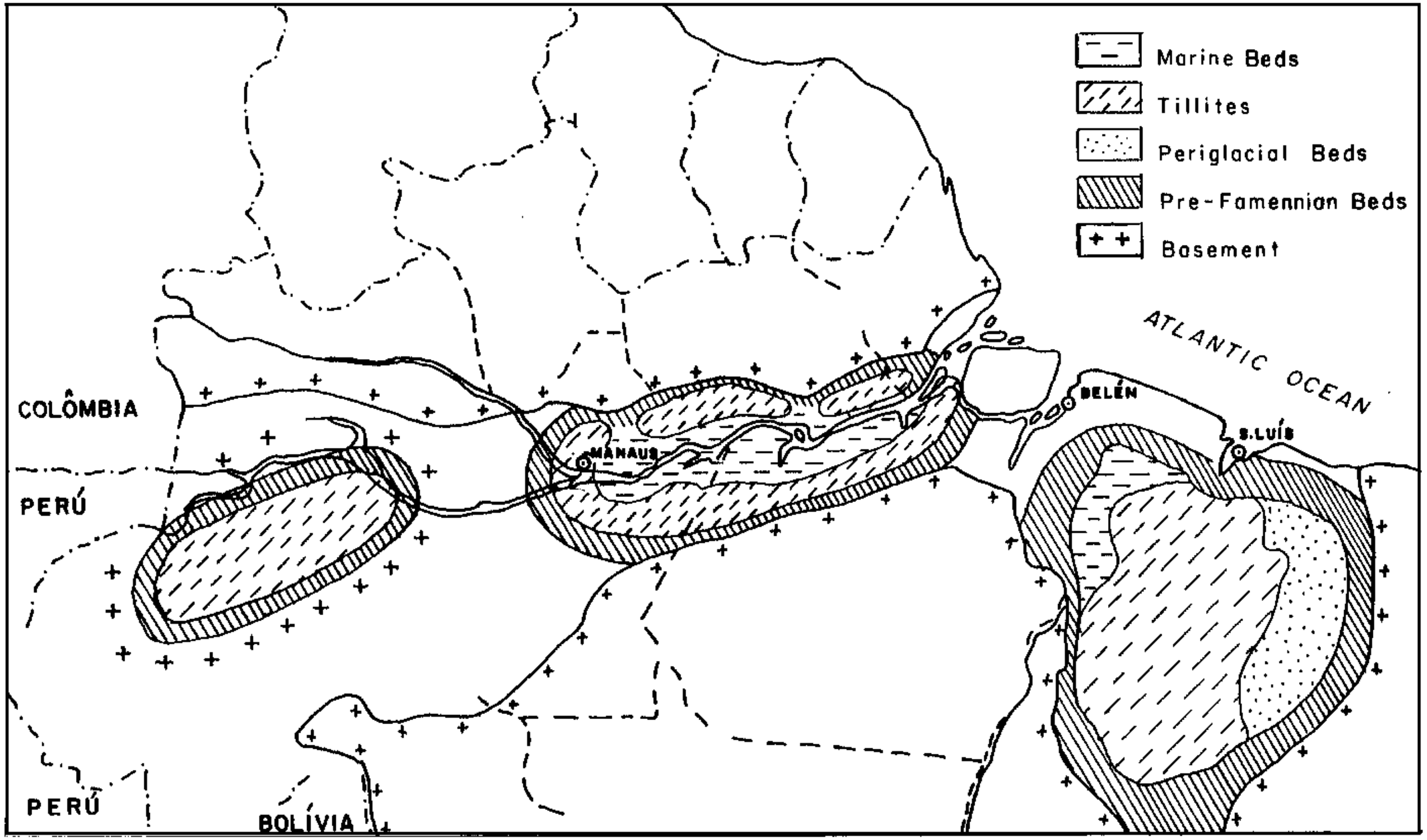

Figure 87 - Distribution of Late Devonian tillites in the Solimões, Amazonas and Parnaíba Basins.

a normal succession between the *Protosalvinia* zone and the glacial beds in both basins.

Palynomorphs found in the Curiri and Cabeças formation diamictites were dated as Early to Mid-Famennian and sedimentary rocks above the tillites were dated as Late Famennian by Daemon and Contreiras (1971a,b). In these conditions the main glacial activity is bracketed between Early Famennian and Late Famennian. Therefore, a Mid-Famennian age is here attributed to these tillites.

Because diamictite itself does not indicate under which climatic conditions it was generated it is necessary to use many criteria (previously discussed here) as evidence of the presence of a former glaciation. There is strong evidence that the Late Devonian diamictites are tillites in northern Brazil, some of these evidences are:

1. Presence of characteristic flat-lying diamictites (Photos 1, 3, 4, 5, 10), with great petrographic variety deposited over a distance of 3,500 km in diff'erent huge intracratonic basins, in the same time interval, rules out a tectonic origin for these rocks (Figure 87). In contrast, although pseudotillites may be frequent, they appear at irregular intervals and are localized (Harland and others, 1966). Underlying and overlying sandy and shaley conformable formations show the same characteristics over wide areas, indicating no tectonic disturbance during the Late

Devonian time in northern Brazil.

2. The dip of the glacigenic formations and striated pavements were lower than half a degree when they were formed, indicating the existence of a flat substrate not adequate for large-scale debris flows and graded beds, characteristic of turbidity currents, were not recognized in the studied sections.

The maximum dip found in wells in the Curiri Formation is 1º 30'. This larger dip is due to the additional Permo-Carboniferous, Mesozoic and Cenozoic subsidence. The present dip of the Cabeças Formation diamictites is lower than half a degree. There is no indication of alpine-like teetonism around the Amazonas and Parnaíba basins since Early Ordovician time as can be observed in the tectonic development of the Amazonian and Brazilian platforms.

Basement clasts and erratics, found more than 100 km from their sources, favor a glacial origin. Debris flow and slides can transport large blocks, but generally not over long distances in a shallow sea or lake.

3. By analogy with the Pleistocene tills faceted and striated stones should not be abundant in older glacial sequences, although some should be observed.

I found in the Amazonas and Parnaíba basins, striated, faceted and some polished clasts (Photo 13), in respective outcrop areas and in the subsurface. Some clasts have a flatiron shape

(Photo 7). Von Engeln (1930) suggested, qualitatively, that a striated and flatiron form was unique to stones transported by glaciers. Glacial striae are most conspicuously developed on flat surfaces where they form intersecting or subparallel sets while tectonic striae are parallel. Mass flow processes are less effective in scratching hard than soft rocks, and such striae are therefore less marked than those on glacial facets. According to Thornbury (1969), stones striated by mud flow, usually lack facets. In view of their overall characteristics the striated clasts observed in Curiri and Cabeças formations are regarded as a result of glacial abrasion.

4. The relative poverty in volcanic clasts in the Amazonas and Parnaíba basins diamictites suggests that volcanic mud flows and lahars were not playing any role in the diamicton deposition.

5. Glacial abrasion produces characteristic polished rock pavements with striations, grooves and crescentic fractures. Thus, the presence of extensive grooved and striated pavements may indicate glacial abrasion especially if they are overlain by diamictites.
At about 49 km from Canto do Buriti town, on the old road to são Raimundo Nonato town, Malzahn (1957) and Bigarella (1973a) documented a section composed of cross-bedded sandstone, diamictite and varve-like sediments with dropstones at the top. There is a horizontal

striated surface between the sandstone and the diamictite beds. The orientation of parallel striae is not coincident with cross-bedding strike, indicating that lineations are not the result of intersection between cross-bedding and the striae in ground surface, nor with joint sets. The striated pavement is located close to the top of Cabeças Formation which is not tectonically deformed. The striated pavement was probably frozen during the glacier motion allowing striae formation on a sand substrate. On the road to Santa Iria, about 4.5 km from Canto do Buriti town, striated pavements cover areas of about 200 m across (Photo 18). Irregular width is observed between lineations and grooves which shows a relief of about 1 to 10 mm in depth, and they cut through the cross bedding strike in a variable angular fashion. Striae follow N15ºE, N25ºE and N40ºE directions. I found other striated pavement in the upper Cabeças Formation at Morro Comprido. It is located about 26 km from the town of Canto do Buriti on the old road to São Raimundo Nonato (Photos 16, 17).

6. At Morro Comprido a silicified conglomerate boulder 1.4 m across derived from the basement lies on the ground. It appears to be a far travelled erratic left by glaciers. The existence of striated pavements overlain by diamictites and or erratics indicates that glaciers were at work in north Brazil during Famennian time.

7. Icebergs are capable of rafting and releasing coarse rock frag-

ments into finer well-sorted sediments. In the Parnaíba Basin, Carozzi and others (1975) documented the presence of rhythmic varve-like sediments which contain small pebbles (Photo 12) in cores from the "Tem Medo" well (1-TM-1-Ma) and Malzahn (1957) and Della Piazza and others (1966) also described varve-like beds with scattered pebbles overlying diamictites. These oversized clasts are interpreted as dropstones because they are isolated in a much finer grained matrix. They were apparently rafted on the surface of lakes. The known dropstones are 1 to 9 cm across and show disruption of laminations at the bottom and draping of sediment over their tops. I observed that the varve-like section overlies and underlies diamictes in the Tem Medo well (1-TM-1-MA). Dropstones could be ice-rafted or plant-rafted, but the rhythmic banding typical of glacial lake varves and the presence of diamictites strongly suggests that these dropstones are ice-rafted. At the least they indicate very cold climate environment.

In the Amazonas Basin, small pebbles and granules included in marine laminated shales have been observed by me and other investigators (Schneider and others, 1975; Macambira and others, 1977; Carozzi and others, 1973), suggesting the presence of icebergs in the Curiri Formation in the Amazonas Basin in Late Devonian time.

8. Deformed and structureless sandstone beds are present everywhere over the Cabeças Formation (Photos 19, 20, 21). All

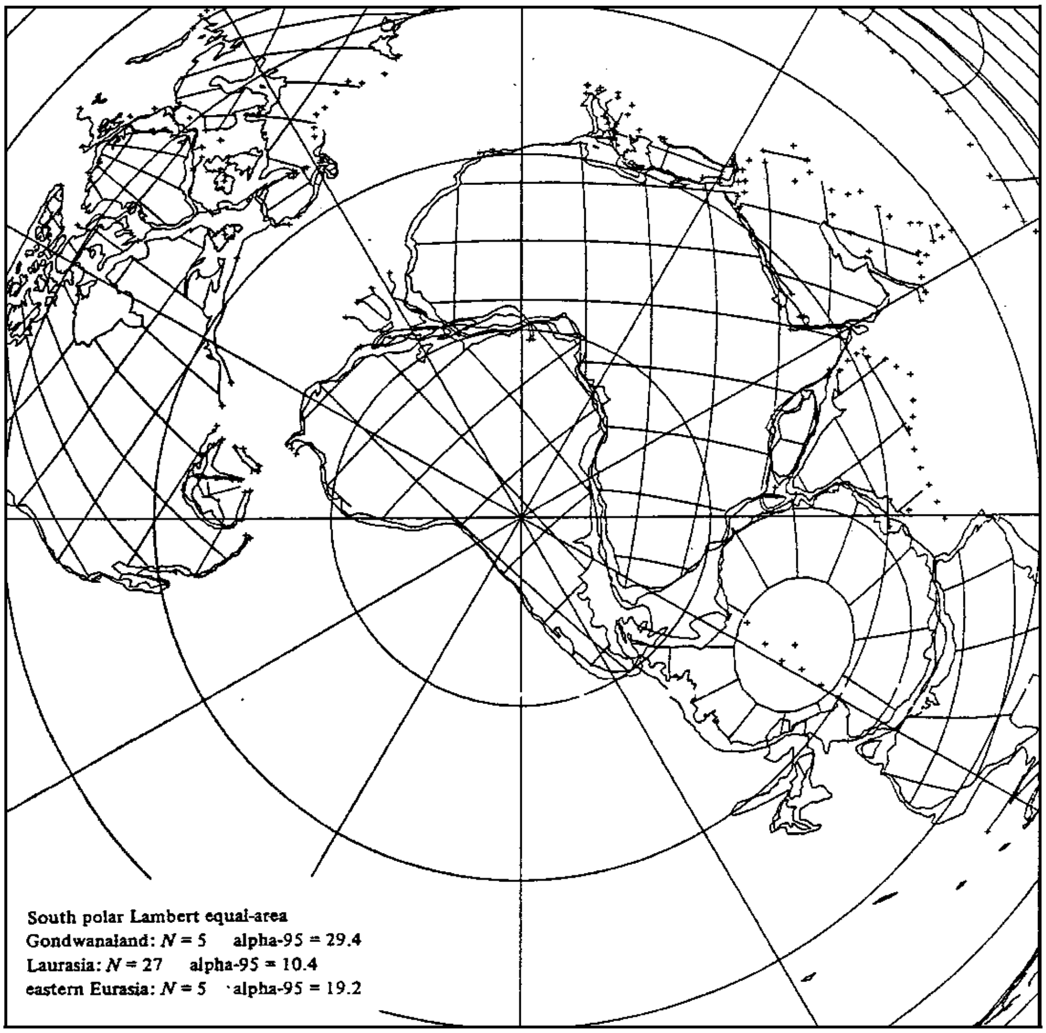

Figure 88. The common boundary area between Argentina and Bolivia was located on the South Pole during Late Silurian time (Ludlovian) according to paleomagnetic data (after Smith and others, 1981).

geologists who have mapped the unit have observed extensive "slumped beds", "convolute folds" or "load cast" features. The deformed sandstone beds may extend over an irregular surface and show undulating shapes; or in other places, massive sandstones may overlie a parallel or cross-bedded sandstone body. These deformed and massive sandstone beds are here interpreted as a result of colapse and fabric destruction when stagnant ice underneath and ice cement melted. Some water-escape structures are seen in many outcrops. Plummer (1948) described sandstone dikes and veins. Possibly the sand may have filled patterned ground cracks from above in a permafrost environment.

The multiple lines of direct evidence presented in this study confirm a glacial event in the Late Devonian time. Although some of these features can be attributed to different environments, all of them have a factor in common: a glacial and periglacial origin.

Furthermore, other lines of evidence, though indirect, may be cited:

9. One of the causes of the fall and rise of sea level is the change of amount of ice tied up in continental glaciers. Eustatic changes in sea level may be produced by several other causes (Crowell, 1983), but glacioeustatic changes are typically very fast geologically speaking. Regressive beds or unconformities on

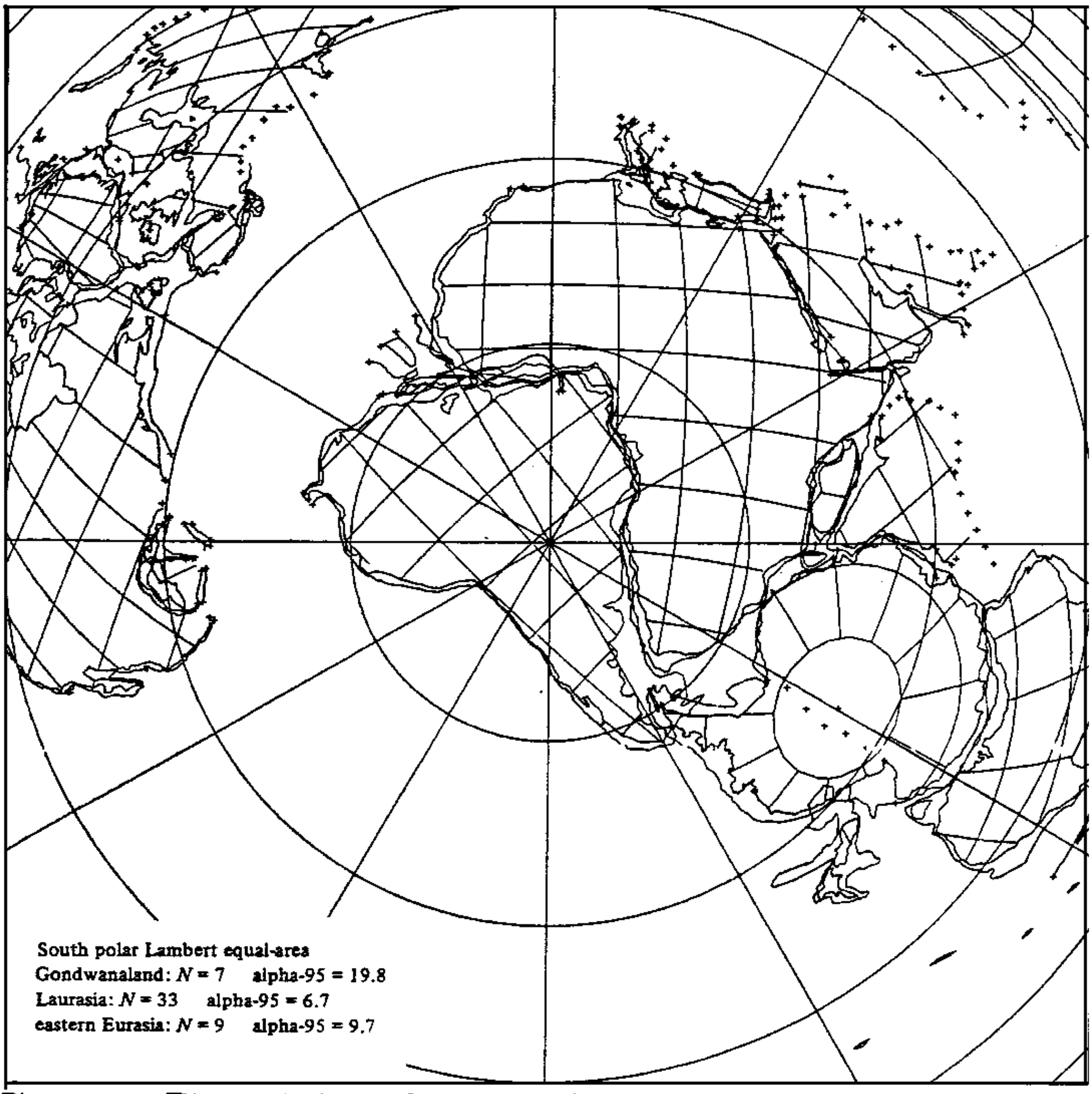

Figure 89. The State of Mato Grosso do Sul area was located on the South Pole during Late Devonian time (Frasnian Stage) according to paleomagnetic data (after Smith and others, 1981)

the world geological record should be expected during times of extensive glaciation. Transgression-regression curves, presented to date for the Paleozoic Era, are not precise enough; they are too generalized to show short, but extensive sea-level changes. In the Solimões, Amazonas and Parnaíba basins, the Famennian sediments are regressive. In Andean basins Famennian beds are not known, therefore, a gap occurs between Devonian and Carboniferous beds. In northern Africa, Famennian sediments are typically regressive (Freulon, 1964). Basset and Stout (1968) documented a through-going unconformity between the Frasnian and the Famennian stages in Canada. The Euramerican record for the Frasnian (and Early Famennian) is strongly transgressive, but Mid-Famennian is regressive in all nearshore locales (Johnson, written communication to Professor Crowell, 1983). The progressively regressive spread of red beds and conglomerates in the New York State in the immediately post-Frasnian is normally attributed to Acadian orogenic movements (House, 1975), but glacioeustatic regression may also be involved. Regressive sandstone deposition and important facies-changes have been recorded in the British Isles at the Late Devonian time. The Mid-Famennian saw general regressional conditions in Russian platform (House, 1975).

The above mentioned examples suggest that the Mid-Famennian regression was worldwide and it correlates very well with the cli-

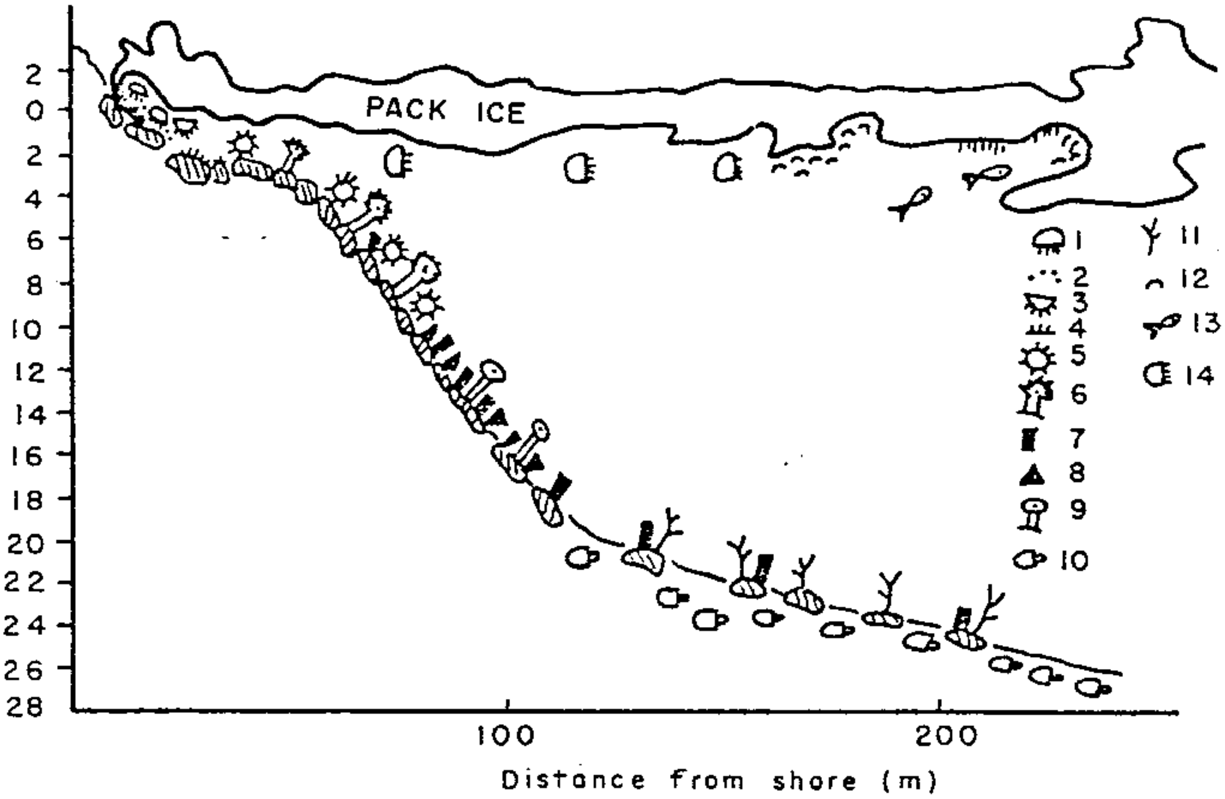

Fig.90 Vertical distribution of bottom biocoenoses on stony and muddy grounds under ice cover near Alexandra Land (Franz Josef Land). 1- Cryolittoral *Gammarus setosus* biocoenosis (1)+ Diatomea g. sp. (2) (the lower horizon of the littoral in ice lacunae). II- *Rhachotropis aculeata* biocoenosis (3)+ *Acrosiphonia sonderi* (4) (at a depth of 0.3m, boulders, stones). III - *Tealia crassicornia* biocoenosis (6)+*Strongylocentrotus* sp. (5) (at 3-10m, boulders, stones). IV - *Hiatella arctica* biocoenosis (7); typical forms: *Balanus balanus* (8), *Hormatella josefi* (9) (at 10-13m, boulders, stones). V- *Cellepora surcularis* biocoenosis (11)+ *Hiatella arctica* (7)+*Mya truncata* (10) (at 18-25m, muddy clay, stones). VI - *Apherusa glacialis* biocoenosis (12); typical forms: *Acrosiphonia sonderi* (4), *Boreogadua saida* (13) *Gammarus wilkitzkii* (14) (at the lower surface of the pack ice).
After Golikov and Averincev, 1977.

Figure 90. Vertical distribution of bottom biocoenoses on stony and muddy grounds under ice cover near Alexandra Land.

max of glaciation in South America. Before the end of the Devonian Period, glaciation ended and a new worldwide transgression occurred. The overlying Longá, Lower Oriximiná and Upper Jaraqui formations document this transgressive event in north Brazil.

10. Loess deposits may indicate periglacial reworking. In Famennian time ice-caps reached the shores and flanks of the Solimões, Amazonas and Parnaíba basins, but in Frasnian times ice-caps were less developed. Beyond them glaciofluvial and aeolian sedimentation may have taken place. In the Amazonas Basin, beneath diamictite beds silty shale beds are present in the Upper Barreirinha and Lower Curiri formations. The great amount of silt might have originated from strong wind activity upon glacially derived material.

Siltstone deposits as thick as 300-400 m were deposited in Adrar of Mauretania, Northwest Africa, in Frasnian time (Nahon and Trompette, 1982). These sediments are interpreted by me as loess resulting from glacial grinding during the Late Devonian in northern South America and probably in central Africa. This conclusion is contrary to that of Nahon and Trompette (1982) who explain these siltstone as generated in tropical zones where weathering processes are very active.

11. Continental reconstruction, based on new paleomagnetic data, shows northern Brazil at high latitudes from Late Ordovician to

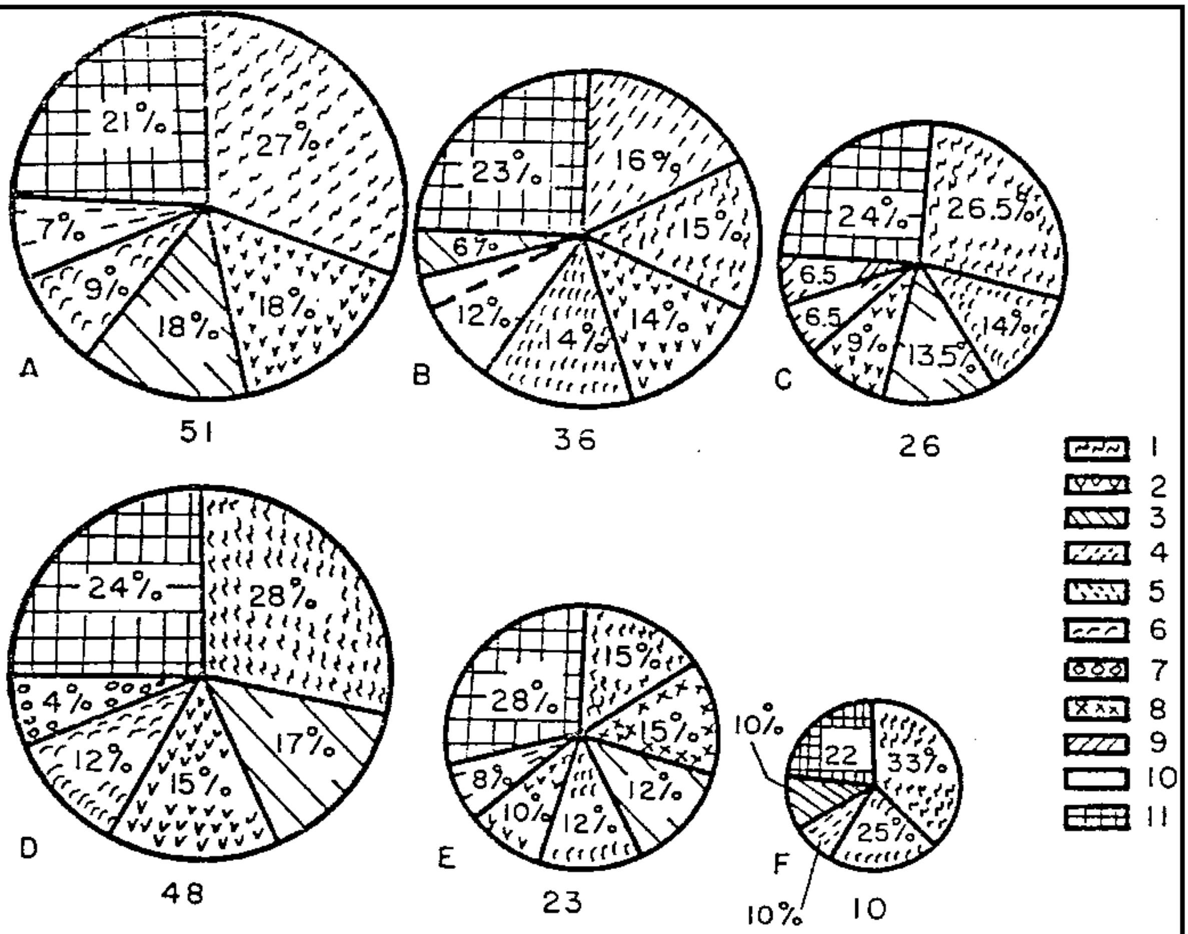

Fig.91 Relationships between large taxonomic groups in biocoenoses according to their species number at Heis Islands (A), Victoria Island (B), Alexandra Land and Rudolf Island (C) on stony and silted grounds (Franz Josef Land), west off De Long Islands, on muddy grounds (D); at Stolbovoj and Belkovsky Island (the Novosibirskye Islands), on stony grounds (E), in the Southern part of the Laptev Sea on silted grounds (F). The Diagram area is proportional to the average number of species in ecosystems. 1 - Polychaeta, 2 - Gastropoda, 3- Bivalvia, 4 - Bryozoa, 5 - Hydroidea and Actiniaria, 6 - Crustacea, 7 - Holothuroidea, 8 - Spongia, 9 - Ophiuroidea, 10 - Algae, 11 - other taxonomic groups. Afer Golikov and Averincev, 1977.

Early Carboniferous time (Smith and others, 1981). Particularly during Silurian (Figure 88) and Devonian (Figure 89) Periods, the Bolivia-Argentine border area and the central part of Brazil was respectively located at the South Pole. The paleomagnetic evidence is consistent with the existence of glacigenic beds since Late Ordovician to Early Carboniferous time in northern Brazil, however, the best glacial imprint was left during the Late Devonian time when central Brazil was located closer to the South Rotational Pole.

It seems necessary that the Earths pole must be sufficiently inland before ice sheets can be developed, and when, the pole lies only on the edge of a continental block oceanic circulation is adequate to maintain equable conditions on that continent (Tarling, 1978).

12. Continental reconstruction based on lithic data suggests that South America was located at a high latitude in Devonian time. The absence of warm and hot lithic climatic indicators in the diamictite units, as well as in adjacent formations suggests that deposition occurred in cold climate.

Andrade and Daemon (1974) alleged that the biologic content of the diamictite-bearing formation is incompatible with glaciation in the Parnaíba Basin in Late Devonian time. However, Golikow and Averincev (1977) investigated present organism distribution in polar oceans. They observed the presence of a great number and low diversity of adapted invertebrates directly living in contact with and beneath ice packs (Figures 90, 91).

Copper (1977) stressed that the fossil assemblage found in the Amazonas and Parnaíba basins are depleted in tropical taxonomic groups due to very cold climatic conditions in Early to Late Devonian time. He recognized a climatic deterioration in Brazil in approaching the Late Devonian time as being responsible for the worldwide Famennian biotic massive extinction. The Malvinokaffric cold fauna was dominated by groups of organisms that survived the worldwide Late Devonian mass extinction. Invertebrate groups restricted to Devonian equatorial belts of North America, Eurasia and Australia were decimated in the Late Devonian (Copper, 1977). Growing cold spells may have eliminated organism communities in tropical areas. Subsequent repopulation of the Carboniferous seas was accomplished by hardy, eurythermal (tolerant of considerable difference in temperature) invertebrate taxa present in cold and as well as tropical regions (Copper, 1977). Therefore, some members are adaptable, and can survive and flourish under the new range of environmental conditions. So the arguments presented by Andrade and Daemon (1974) that the fossil record of the diamictite-bearing formation is incompatible with the presence of glaciation are untenable.

CHAPTER 13. MIGRATION OF GLACIAL CENTERS ACROSS GONDWANA DURING PALEOZOIC ERA

OVERVIEW

Different Paleozoic glaciations upon Gondwana are reviewed here in order to recognize any relationships between their development. Because Paleozoic glacial strata occur in different places at different times the question arises as to why these glacial centers change with time. Paleozoic glacigenic beds of other Gondwanan basins are therefore here reviewed according to their age and geographic distribution in order to understand the pattern of ice-center changes and to relate the Brazilian glacigenic rocks to this pattern.

Much new information has been obtained recently concerning the depositional environment and age of Paleozoic strata occurring in separate areas across the wide reaches of Gondwana. Since continental glaciation was first documented in India (Blanford and others, 1856), enough stratigraphic and tectonic information has now accumulated so that generalizations concerning Paleozoic world geography and climate are feasible.

Moreover, with the acceptance of the concept of the drift of Gondwana and its later fragmentation beginning in the Mesozoic Era, reconstructions of the huge Gondwana supercontinent and the distribution of sedimentary facies upon it and around its margins are meaningful (Crowell, 1981). Any satisfactory reconstruction of past geography, however, depends upon fitting together data of many types from many sources, and applied to a limited time interval, so that correlation precision lies at the heart of such efforts. Glacial sedimentary facies are both useful in correlation and provide much information of paleogeographic and paleoclimatological significance. This study adds new observations and interpretations pertaining to Late Devonian glaciation and reviews Paleozoic stratigraphy related to glacial rocks with an emphasis on Early and Middle Paleozoic strata of South America and Africa.

Time designations used for the Paleozoic glacial facies traditionally depend on sparse fossils (both marine and nonmarine), correlation based on physical properties, and on stratigraphic sequence and position. Only rarely are there dates available from geochronology, magnetostratigraphy, seismostratigraphy and other methods. In the absence of near continuous outcrops and corroborated ages, judgments are needed to date the beds at many places. What is a reasonable rate of facies change within a sedimentary basin, both laterally and vertically? How long did it take to deposit a given stratigraphic section? In order to elucidate the earth's past history one can only work with the data available and employ applicable methods, although one must be on guard against circular reasoning in regard to both age and facies assignment. In examining the glacial and periglacial deposits and their geographic distribution through time it is concluded that glacial centers migrated systematically.

In brief, Late Precambrian and Cambrian tillites and associated glacigenic sediments have been described in northwest Africa, in small basins around the northeastern margin of the Western African craton, although their precise age based in geochronologic data is controversial (Deynoux, 1980; Deynoux and others, 1978; Caby and Fabre, 1981a,b). Late Ordovician to Early Silurian glaciation has been well documented in North Africa, South Africa and South America (Beuf and others, 1971; Rust, 1981; Crowell and others, 1981). Evidence of the Late Devonian glaciation is now confirmed in northern South America and Early Carboniferous glaciation proved in western South America is now also considered as established in northern South America. Late Carboniferous to Mid-Permian glaciation is widely preserved on all of the continents that once constituted Gondwana (Crowell and Frakes, 1975). In this study the Late Ordovician-Early Silurian and Late Devonian-Early Carboniferous glaciations in northern Brazil are primarily dealt with because new data are added by me. Only secondary attention is given to the Cambrian and to the Late Carboniferous to Middle Permian glacial records because they have been treated elsewhere.

LATE ORDOVICIAN-EARLY SILURIAN GLACIATIONS

Ordovician-Silurian glaciation was first identified in the Table Mountain Group, Cape Ranges, South Africa (Rogers, 1902, 1904; Rust, 1973, 1981). In North Africa, Debyser and others (1965) presented incontestable evidence of a glacial event in the Sahara region although Lulubre (1952, in Freulon, 1964) had already suggested glaciation on the basis of rock texture in Ordovician-Silurian strata. The paper by Debyser and others (1965) was followed by several publications documenting glaciation in north and northwest Africa. More recently, complete reports have been provided by Beuf and others (1971), Fairbridge (1974), Deynoux (1980), and Biju-Duval and others, (1981). In South America, beginning with Schlagintweit (1943), a glacial imprint within Silurian strata has been recognized (Crowell and others, 1980, 1981). On the following pages some of these stratigraphic sections in Africa and South America are commented upon briefly.

ARABIA AND NORTH AFRICA

Arabia. Upper Ordovician glacial strata lying upon the northeastern side of the Precambrian Arabian craton, have recently been described by McClure (1978) and McClure and Young (1981).

Here diamictites are present in the Tabuk Formation at the top of the Ra'an Shale Member, and occur as discontinuous bodies up to 3 m thick. The diamictite is poorly sorted and contains sedimentary, igneous and metamorphic clasts, supported in a clayey and sandy matrix, with some clasts striated, faceted and polished. An unconformity is recognized between the shale and overlying coarse sandstone beds where paleovalleys filled with sandstone beds are preserved, some of which interfinger with the diamictite beds. Paleovalleys also contain rhythmites with isolated stones and sandstone bodies that are interpreted as varves and kettle hole deposits, respectively.

Striations on boulders embedded in the diamictite, which may constitute a boulder pavement, show a NNE orientation. According to McClure (1978) the diamictite beds are tillites and the sandstone beds were probably laid down as glaciofluvial outwash deposits. Such evidence for glaciation extends for more than 150 km along the NW trending strike of the formation.

On the basis of graptolites, trilobites and chitinozoans, the Ra´an Shale underlying the tillite is considered to be Late Caradocian (early Late Ordovician) in age, although an Early Ashgillian (late Late Ordovician) age cannot be ruled out. The sandstone unit overlying unconformably the tillite is inferred as deposited in early Early Llandovery (Earliest Silurian) time

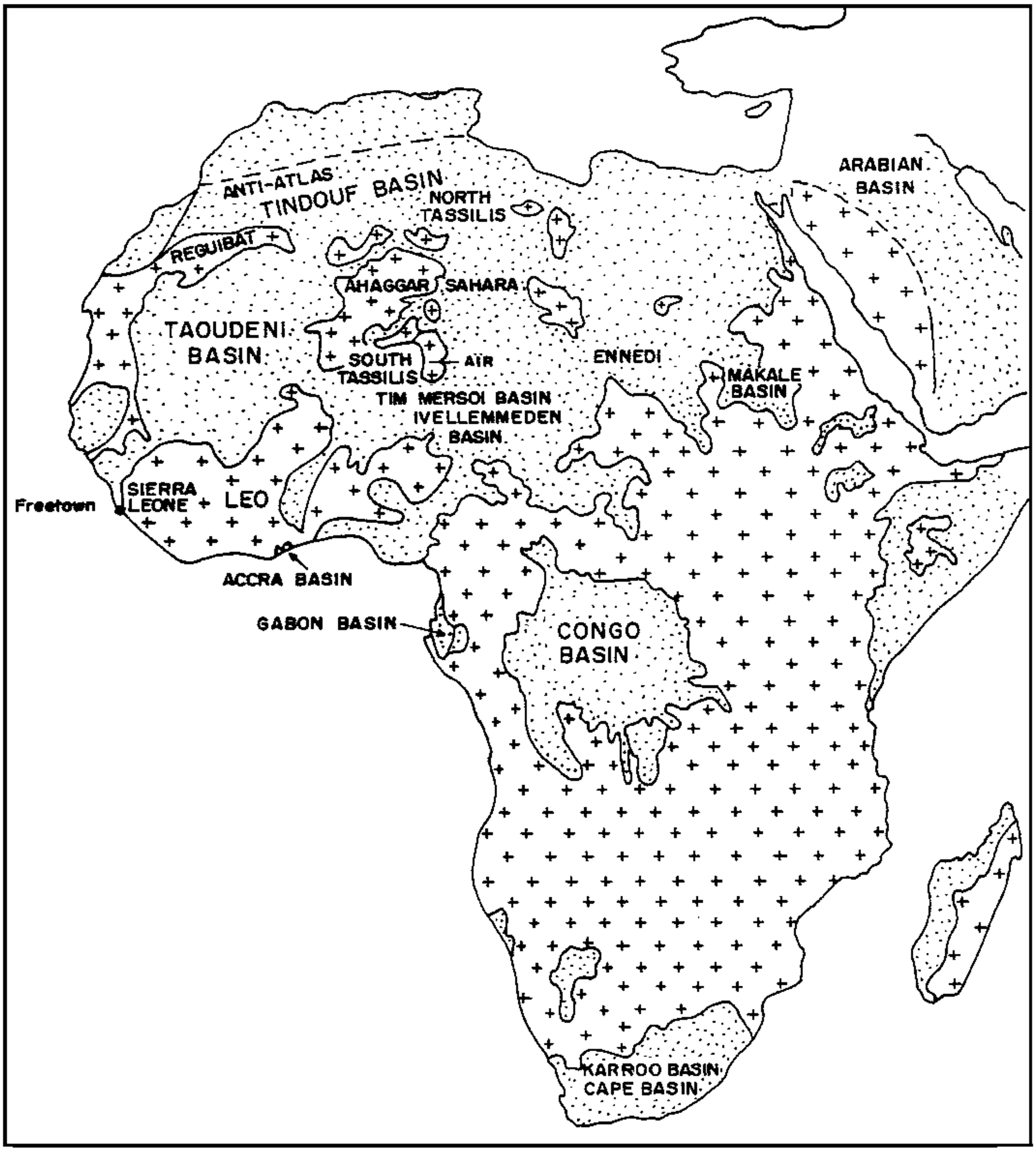

Figure- 92. African Paleozoic Basins indicated as discussed in the text.

because the Qusaiba shale overlying the sandstone beds contains graptolites, chitinozoans and acritarchs of Mid-Llandovery age (Idwian substage) (Early Silurian). Late Ashgillian (Uppermost Ordovician) beds are missing.

The presence of rocks of the basement and the striation directions on boulder pavements suggest that ice movement was in a general NNE direction from the craton towards a shallow continental shelf or sea. The unconformity at the Ordovician-Silurian transition may have resulted when oceanic waters receded during glaciation or to the formation of a forebulge in the area related to a major ice load during glaciation maxima in southern Saharan region. Figure 92 shows African Paleozoic basins as discussed in the text.

Central Sahara Basin. Descriptions of Ordovician glaciation in the Saharan region are presented by Beuf and others (1971), Bennacef and others (1971), and Biju-Duval and others (1981). The Tamadjert Formation, including glacial deposits, occurs over a large area in the Central Sahara region, around the margins of the Hoggar and Tibesti massifs (Touareg and Toubou shields respectively). These beds crop out in Algeria, Mali, Libya and Chad at the top of a Cambro-Ordovician clastic sequence which overlies unconformably Pan-African fold-belt rocks.

The Tamadjert Formation ranges in thickness from 10 to 450 m and has at its base either an angular unconformity or a disconformity. Its upper limit is also unconformable. The unit consists of diamictite, siliceous shale and siltstone containing striated quartzite boulders and basement pebbles that fill U-shaped paleovalleys, with polished and striated floors at places. Intraformational erosion surfaces are exposed at several stratigraphic levels. Features supporting a periglacial origin for some of the facies include deformed sandstone bodies and varve-like clays with rhythmic bedding containing dropstones.

Glacially striated pavements are exposed for many kilometers and at many places displaying chatter marks, crescentic gouges and crescentic fractures and showing ice motion from south to north. These pavements are overlain by diamictite with a sandy matrix, considered to be tillite. The deformed sandstone beds are interpreted as collapsed outwash sandstone beds overlying bodies of stagnant and melting ice. Patterned ground is locally preserved where former ice wedges are now filled with sand (Biju-Duval and others, 1981). A northward slope into the basin is suggested by an increase in sediment thickness and also by the facies distribution where massive tillites change laterally into probable glaciomarine deposits.

Boucot (1982, personal communication to J. Crowell), upon examining graptolites and brachiopods, stressed that the earliest part of this glaciation is recorded in the Caradocian (early Late Ordovician) of Libya and its latest part occurs in the Middle Llandoverian (Mid-Early Silurian) of Algeria with a maximum in the Ashgillian (late Late Ordovician). The present known extent of the former Saharan ice-sheet stretches 4,000 km from near the Red Sea to the Spanish Sahara on the west and 2,000 km from the Anti-Atlas in the north to southern Mauritania on the south (Tucker and Reid, 1973). The main ice spreading centers may have been located in a region south of the Saharan area and moved towards a broad shelf to the north and northwest and also on the Arabian craton where ice moved toward the northeast.

Tindouf Basin. Upper Ordovician sediments of possible glacial origin have been described by Destombes (1968, 1976, 1981) in the Anti-Atlas region, although he stresses that as yet no detailed studies have been carried out. Along the length of the Anti-Atlas chain and towards the south, the possible glacigenic Upper Second Bani Sandstone Formation up to 50 m thick rests unconformably on Precambrian rocks and strata that range in age from Arenigian to Ashgillian (Middle to Late Ordovician). it underlies disconformably Lower Silurian deposits.

The formation consists of white and pink massive and cross-stratified sandstone beds with conglomerate interbeds characterized by chaotic and heterogeneous structure. Diamictite containing pebbles of limestone, granite, rhyolite and bryozoan fragments occurs at the base and at the top of the section. Evidence for glaciation is based on the rock texture and on a single striated pavement underlying diamictites. The diamictites may be tillites, and conglornerate and cross-stratified sandstone beds may constitute periglacial outwash deposits.

Destombes (1981) considers the age of the diamictites, based on brachiopods, as Latest Ordovician (Late Ashgill: Hirnantian Stage) and the age of the overlying Ain Deliouine Formation, based on graptolites, as Early Llandoverian (Earliest Silurian). A slight time gap exists between the Upper Second Bani and the Ain Deliouine Formation.

Taoudeni Basin. Uppermost Ordovician glacial beds are widely exposed in the Taoudeni Basin and in the Upper Paleozoic West African mobile belt (Mauritanides). Deynoux and Trompette (1981) proposed the name Tichit Group as a general designation for the Upper Ordovician glacial formations to encompass several local names in use in the Taoudeni Basin. The Tichit Group rests unconformably on Cambro-Ordovician rocks and underlies conformably the Aratane Group composed mainly of marine shale of Siluro-Devonian age (Deynoux, 1980).

Lower units of the Tichit Group consist mainly of massive sandstone, clayey sandstone, and poorly sorted argillaceous conglomeratic sandstone and diamictite beds that fill deep paleovalleys. Upper units consist mainly of well-stratified fine- to medium-grained feldspathic sandstone and diamictite interbedded with clay-rich fine-grained sandstone. A distinctive laminated shale with scattered sand- to boulder-sized clasts, some striated, occurs at the top of the section and contains graptolites (Deynoux and Trompette, 1981).

Beds interpreted as tillites consist of green or purple, clayey, poorly sorted, massive sandstone with scattered sandstone and granite boulders, some striated. Many tillites are bedded and carry a higher percentage of clay and mica and a lower percentage of clasts.

According to Deynoux and Trompette (1981), sandy tillites are terrestrial, whereas those that are muddy and laminated, are generally laid down in water. Most of the deposits consist of cross-stratified or regularly bedded sandstone layers filling paleochannels. These are interpreted as periglacial sediments or outwash fans similar to Icelandic sandur formed when stream waters rework glacigenic deposits. Several glaciated pavements and "roches moutonées" (sheepback) underlying tillites occur but periglacial features such as polygonal structures with sandstone wedges are unknown.

The age of the base of the Tichit Group is not known because its lower part is unfossiliferous and it rests unconformably on

Upper Cambrian-Lower Ordovician rocks (Deynoux and Trompette, 1981). The age of its upper part is as old as Late Ashgillian on the basis of graptolites (Deynoux, 1980).

Striated pavements and "roches moutonées" indicate a NW or WNW direction of ice movement in the Taoudeni Basin (Deynoux and Trompette, 1981). In the entire area from the Arabian craton to the Taoudeni Basin, ice movement directions spread out almost 180º in northerly directions, suggesting the presence of major spreading centers on the south near the Gulf of Guinea (Deynoux and Trompette, 1981) or near the eastern tip of South America in a pre-drift reconstruction of Gondwana.

Tucker and Reid (1973) described laminated rhythmic clay with dispersed clasts interpreted as ice rafted dropstones in the Waterfall Formation of Sierra Leone. Although they were convinced of glacial origin for the beds, no striated surfaces or boulders were discovered, nor were fossils found.

A Middle to Late Ordovician age was assigned solely on litho-logic correlations with strata occurring in Guinea and Senegal where the beds are given an age of Middle to Late Ordovician. Turbidite features within units of the Waterfall Formation have not been observed elsewhere in Upper Ordovician glacial units of Waterfall Firmation, and so more studies are needed to determine the true depositional environment as well as their age.

Accra Basin. In comparison with the huge Taoudeni Basin (more than 2,000,000 km^2) the Accra Basin as now exposed is quite small (200 km^2). As a consequence it is difficult to find sedimentary features sufficient for paleoclimatic interpretation. Within the basin, Paleozoic strata comprising the Asemkaw, Ajua and Elmina formations lie in small isolated areas along the coast of Ghana and may have been deposited in glacial and periglacial environments as suggested by Junner (1939) and Crow (1952), but disputed by Anderson (1961) and Talbot (1975, 1981).

The Asemkaw Formation consists of a basal bed up to 4 m thick, composed of large rounded or subangular boulders derived from the underlying basement immersed in a matrix ranging from clay- to granule-size. Above this there are cross-laminated and cross-stratified sandstone beds with some conglomerate interbeds. The Ajua Formation consists of finely interlaminated, ripple-marked, micaceous feldspathic sandstone and siltstone, fissile shale, and diamictite. The Elmina Formation, the youngest unit, consists of coarse-grained sandstone with conglomerate and mudstone interbeds. At the top of the section there is a breccia horizon with clasts up to several feet in diameter (Crow, 1952). In some places, the Elmina Formation has cut through the underlying formations and rests directly on basement rocks. It is not known whether the Asemkaw and Ajua formations pinch out against basement or whether the Elmina Formation fills large paleovalleys.

Scattered pebbles and boulders up to 30 cm across occur throughout the Ajua Formation, but no striated clasts have been identified. The clasts are composed of granite, quartzite, quartz, tuff, diorite, lava, sandstone and conglomerate (Crow, 1952). Primary sedimentary structures are commonly disrupted by soft sediment deformation (Crow, 1952; Talbot, 1981).

Crow (1952) stressed that beds are disturbed beneath large pebbles and boulders, but Talbot (1981) pointed out that the deformation is only the result of compaction. On the other hand, Talbot (1981) interpreted blade-shape structures 1-5 cm long and O.1-0.3 cm wide filled with sand and developed in silt laminae as ice crystal pseudomorphs. He interpreted the Asemkaw formation as deposited by braided streams, where with periodic aereal exposures, the dry river sediments were redeposited by wind and the Ajua Formation as deposited in a shallow marine environment in which river or beach ice rafting took place. The overlying Elmina clastics is regarded as deposited in delta front fans and delta fans.

The Elmina sandstone was dated by Bär and Riegel (1980) as Latest Ordovician or Earliest Silurian on the basis of palynological studies. Its palynological content is similar to that of the Serra Grande Group of the Parnaíba Basin (Tianguá and Jaicós formations).

In a pre-fragmentation reconstruction of Gondwana, the Ghanaian outcrops of the Asemkaw, Ajua and Elmina formations fit against the Brazilian outcrops of the Serra Grande Group of the northern Parnaíba Basin. The Accra Basin may therefore have been a piece of the northern Parnaíba Basin of Brazil. Paleocurrent determinations suggest a generally north to south direction of sediment transport and soft sediment deformation features suggest a south or south-easterly paleoslope (Talbot, 1981). These directions agree with those for source areas for the northern Parnaíba Basin basal sediments.

Gabon Basin. For more than 100 km along the eastern margin of the coastal Gabon Basin, within a graben formed during Mesozoic times, the N'Khom Series of glacigenic deposits of possible Late Ordovician-Early Silurian age is resting unconformably on Precambrian and Early Paleozoic basement.

According to Micholet and others (1971) the N'Khom Series, 10 to 55 m thick, consists of a basal reddish brown diamictite with abundant clasts of gneiss, granite, and sandstone immersed in a sandy and clayey matrix. Some striated boulders lie within paleovalleys where "roches moutonées" and pavements with E-W striae are exposed. Overlying these beds, interpreted as tillite, are clay layers with rhythmic bedding and with dispersed angular or rounded rock fragments. At places clay beds display slump structures, and above and laterally sandstone and conglomerate strata show fluvial characteristics.

The presence of diamictite with striated clasts overlying striated pavement and their association with rhythmites containing outsized rafted pebbles are the main arguments for interpreting the N´Khom diamictites as glacigenic (Rocha-Campos, 1980).

The Bekang Member overlies the N´Khom Series and consists of cross-stratified, poorly sorted, coarse-grained sandstone and conglomerate beds interpreted as deposited in outwash streams, overlain in turn by black shale of the Agoula Series (Hambrey, 1981).

Micholet and others (1971), because the overlying Agoula Series has a Late Carboniferous or Early Permian age, based on palynological grounds, suggested a similar age for the N´Khom Series. This age determination may be incorrect, however, because there is an implied great difference between the cold climatic indicators for the unfossiliferous N'Khom Series as suggested by the tillites, varve-like sediments with dropstones and outwash deposits and warm climatic indicators for the overlying Agoula Series, such as black and red shale, dolomite, gypsum, anhydrite and cherty limestone. There may be an unrecognized unconformity between them.

In Gondwanan reconstruction, the pull-apart Gabon Basin fits next to the pull-apart Sergipe-Alagoas Basin of Brazil (Schaller, 1969). The N'Khom Series correlates with the Brazilian Mulungu Formation of the Sergipe-Alagoas Basin and the Ipu Formation of the Serra Grande Group (Parnaíba Basin) based on lithologie similarities and stratigraphic position. The Agoula Series of Gabon correlates with the upper Batinga Formation (Boacica Member) of the Sergipe-Alagoas Basin based on lithologic similarities and palynological data. Moreover, surface striations and paleovalley directions measured by Micholet and others (1971) indicate an east to west direction of glacier transport, a direction in agreement with the projected directions to the source areas for the lowermost sediments of the Sergipe-Alagoas Basin.

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Sergipe-Alagoas Basin. Probable Late Ordovician to Early Silurian deposits crop out over an extent of 70 km by about 10 km in the Sergipe-Alagoas Basin of northeastern Brazil, (Schaller, 1969; Rocha Campos, 1981a).

The stratigraphic section consists of the Batinga Formation, composed of three members. The basal Atalaia Member up to 25 m thick fills paleovalleys and is made up of feldspathic, immature, white, coarse-grained cross-bedded sandstone. The overlying

Figure-93. South American Paleozoic Basins indicated as discussed in the text.

Mulungu Member consists of gray or maroon sandy diamictite with clasts of dispersed fragments of quartzite, quartz, granite and mica-schist, some of which are faceted and striated (Rocha-Campos, 1981a) and with shale interbeds. These beds grade upwards and laterally into finely-stratified pebbly sandstone or tillite and rhythmically bedded strata charged with centimeter-sized clasts.

The Precambrian and Lower Paleozoic basement beneath the Batinga Formation shows parallel elongate, polished and finely-striated low bosses (Roche moutonée), oriented 135º (Rocha-Campos, 1981a). The uppermost Boacica Member consists of laminated siltstone and sandstone beds up to 170 m in thickness.

The existence of tillites with striated clasts and associated rhythmites directly overlying polished and striated basement surfaces are the main evidences supporting a glacial origin for the Mulungu Member (Rocha-Campos, 1981a). The Atalaia Member may have been deposited in proglacial or outwash streams occupying paleovalleys before expansion of glaciation or during previous interglacial stages.

Controversy over the age of these strata exists. The unit was considered to be of Late Carboniferous or Early Permian age on the basis of spores Florinites found in the upper Boacica Member of the Batinga Formation (Uesugui and Santos, 1966).

In the adjacent Jatobá and Parnaíba basins, Late Carboniferous sediments (Curituba and Piauí formations) are interpreted as deposited in warm and hot climate as indicated by red fluvio-aeolian, marine limestone and evaporite interbeds. Although these beds are not yet satisfactorily dated it is speculated here that the Brazilian units (basal Atalaia and Mulungu members) correlate with the lower part of the Sekondi Series (Asemkaw and Ajua formations) of Ghana, and with the N'Khom Series of Gabon. These, in turn, may correlate with the Ipu Formation of the Serra Grande Group, Parnaíba Basin, Brazil. The overlying rocks are dated as Middle Devonian for Ghana (Accra Basin), Late Carboniferous for Brazil (Sergipe-Alagoas Basin) and Early Permian for Gabon. If the Atalaia and Mulungu members of the Batinga Formation are as old as Ordovician-Silurian they should be separated from the Boacica Member and raised to the rank of formation.

In a Gondwanan reconstruction, South America has to be rotated 45º counterclockwise with respect to Africa in order to fit properly. If this rotation is taken into account, the 135ºN orientation determined by Rocha Campos (1981a) for striae on the basament of the Sergipe-Alagoas Basin changes to an E-W direction; that is, the same direction as that found by Michelet and others (1971) on the floor of the Gabon Basin paleovalleys. Although the sense of glacier movement is not known in Brazil, paleovalleys draining westward from the Gabon Basin would account for the great volume of sediments laid down in the Parnaíba trough. Figure 93 shows South American Paleozoic basins as discussed in the text and Figure 94 shows the stratigraphic columns of these basins.

Parnaíba Basin. The Late Ordovician to Early Silurian Serra Grande Group (200 to 900 m thick) is exposed in the outcrop belt marginal to the Parnaíba Basin, Brazil, and in the area between the eastern Parnaíba outcrop belt and the eastern Atlantic coast. In this area 600 km wide, several small patches of the Serra Grande Group are apparently preserved within down-faulted blocks as at Jatobá, Mirandiba, Rio do Peixe, and in the Cariri Valley, all areas located in the states of Maranhão, Ceará and Pernambuco (Mabesoone, 1977, 1978). The Serra Grande Group rests unconformably on Precambrian and Lower Paleozoic rocks and is composed of three formations.

The lower (Ipu Formation) is mainly made up of coarse-grained, massive and cross-bedded sandstone, conglomeratic sandstone, conglomerate and diamictite at the top, with soft sediment deformation structures in some places. In the Cariri Valley the unit is in part composed of dispersed rounded and angular pebbles and cobbles, derived from igneous and metamorphic terranes, and supported by a sandy and clayey matrix (Mabesoone, 1977, 1978). The clasts found in the outcrop belt consist mainly of quartz and quartzite along with some blocks of granite, gneiss and slate. Some clasts are faceted but striated fragments were not noted (Kegel, 1953). In the outcrop belt diamictite horizons with striated boulders cap coarse sandstones, conglomeratic sandstones and conglomerates in the region of Ipueiras. It was inferred a Late Ordovician to Earliest Silurian age to the Ipu Formation because it is overlain by apparently conformable Early Silurian beds.

The middle formation of the Serra Grande Group (Tianguá Formation) consists of interbedded marine shale, siltstone and fine-grained sandstone. The unit exposed only in the northeast part of the basin, but in the subsurface may reach about 250 m in thickness and wide distribution. The upper unit (Jaicós Formation) is composed of cross-bedded sandstone, a few conglornerate beds and infrequent shale beds. Its upper boundary is difficult to locate due to lithological similarities with overlying Devonian strata.

The tillite-like rocks make up a small percentage of the total thickness of the Ipu Formation, and the major part of the section probably consists of glacigenic deposits reworked by periglacial outwash streams. The presence of faceted and striated stones in diamictite beds, a huge number of clasts, a high proportion of angular sand-sized quartz and feldspar grains, and low proportions of shale beds intercalated in sandstone beds, is consistent with a fluvioglacial origin. I found several boulders 20 to

30 cm in diameter in the outcrop belt of the Ipu Formation. These may be stones left behind by strong floods, perhaps jokulhlaups, when glacial erratics were carried from a wasting ice front out upon an outwash plain, such as a sandur (Bluck, 1974).

The Tianguá Formation was deposited under marine conditions following the Early Silurian transgression, interpreted as due to glacier melting. The Silurian age of the Tianguá Formation was determined by Müller (1962) and Carozzi and others (1975) on palynological grounds (chitinozoans and acritarchs).

Schermerhorn (1971) contested the presence of ice sheets in northwest Africa in Late Ordovician time. Among questions he raised was: Why did no corresponding glacial centers develop in the originally adjoining parts of the Americas that also were situated in polar zones? This study tentatively concludes that ice sheets did develop in adjoining northeastern South America.

Amazonas Basin. The Nhamundá Formation, of the Upper Ordovician-Lower Silurian Trombetas Group of the Amazonas Basin, northern Brazil, contains diamictites which are interpreted as glacigenic (Caputo and Vasconcelos, 1971 (Photo 2); Caputo and others, 1971, 1972, and Rodrigues and others, 1971). The Trombetas Group is composed of four formations, from bottom to top: Auta's-Mirim, Nhamundá, Pitinga and Manacapuru.

In regard to age, the Pitinga Formation is dated as late

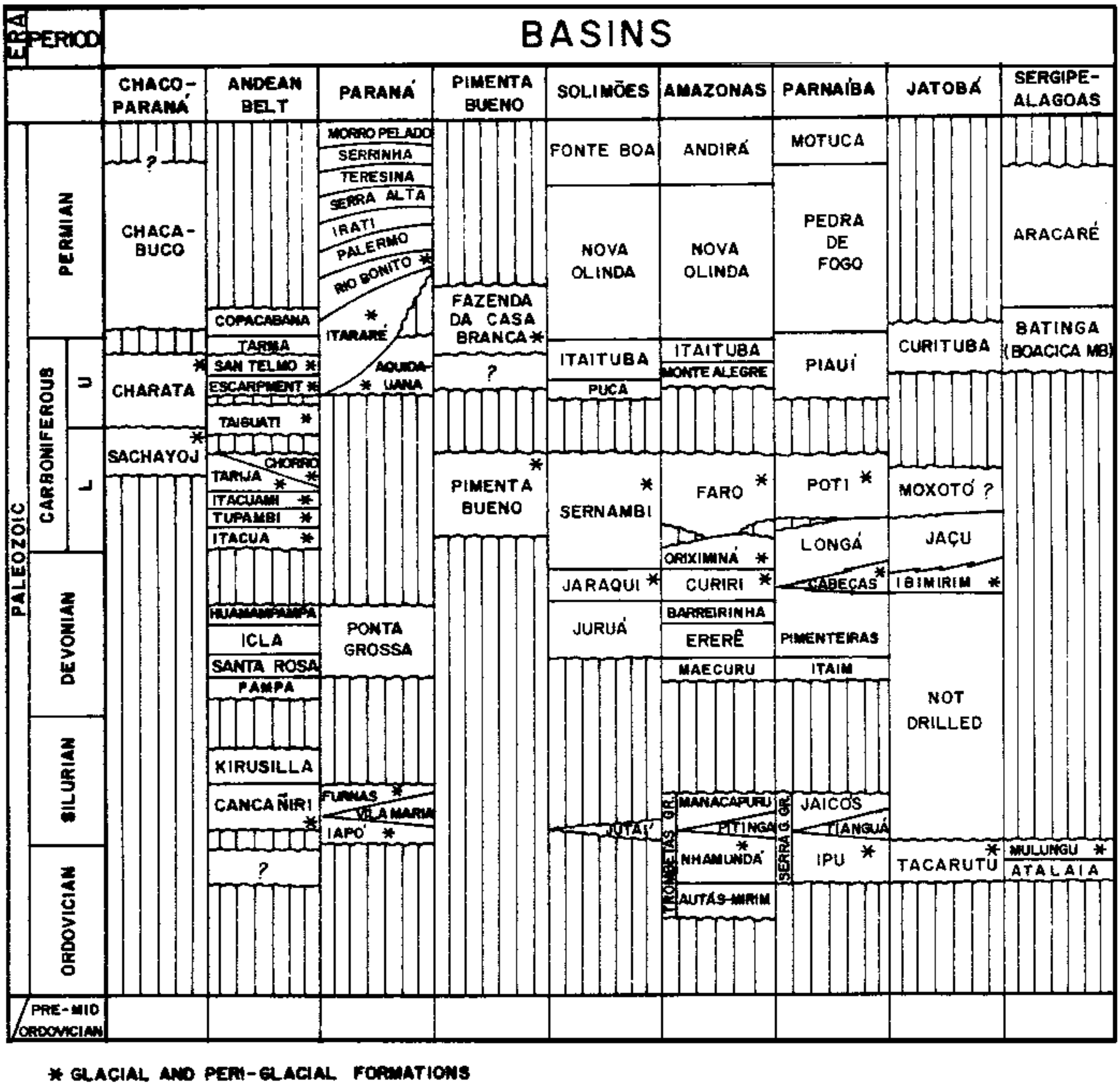

Figure 94. Time - stratigraphic correlation chart of South American Paleozoic sediments of basins discussed in the text.

Early Llandovery to Middle Llandovery based on graptolites such as *Climacograptus innotatus* Nicholson var. *brasiliensis* Ruedemann. An early Early Llandovery age is assigned to the uppermost Nhamundá diamictite and an Ashgillian age is inferred for the two lower ones.

The Nhamundá Formation consists of white and light gray, fine-grained sandstone with three diamictite horizons (Rocha-Campos, 1981d). The widespread unit crops out in the northwest part of the western Amazonas Basin, but it subcrops the entire basin and is cored in wells more than 150 km from the present surface outcrop position of basement.

The upper diamictite, 10 to 15 m thick, is best known and contains coarse sand grains and rounded or angular clasts of rhyolite, quartzite, and sandstone up to 6 cm in diameter, supported within a massive micaceous, sandy, silty, and clayey matrix, gray to brown in color. Chlorite is dominant over illite in the matrix. Since information concerning it comes primarily from few well cores such features as striations and facets are difficult to recover, but in geological surveys they were found (Caputo and Sad, 1974).

The widespread Nhamundá Formation has a dip of about 1/2º after removal of post-Silurian deformation due to subsidence, and available isopachs show regular thicknesses suggesting no tectonic activity, only restricted to subsidence after its deposition. Bioturbated horizons in the formation suggest a shallow marine environment, but the diamictite beds occur between cross-stratified fluvial beds above and below. The absence of graded beds and related sedimentary structures suggests that turbidity currents played no role in the diamictite deposition and ice becomes the most likely mechanism for carrying such a range of clast immersed in a fine matrix, in flat terrains for distances as great as 150 km.

The stratigraphic units above and below the diamictites, also deposited at times of tectonic calm, are free of warm climatic indicators and contain abundant mica, chlorite, illite, and feldspar minerals indicative of deposition under cold and dry climate.

Biological climatic indicators also suggest high latitudes of cold climate for North Africa during Late Ordovician (Spjeldnaes, 1961) and Early Silurian (Berry and Boucot, 1967) times, as well as the adjoining parts of northern South America.

In the Amazonas Basin as well as in the Saharan basins, Lower Silurian shale transgresses across marine sandstone beneath which glacial beds occur. The glacial horizons were deposited on both the northern and southern flanks of the Amazonas Basin so glaciers may have been developed on both the Guyana and Brazilian shields.

Paraná Basin. The unfossiliferous lapó and Furnas formations of probable Early Silurian age overlie unconfonnably the Precambrian to Mid-Ordovician basement in the Paraná Basin, southern Brazil, and are considered to be of glacial and periglacial origin respectively. The age of both formations is constrained by Late Ordovician rhyolites of the Castro Group below (450 ± 25 m.y.) (Bigarella, 1964, 1970) and by overlying late Early Devonian beds (Emsian) of the Ponta Grossa Formation which contain trilobites, brachiopods, pelecypods and chitinozoans.

In the State of Paraná, the lapó Formation is about 16 m thick along its outcrop belt (Rocha-Campos, 1981e), but it is more than 70 m thick in the subsurface where I have identified in a half-dozen wells (for example, well JT-1-PR). The unit, red diamictite in the outcrop area, but gray in the subsurface, is crowded with angular and subangular clasts, up to 25 cm in diameter, supported by a structureless sandy, silty and clayey matrix. Pebbles commonly consist of granite, gneiss, rhyolite and quartzite, derived from the underlying igneous and metamorphic basement.

Other clasts, such as sedimentary rocks, quartz, chert and phyllite appear in smaller percentages (Bigarella and Salamuni, 1967b). Some are faceted and striated (Maack, 1947).

Maack (1947, 1964) considered the lapó Formation to be a glacial deposit based on rock texture, the occurrence of faceted and striated pebbles, and because it has the same stratigraphic position as that of the South African glacial deposits in the Cape Ranges. Moreover, he also gave it a possible Silurian age in view of the same suspected correlation. Caster (1952), Rich (1953), Buerlen (1955), Martin (1961), Ludwig and Ramos (1965), also interpreted the Iapó Formation as glacial, but this interpretation was challenged by Bigarella and others (1966), who held that the sequence was deposited by successive subaerial mudflows, each several decimeters thick. Later Bigarella (1973a,b) suggested that the diamictites might be reworked periglacial deposits. This study considers the Iapó Formation diamictites to be tillites deposited directly from ice sheets.

The poorly fossiliferous Furnas Formation was traditionally considered as Devonian due to its position beneath the Devonian Ponta Grossa Formation, but du Toit (1927), Bigarella and others (1961, 1966), Bigarella and Salamuni (1967a,b), Bigarella (1973a,b) and Ludwig and Ramos (1965) suggested that this unit or part of it may be separated from the overlying Ponta Grossa Formation.

The Furnas Formation consists of fine-, medium- and coarse-grained, cross-bedded, feldspathic, white sandstone and conglomeratic sandstone with interbedded conglomerate at several levels. Angular to subrounded clasts may be as large as 20 cm across. Shale and siltstone intercalations are rare.

Maack (1947, 1964) studied the Furnas Formation and pointed out that sparsely distributed faceted and grooved quartzite clasts occur over a distance of 50 km from basement. The formation probably formed as periglacial outwash, but detailed studies of the unit are still lacking. The Furnas Formation may correlate with the uppermost Nardouw Formation of the Table Mountain Group of South Africa.

In the State of Goiás, central Brazil, north of the Paraná Basin, the Vila Maria Formation of Silurian age (De Faria, 1982) lies at the base of the Paleozoic section along the northern outcrop belt of the huge Paraná Basin. It consists of diamictite, shale, siltstone and sandstone, with diamictite and shale predominating at the base.

Diamictite beds are greenish gray, unsorted, with poly-mictic coarse clasts ranging from granule- to boulder-size (1 m across). The smaller clasts are mainly angular and the larger ones are chiefly subrounded. These clasts consist of granite, gneiss, quartz, quartzite, rhyodacite, phyllite, metamorphosed tuff and basic rocks that are supported by a sandy or silty and clayey matrix. Interbeds of fossiliferous, laminated, micaceous greenish shale up to 1 m in thickness also occur.

The diamictite section is overlain by laminated, micaceous, fossiliferous, greenish gray shale beds up to 30 m thick, with some siltstone interbeds. The upper part of the shale section is characterized by laminated siltstone with some sparse granule-sized quartz and fine-grained sandstone beds up to 10 m thick.

The shales contain brachiopods, pelecypods, gastropods, graptolites and the worm *Arthrophycus* sp. that indicate an Early Silurian age (De Faria, 1982). The large size of the gastropod genus *Plectonotus* may indicate very cold water milieu. The shale section underlying the Ponta Grossa Formation, described by De Faria (1982) was observed in many boreholes to thicken toward the well Known Paraguayan Silurian outcrops, in the western side of the Paraná Basin.

This study restricts to the part of the section consisting of the Vila Maria Formation shales and diamictites. The basal diamictites probably correlate with the Iapó Formation in the State of Paraná, with the Nhamundá Formation (Amazonas Basin) and with the Ipu Formation (Parnaíba Basin). The shale beds are viewed as correlating with the Pitinga (Amazonas Basin) and Tianguá (Parnaíba Basin) formations. The Furnas Formation correlates with the Manacapuru (Amazonas Basin) and Jaicós (Parnaíba Basin) formations. In Paraguay, a thicker similar stratigraphic section exists in the Caacupe Group, although diamictite or tillite have not been recognized yet.

De Faria (1982) interpreted the sedimentary sequence as deposited in a shallow sea influenced by tides and invaded periodically by debris flows. The transitional contact of the Vila Maria Formation with the overlying Furnas Formation suggested to him an evolution from coastal to fluviatile conditions. However, the diamictite beds are interpreted here as a result of glacial activity on the Brazilian and Guaporé shields. This is suggested by their stratigraphic position involving correlation with quite well established tillites in other mentioned basins. Stone types within the diamictites apparently match those within the shield regions.

The diamictite beds of the Vila Maria Formation are intercalated with shale of Llandoverian age, suggesting that the glaciation is largely Silurian in age, and also glaciomarine. Perhaps the shale beds were laid down during a marine transgression following ice melting also in central Africa. This Early Silurian transgression may have reached the Paraná Basin from the Paleopacific Ocean because the shales thicken toward Paraguay and paleocurrent studies of the overlying non-marine Furnas Formation suggest that supposed rivers emptied into a general western direction (Bigarella, 1973a,b), probably Paleopacific Ocean during a general regression that began in the Late Llandoverian time (late Early Silurian) in South America and Africa (McKerrow, 1979).

Andean Region. In the eastern Andes and Andean foothills, glacigenic units of Early Silurian age crop out along a belt about 1,500 km long from northern Argentina across Bolivia to Peru (Crowell and others, 1980, 1981). Names applied to these units include Zapla (Schlangtweit, 1943), Cancañiri or Sacta in Bolivia, San

Gaban in Peru, and Mecoyita in Argentina. Beds containing the glacial sequence overlie unconformably several different levels of the Ordovician system. They include structureless micaceous diamictite up to 60 m thick, intercalated with sandstone, conglomerate and marine black shale.

The diamictite varies from sparse-stoned to very stone-charged and it is highly fractured with incipient cleavage. Crowell and others (1981) report oversized stones up to 150 cm across, but in most exposures none is larger than 15 cm. The subangular to sub-rounded clasts consist of vein-quartz, granite, gneiss, quartzite, pink granite, grey pebble conglomerate, slate, sandstone and black shale. Some faceted pebbles and a few that are striated have been recovered. Slump structures are common in the central parts of the basin.

Glacial origin for some of these sediments rests primarily upon the occurrence of striated and faceted clasts and the wide extent of diamictites at the same stratigraphic level, and upon the inferred distant source for the stones which could most successfully be transported by ice. It is postulated that ice locally rimmed a marine basin, and that glaciers on the Brazilian shield to the east and on basement highlands to the south and west contributed debris (Crowell and others, 1980, 1981).

Overlying conformably marine deposits containing sparse graptolites as old as Late Llandovery, may indicate a Late Llandovery age (Berry and Boucot, 1972) for the diamictites. Underlying beds, at places, unconformably beneath the presumed glacigene strata, are mainly Caradocian, and the Ashgillian is largely missing. The glaciation here is therefore largely Silurian in age, and may not include the latest Ordovician.

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Cape Ranges. Glacigene beds occur in the Pakhuis Formation, Table Mountain Group, in the western part of the Cape Fold Belt and northward into the Cedarberg Mountain Range (Rust, 1981). This ice age has been named the Winterhoek Glaciation by Rust (1973). Immediately underlying the Pakhuis Formation, within the top of the Peninsula Sandstone Formation, complex soft-sediment folds are interpreted as the result of glaciers riding over unconsolidated strata. Discontinuous masses of the Sneeukop Tillite are lodged within the synclines and form pod-like cores, and clastic dikes and patterned ground are associated (Daily and Cooper, 1976). The Oskop Sandstone, 2 to 3 m thick, lies above these deformed beds and the patches of Sneeukop Tillite. The top of the Oskop displays a distinctive glacial pavement, which, in turn is overlain by the Steenbras Tillite. This tillite unit thickens irregularly to the south and grades laterally into the

Kobe Tillite. The Cedarberg Shale caps the tillite, and consists of fine-grained carbonaceous shale carrying isolated sand grains that are interpreted as ice rafted. A laminated sequence, varve-like in aspect, has been observed at one place.

An assemblage of marine brachiopods, bryozoans, trilobites, and crinoids of Llandoverian age has been recovered from part of the Cedarberg Shale (Berry and Boucot, 1973). The overlying Nardouw Formation caps the sequence and has the same characteristics as those of the Peninsula Formation.

Stones within the diamictite units range in size up to 50 cm and consist of many lithologic types, including chert, stromatolitic chert, sandstone, limestone, and shale that are probably derived from the Nama Group in Namaqualand and Namibia. One granitic clast, a few volcanic stones, white quartzite and vein quartz have also been recovered. Stones contained within the Kobe Tillite are notably angular to subangular and are commonly polished, faceted, and striated (Rust, 1973, 1981).

The underlying Peninsula Formation, depleted in mudstone beds, is interpreted as consisting of shallow shelf, littoral (Hobday and Tankard, 1978), and possibly fluvial deposits. Perhaps it was derived from the reworking of glacigenic material generated around the southern rim of huge ice-sheets occupying central West Africa during Late Ordovician time. Groove and striations, and kinematic interpretations from the shapes of soft-sediment folds, suggest flow of ice from the north and east.

LATE DEVONIAN GLACIATIONS

Documentation of Devonian glaciation comes largely from northern Brazil. Principal evidence for glaciation is very widespread diamictites in the Solimões, Amazonas and Parnaíba basins of northern Brazil. In each basin, sedimentary textures, structures, and depositional setting support interpretation that the diamictites are of glacigenic origin. De Moura (1938) first interpreted diamictites as deposited under glacial conditions in borehole cores from the Amazonas Basin, and Kegel (1953) those from cores from Carolina well (1-Cl-1-MA) from the Parnaíba Basin.

Other areas where probable Late Devonian glacial and periglacial deposits may have been laid down include the following: a) Famennian sediments of the deepest parts of the Jatobá Basin (Brazil); b) The Takoradi beds of Accra Basin, Ghana from which Bär and Riegel (1974) recovered diamictite of suspected glacial origin; c) The Teragh Formation of probable Late Devonian age of the Tim Mersoï Basin, Republic of Niger, that Hambrey and Kluyver (1981) demonstrated its glacial genesis; d) The Miller Diamictite Formation of the Kommadagga Subgroup of the Witteberg Group (Karoo Basin, South Africa) of possible glacial nature (Swart and Hille, 1982).

Solimões Basin. Late Devonian (Famennian) diamictites of the Jaraqui Formation up to 100 m thick occur only in the subsurface of the Solimões Basin (Upper Amazonas Basin) because Paleozoic formations are concealed and overlapped by Mesozoic and Cenozoic cover. Close to the margins of the Paleozoic basin, the Jaraqui Formation is truncated by a pre-Pennsylvanian unconformity. The diamictite unit lies between a lower sandstone unit and an upper shaley unit and appears conformable with them. It is hard, medium gray to dark gray and is non-stratified and very poorly or non-sorted, and contains angular to sub-rounded clasts ranging up to about 6 cm across as seen in a few borehole cores (Photo 1). gray
The clasts consists of quartz, quartzite,igneous, metamorphic and sedimentary rocks and show no preferred orientation. The structureless matrix is composed mainly of quartz, feldspar and mica of silt-size, and particles of clay-size. Clay minerals are chlorite and illite, and the rock is generally rich in organic matter, pyrite and siderite. In the lowermost part of the stratigraphic section, gray conglomeratic sandstone beds are intercalated with the diamictites and in the upper part black shale as thick as 20 m is present. Some of the sandstone beds display soft sediment deformational features.

According to Lima (1982, personal communication) from Petrobras paleolab the age of the diamictite based on palynological data is Mid-Famennian. That is the same as that determined for Devonian diamictites in the Amazonas and Parnaíba basins.

Amazonas Basin. Upper Devonian glacial strata have been described in the Curuá Group (Caputo and Vasconcelos, 1971). Two diamictite horizons were identified in subsurface in the Curiri Formation up to 300 m in thickness and one in the lowermost part of Oriximiná Formation (Photo 3). These formations dipped at about 7 m per km, or at about 1/3° when they were deposited.

The diamictite consists of massive micaceous clay with sand grains, granules and pebbles chaotically dispersed throughout (Caputo and Andrade, 1968). Some striated clasts were described from cores (Bouman and others, 1960; Rodrigues and others, 1971) and faceted and striated as well as polished clasts recovered from surface outcrops by me (Photos 3, 6, 7, 8, 13) and Macambira and others (1977). The clasts are angular to rounded, and consist of quartzite, vein quartz, shale, siltstone, diabase, rhyolite, igneous rocks and carbonates. Faceted and striated clasts were also dispersed in laminated shale in the southern outcrop belt (Macambira and others,1977) and in cores. Sandstone lenses intercalated within diamictite display folded and disrupted structures (Rocha-Campos, 1981b) and fine- to medium-grained sandstone sheets are intercalated between diamictite layers.

A glacial origin for the Late Devonian diamictites in the Amazonas Basin was first proposed by De Moura (1938) based upon the rock texture. Ludwig (1964) interpreted them as resulting from submarine slumping and turbidity currents, and explained the deformed sandstone bodies as convolute folds (Photo 9). Volcanic rock fragments and supposed volcanic glass-derived material in thin section led him to propose that volcanism may have caused submarine slumping and turbidity currents, but he did not document the existence of graded beds. Caputo and Vasconcelos (1971) and Rodrigues and others (1971) reinterpreted the diamictites as a result of glacial activity, and regarded the Curiri and Oriximiná diamictites as both terrestrial tillites and glaciomarine sediments. Striated and faceted clasts occur more than 150 km distant from present basement sources and very gentle dips, and include dropstones in marine laminated shale. Sparse volcanic fragments and volcanic-derived material may have originated from older Precambrian rhyolites within the basement (Uatumã Group) that was then transported by glaciers into the basin.

The age of the Curiri Formation, previously discussed, is Mid-Famennian on the basis of palynological studies.

Diamictite-bearing strata are spread over a distance of more than 2,000 km in the Solimões and Amazonas basins from close to the Brazilian-Colombian border to the rifted Marajó basin near the Atlantic Ocean.

Parnaíba Basin. Three Devonian formations (Pimenteira, Cabeças and Longá) within the Parnaiba Basin have been considered as in part glacigenic. Because the Devonian Cabeças Formation is lithologically similar to the Serra Grande Group, it was confused by Blankennagel and Kremer (1954) with this group of Ordovician-Silurian age. As a result Malzahn (1957), who correctly interpreted the beds as of glacial and periglacial origin, misplaced them stratigraphically; these diamictites actually belong to the upper part of the Cabeças Formation of Late Devonian age and not to the upper part of the Serra Grande Group of Ordovician-Silurian age. The Cabeças Formation sandstone pinches out between the underlying Pimenteira Shale and the overlying Longá Shale, but its diamictite beds continue at the same level and truncate major formation boundaries.

I found striated, faceted and polished clasts in the south of the town of São Francisco do Piauí and in the western part of the basin in the Tocantins River cutbanks (down and up river from the town of Pedro Afonso). In this region Barbosa and others (1966) had previously observed tillites intercalated in the supposed Pimenteira Formation.

The diamictite is composed of unoriented rounded to angular clasts of quartz, quartzite, sandstone, shale, granite, gneiss, and other rocks, some of which are striated and polished, supported in a micaceous silty and clayey matrix (Photo 5) or supported in a massive sandy matrix. Striations on a pavement show a N15ºE trend (Malzahn, 1957; Bigarella, 1973b; Rocha-Campos, 1981c) and faceted, striated, and polished stones occur within diamictite layers. I found a pavement located about 26 km from the town of Canto do Buriti (Photos 16, 17, 18) on the old road to são Raimundo Nonato and in the same place a silicified conglomerate boulder (Photo 15) derived from the basement is present. It appears to be a far travelled erratic left by glaciers.

In areas where the underlying sediments are sandstone beds, the diamictite tends to be sandy and massive. Rhythmically bedded siltstone and shale with scattered dropstones was recorded by Malzahn (1957) from field outcrops and by Carozzi and others (1975) (Photos 11, 12) from borehole cores. The dropstones show disruption of laminations at the bottom and draping of sediment over the top of stones. In addition, sands and deformed sandstone bodies are extensive in the Cabeças Formation, perhaps due to collapse when ice underneath melted.

The glacial horizons are situated in the biostratigraphic intervals VII and VIII (Famennian of the Upper Devonian) based on palynological data (Andrade and Daemon, 1974).

Data from many wells drilled by Petrobras, supplemented by mapping and palynological studies, demonstrate that supposed different diamictites occurring in three formations are actually situated at the same stratigraphical level because the tillites cross major formation boundaries. The tillites of the Parnaíba Basin correlate with the Devonian tillites of the Amazonas Basin.

Within the small rifted Jatobá Basin, located close to the northeast Brazilian coast, marine Paleozoic sediments indicate that strata of the Parnaíba Basin previously extended beyond the area of the Jatobá Basin and have been preserved in downfaulted areas. In addition, Late Famennian (upper Upper Devonian) sediments of the Ibimirim Formation (Juatobá Basin), which is equivalent to the Cabeças Formation, consist of conglomerate and sandstone with a clayey matrix. Some coarse-grained sandstone beds contain slump structures. The character and age of the strata in the well 2-IM-1-PE drilled in Ibimirim in State of Pernambuco (Ludwig, 1964; De Quadros, 1980), suggest that they were deposited under glacial and periglacial conditions. If this is correct, 1,500 km more can be added to the glaciated area, so that it extends from the rifted Marajó Basin, at the mouth of the Amazon River to the rifted Jatobá Basin, close to the eastern Brazilian coast.

NORTH CENTRAL AFRICA

Tim Merso´i Basin. Along the eastern margin of the Tim Merso´i Basin the probable Late Devonian Teragh Formation with glacial characteristics overlies Precambrian igneous and metamorphic basement of the Aïr, north of the city of Agadès, northern Niger. The glacial origin of these sediments was demonstrated by Valsardieu and Dars (1971) and described by Hambrey and Kluyver (1981) upon whose papers this summary is based.

The stratigraphic section consists of the Terada Series with the· Téragh Formation at the base, which carries the glacial imprint, and overlain with a slight unconformity by the Talach Shale. The Téragh Formation is about 200 m thick in the north, but it thins considerably southward. From the base upwards, it consists of polymictic conglomerate, with clasts of the basement, supported within a sandy and pelitic matrix (3 m thick), mudstone or diamictite with boulders up to 50 cm in diameter (2-3 m thick), rhythmically bedded siltstone and shale with dropstones (5 m thick), siltstone and brown mudstone beds with traces of plants (3 m thick), and cross-stratified, feldspathic, muddy sandstone carrying angular clasts.

In the area beneath the Téragh, the basement displays polished surfaces of quartzite and with "roches moutonnées", some with cirque forms and stepped troughs as the result of glacial plucking. Paleo-valleys several meters deep are filled with conglomerate beds of irregular thicknesses. Clasts within the Téragh Formation, ranging in size from granules to boulders, consist of many shapes and lithologic types such as granite, quartzite, quartz, rhyolite, metamorphic rocks, and siltstone. Some pebbles possess an alteration veneer and display concave angular depressions where they have been pressed together. These units are interpreted as lodgement tillite, overlain by glacial marine or lacustrine beds as shown by mudstone with outsized clasts and rhythmites with dropstones, and capped by glaciofluvial deposits at the top of the section. Synsedimentary deformation structures are explained as due to undermelting of ice bodies and glacial shoving.

Unfortunately, the exact age of the Téragh Formation is questionable. The overlying Talach Shale indicate, on the basis of brachiopods and bryozoans, an Early Visean age (mid Early Carboniferous) (Valsardieu and Dars, 1971). The question involves the time interval represented by the slight angular unconformity beneath it. Plant remains recovered from the Téragh Formaton suggests that it is younger than the Ordovician-Silurian glacial sediments of northern Africa because land plants are unknown in Silurian time. The formation is therefore of either Visean or Tournaisian age (if the unconformity represents a short time lapse), or it may be of Late Devonian age and correlates with the Cabeças Formation from Brazil. Moreover, there are unconfirmed indications of similar glacial sediments of possible Late Devonian age about 100 km to the north (Hambrey and Kluyver, 1981). If future investigations show that the Téragh Formation is Late Devonian in age, more than 1,500 km may be added to the width of the Late Devonian glaciated area.

DISCUSSION

If all known and suspected Devonian glaciated areas are considered, extending from the Peru-Brazil border and Bolivia to northern Niger, the diameter of the region may have been more than 5,000 km. If the probable extension to Niger is excluded, it is about 3500 km wide (Figure 87).

Silurian and Devonian paleomagnetic poles (Figures 88, 89) were located in the Bolivia-Argentina border and in the Brazilian shield respectively (Smith and others, 1981) and world paleogeographic reconstructions based on lithic and biological climatic indicators suggest that western Gondwana was positioned in very high latitude areas (Ziegler and others, 1979).

In Late Devonian (Famennian) time glaciation may have had a significant worldwide effect. Perhaps glaciation was responsible for the rapid Mid-Famennian sea-level regression around the world and for massive biotic extinction (Copper, 1977) similar to the large extinction that occurred in the Late Ordovician. In addition, in the Late Devonian less intensive evaporation due to worldwide refrigeration may have reduced the formation of large quantities of saline water and as a consequence reduced the amount of evaporite deposition on platform interiors. Perhaps the salinity also changed in restricted environments due to low evaporation rates in times of general refrigeration. The warm carbonate belt extending to 40° latitude away from the equator was reduced to about 20° (Figures 4, 95) from Middle Devonian to Late Devonian (Heckel and Witzke, 1979, text-figure 3) (Figure 95). This shrinkage in the carbonate belt (Figure 4,) would be expected when a worldwide refrigeration occurs (Fairbridge, 1964).

At times following a glacial maximum a shale-siltstone maximum may occur due to the power of the winds in front of ice-sheets to sort out clay and siltstone from coarse material. As ice melts in one place and accumulates in another, large quantities of loess are picked up by the wind and delivered elsewhere.

At many places over the world at this time shale succeeded carbonates and evaporites, as for example the Chattanooga shale (Conant and Swanson, 1961) of southeastern United States, referred to as enigmatic by Heckel and Witzke (1979). These associations are here interpreted as due to climatic change and transgression and an

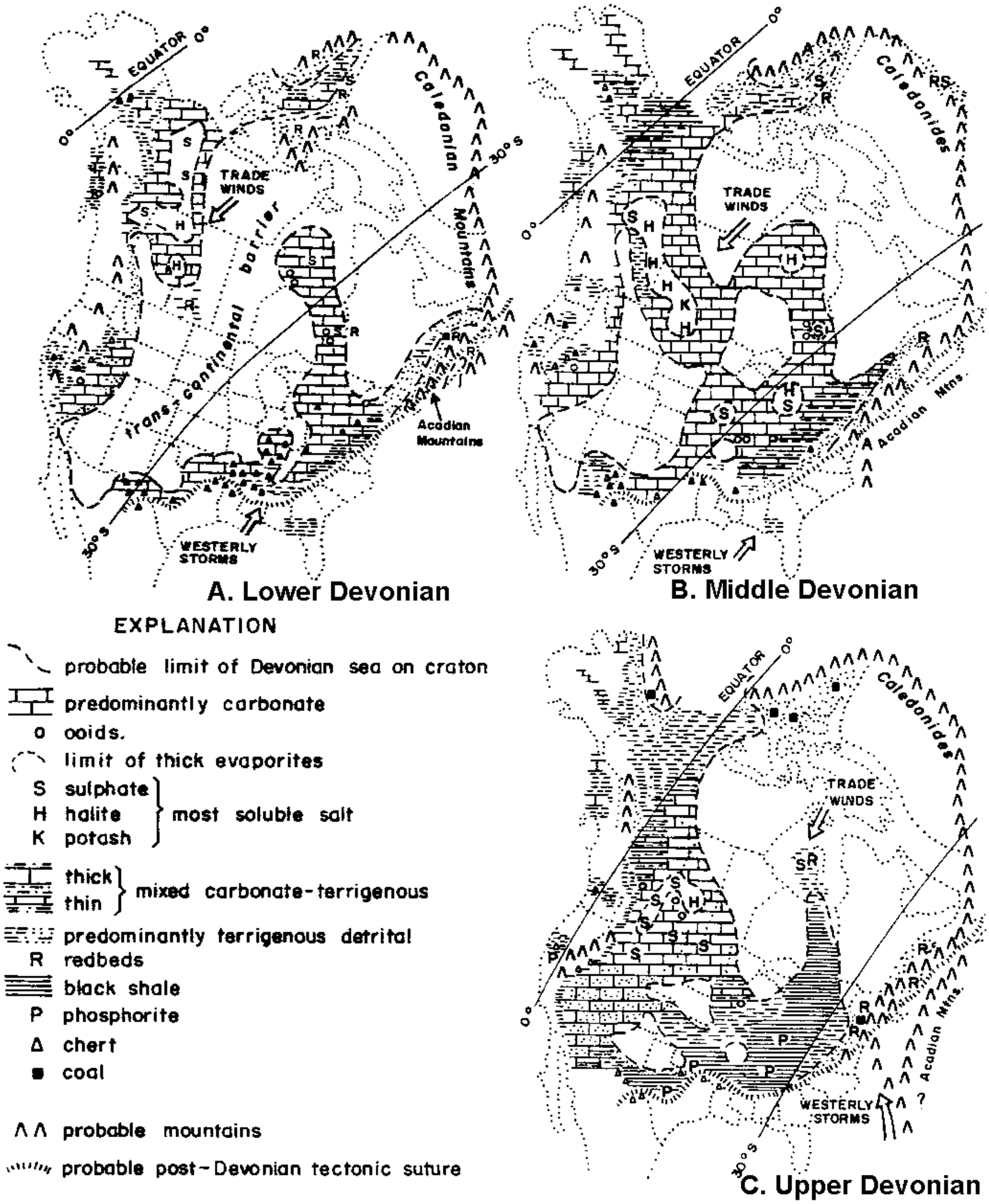


Figure 95. Generalized Devonian lithofacies map of North America after Heckel and Witzke (1979)

increase in the amount of clay and loess in the world ocean.

The North African black limestones (Hollard, 1967, p. 239) and siltstones (Nahon and Trompette, 1982) of Late Devonian age may have been caused by a similar association of events, including the presence of an ice-sheet in northern Brazil and part of west central Africa. The abundant North American shales of Late Ordovician age referred to by Dott and Batten (1976) as an oddity for the craton may also have the same origin.

EARLY CARBONIFEROUS GLACIATIONS

Documentation of Early Carboniferous glaciation is also controversial. Up to the present, Lower Carboniferous glacial sediments have been identified in western South America in Argentina and Bolivia (Frakes and Crowell, 1969a), and in Brazil, in areas adjacent to Bolivia in the Pimenta Bueno Basin (Pinto Filho and others, 1977). Farther east in Brazil, in the Amazonas and Parnaíba basins, Lower Carboniferous sediments show some characteristics of glacial deposits, but the data are poorly known and require further investigation.

In Africa, glacial sediments at the base of the section within the Congo Basin, and within the Zambezi block (Chappel and Humphreys, 1970) may have been deposited in Early Carboniferous time (Frakes and Crowell, 1970a). Glaciation in the Karoo Basin may have begun, according to present data, in either the Late Devonian or Early Carboniferous (Plumstead, 1964).

SOUTH AMERICA

Northern Argentina and Bolivia. A sequence over 2,000 m thick of sandstone, diamictite, shale, and siltstone deposited in both marine and continental environments crops out discontinuously in a belt about 1,300 km in length and 100 to 350 km in width within and along the eastern Andes and extending eastward beneath the Gran Chaco and Beni plain. The strata lie unconformably upon older Paleozoic and Precambrian rocks, and are involved in strong Andean deformation. The section is more complete in the southern Sub-andean Zone than to the north where most of the Upper Carboniferous (Pennsylvanian) section is missing (Reyes, 1972).

Five formations of primarily Early Carboniferous (Mississip-pian) age are recognized in both the southern Subandean Zone and the Altiplano area in the high Andes to the west. To the north the upper units are Late Carboniferous (Early Pennsylvanian) in age. From bottom to top the five formations are the following: Itacua, Tapambi, Itacuami, Tarija, and Chorro. The Itacua Formation, up to 50 m thick, consists of purple and gray diamic-tite associated laterally with variegated quartzitic sandstone, gray and red shale and sandy shale (Reyes, 1972). It was laid down upon an irregular surface between topographic highs where it was not deposited. According to Ayaviri (1972) the basal unit records a transgression with glacial influence, and the local development of diamictites may have been related to montain glaciation, where glaciers emptied into fjords.

The overlying Tupambi Formation, up to 450 m thick, consists mainly of sandstone with conglomerate in the basal part and clay-shale, siltstone and diamictite in its upper part. The sandstone beds are both massive and cross-stratified and the diamictites have a sandy to clayey matrix with a dispersed coarser fraction composed of a variety of metamorphic, sedimentary and igneous clasts, some striated (Reyes, 1972).

The next unit, the Itacuami Formation is up to 300 m thick and consists mainly of dark-gray to dark-red clay-shale and siltstone, which change laterally to sandstone. Diamictite interbeds may dominate locally (Reyes, 1972).

The Tarija Formation up to 600 m thick consists of grey to dark green mudstone containing dispersed clasts of sand- and gravel-size (Mingramm and others, 1979). Thick sandstone interbeds and lenses are characteristic, including a 50 m thick bed near the base (Frakes and Crowell, 1969a). At places diamictites constitute more than 90 per cent of the formation but normally only about 50 per cent. The clasts are composed of volcanic, plutonic, sedimentary and metamorphic rocks of variable sizes, some of them faceted and striated.

The Tarija Formation overlies the Itacuami Formation or the upper member of the Tupambi Formation, suggesting an unconformity beneath it, and an unconformity probably surmounts it, at the base of the overlying formations (Taiguati, Escarpment, or Las Peñas formations). The Chorro Formation, a lateral sandstone facies of the Tarija Formation (Reyes, 1972) is up to 600 m thick, and consists of pink, dark red and light violet, and massive to poorly stratified sandstone. It is interbedded with diamictite containing striated and faceted clasts and is interpreted as glaciofluvial.

Although the age of these formations is somewhat problematic, the Itacua Formation is considered Tournaisian (early Early Carboniferous) based on palynological studies (*Leiozonotriletes* zone) (Reyes, 1972). The age of the Tupambi Formation is also Tournaisian, based on palynological grounds (*Leizonotriletes* and *Baltisphaeridium* zones) and the overlying Itacuami is also considered to be Early Carboniferous (Mississippian). The palynological content of the Tarija Formation suggested a Namurian-Westphalian age to Russo (1977), but Salas (1978), on the basis of palynomorphs, assigned a Late Namurian to Westphalian age to the unconformably overlying Las Peñas Formation and equivalent formations. If the unconformity beween the Tarija and Las Peñas formations represents a considerable time interval, a large part of the Tarija Formation may be as old as Visean (mid-Early Carboniferous).

In the Andean Cordillera and Altiplano region near Lake Titicaca, two possible Early Carboniferous units are present: the Cumaná and Kasa formations. The basal Cumaná Formation, up to 1,850 m thick, consists of greenish gray and light pink, cross-bedded, fine- to coarse-grained sandstone, interbedded with greenish- and blackish-gray micaceous shale, and diamictite. The diamictites are restricted to two levels, one near the base (Calamarca Tillite) and the other near the top. The basal diamictite contains white and red granitic clasts and overlies conformably a predominantly shale section 360 m thick (Reyes, 1972). The Kasa Formation, free of diamictites, overlies the Cumaná Formation conformably and consists of variegated sandstone, siltstone and shale with a 5 m thick carbonaceous horizon at the top containing *Lepidodrendon* sp., *Calamites* sp., and *Neuropteris* sp.

In regard to age, the strata underlying the Calamarca Tillite are assigned to the *Baltisphaeridium* zone (Tournaisian Stage of the Early Carboniferous) and the Calamarca Tillite to the *Micro-reticulatisporites* and *Densosporites* zones (Visean of the mid-Early Carboniferous). Some investigators assign the Kasa

Formation to the late Early Carboniferous, while others consider it as deposited in Westphalian to Early Stephanian time (Early to Late Pennsylvanian), based on *Bothrychlopsis* sp. and *Sphenopteris* sp. (Aceñolaza and others, 1972).

Pimenta Bueno Basin. Brazil. According to Pinto Filho and others (1977), glacial rocks of Early Carboniferous age lie within a graben in westernmost Brazil, close to the Bolivian border, between 10º and 12ºS latitude and 60°30' and 62º45'W longitude. The basal unit, the Pimenta Bueno Formation, is composed of shale, sandstone and diamictite, overlain by variegated shale and brown fine- to medium-grained sandstone.

The upper part of the section is composed of mudstone, diamictite, rhythmically bedded shale and sandstone, and conglomeratic sandstone. The diamictites contain cobbles and boulders up to 1.5 m across, some of which are polished and striated embedded in a massive matrix. Clasts, of great petrographic variety, are commonly encased within a calcite veneer. Some laminated claystone beds contain dropstones that disturbed the underlying laminations. Sandstone and claystone beds show slump structures, and fossil patterned grounds were identified. These features support a glacial and periglacial origin for the unit.

The age of the Pimenta Bueno Formation, previously discussed (p 114), Is considered as Early Carboniferous (Visian) on the basis of palynological data provided by Pinto Filho and others, (1977).

The presence of Mississipian spores as well as the glacial nature of the rocks suggest correlation with presumed glacial rocks of the Faro (Amazonas Basin) and Poti (Parnaíba Basin) formations of Visean age (mid-Early Carboniferous). Similar glacial rocks from the Jauru Valley, 600 to 700 km southeast of the Pimenta Bueno graben also contain palynomorphs of Early Carboniferous age (Olivatti and Ribeiro Filho, 1976; Rocha Campos, 1981a). Shales related to the Jauru tillite in Mato Grosso contain *Reticulatisporites* sp., *Convolutispora* sp., *Acanthotriletes* sp., *Cristatisporites* sp., *Verrucosisporites* sp., *Lycospora* sp., *Lophozonotriletes* sp., and *Vestispora* sp. A glacial influence during deposition of the Jauru strata is shown by typical glacial striae and facets on clasts from diamictites.

Amazonas Basin, Brazil. In the subsurface of the central Amazonas Basin, the Faro Formation contains diamictite close to its base that may be glacigenic. The formation, up to 350 m thick, is composed of two sandstone and two argillaceous units.

The lowermost unit consists of fine- to coarse-grained, white to light gray sandstone that is parallel or cross-stratified and interbedded with thin shale and siltstone. The overlying unit is composed of dark gray to black carbonaceous shale with diamictite horizons that contain sand- to pebble-sized stones immersed in a gray micaceous silty to argillaceous matrix consisting mainly of kaolinite or illite.

The diamictites display a characteristic massive texture with "floating" clasts of great petrographic variety. Some diamictites are found about 60 km away from present basement position, but Paleozoic outcrops before erosion since their deposition were located about 150 km away from the basement. The dip of the diamictite-bearing beds at the time of deposition was about 1/4º, a depositional slope much too gentle for mass flow. Graded beds and other sedimentary structures associated with turbidity currents have not been recognized, nor has evidence of local tectonism. According to electrical log studies, the argillaceous interval with the diamictites is widely distributed, but only a few cores are available, so it is difficult to determine the total extent of the diamictites.

Parnaíba Basin, Brazil. In the Parnaíba Basin, the Poti Formation of Early Carboniferous age contains beds interpreted as laid down under glacial conditions. The formation is up to 170 m thick in surface section, where it consists of two sandstone and two argillaceous units previously described in this study.

In the upper shaley unit, variegated shale and siltstone beds change laterally to purple and pink diamictite in which clasts up to pebble-size are disseminated in a massive, micaceous, silty and clayey matrix, at places, calcareous. Massive sandstones are locally interbedded and at the top the section is capped with some thin limestone lenses. The eastern part of the basin contains coal laminae and finer sandstone beds. The Poti Formation is underlain and overlain by erosional unconformities, but its regional dip is less than 1º and was less at the time of deposition as can be observed in bore holes correlation.

The Poti Formation weathers to a ruinform topography at its base along the eastern side of the basin (Lima and Leite, 1978) and sandstone dikes up to 3 cm in thickness and 20 m in length occurs in its upper part (Ojeda and Bembom, 1966). These dikes are tentatively interpreted as fossil ice wedges filled by sand. Rafted quartzite stones up to 25 cm across were identified by Della Favera and Uliana (1979) associated with what they interpreted as suspension deposits, and Kegel (1953) noted that Poti beds display soft-sediment deformation.

Coarser material is found in the western and southwestern parts of the Parnaíba Basin, suggesting that the Central Brazilian shield was the most important source for the Poti Formation diamictites. This is in contrast to source areas of the Ordovician-Silurian strata, which were located at the eastern and southeastern sides of the basin. The source areas for Late Devonian and Early Carboniferous coarser material were located at the southwestern and western sides of the basin.

The reddish diamictites and massive sandstones with dispersed and unoriented clasts of the Poti Formation contain some carbonate, similar to beds in the Pimenta Bueno Formation and in strata of the Subandean basins. Although no striated pavements or clasts were found, a large part of the Poti Formation may have been deposited under glacial and periglacial conditions as can be inferred from the texture and wide distribution of the diamictites. Aguiar (1971), on the other hand, considered the diamictites as deposited by turbidity currents, although no sedimentary structures characteristic of turbidites were identified by me. The cored black diamictite beds are very similar to those found in the underlying Devonian Cabeças section, but their age, previously discussed, is Carboniferous (Biostratigraphic local Intervals XI, XII; Visean-Mid Early Carboniferous) (Daemon and Contreiras, 1971a,b) based on sporomorphs and fossil plants.

Chaco-Paraná Basin, Argentina. The western continuation of the Paraná Basin of Brazil, the Chaco Paraná Basin, has a surface area of about 700,000 km^2. Inasmuch as Paleozoic strata are covered by Mesozoic rocks, information comes only from boreholes. Within it, the stratigraphic section is subdivided into three units. The lower formation named Sachayoj by Padula and Mingram (1967) is 1,200 m thick and is composed of pyritic black shale with two sandstone intercalations. A diamictite layer underlies each sandstone member characterized by dispersed clasts "floating" in a silty and clay groundmass derived from Precambrian and lower Paleozoic terranes. The age of the unit is considered as late Early Carboniferous (Namurian Stage) based on palynological studies (Amos, 1972). Only the texture of the diamictites and their extent argue for glaciation so far as is now known.

DISCUSSION

Information suggesting centers of Early Carboniferous glaciations are sparse and dating of the strata is still unsatisfactory. Some strength is added to both environmental interpretation and age assignment by extending correlations from basin to basin. For example, the Poti Formation of the Parnaíba Basin and the Faro Formation of the Amazonas Basin appear to correlate with glacial strata of the Pimenta Bueno graben and southward into the Jauru section and far to the west to the beds in the Subandean basins. In the Solimões Basin, at the same age level, a "breccia" horizon with a sandy and silty matrix has been cored in a well.

As more wells are drilled and sampled, Early Carboniferous glaciations may also be better documented for this region. During Visean time a significant amount of sand was delivered to basins around the Brazilian shield, and the Chorro (Subandean basins), upper Sernambi (Solimões Basin), Faro, (Amazonas Basin) and Poti (Parnaíba Basin) formations are interpreted as deposited under glacial and periglacial conditions. This study suggests that in Gondwana cratonic areas the reworking of glacial sediments and regressions furnished coarse clastics to the sedimentary basins, and that coarse material derived from tectonic uplift forms a minor proportion.

LATE CARBONIFEROUS AND PERMIAN GLACIATIONS

An incontrovertible record of continental glaciation during the Late Paleozoic occurs on all Gondwana continents. It was first recognized India (Blanford and others, 1856), in Australia (Selwyn, 1859), in Africa (Sutherland, 1870), in South America (Derby, 1888a,b), in Madagascar (Hirtz, 1950) and in Antarctica (Long, 1962). Over the years since discovery, the record has been extended and re-investigated by many geologists, including Leinz (1937), Rocha-Campos, 1967; Rocha-Campos and dos Santos (1981), Farjallat (1970), Crowell and Frakes (1970a,b; 1971, 1972, 1975) and Frakes and Crowell (1967, 1969a,b; 1970a,b; 1972) and by Frakes and others (1971, 1975). In view of these previous publications, here focus is upon some new discoveries and problems.

The time and place of ending of the vast Late Paleozoic glaciations is reasonably well known and documented (Crowell and Frakes, 1975) and occurred in Australia and perhaps also in then adjoining Antarctica in Mid-Permian time (Early Kazanian Stage). The culmination and region most widely affected by the glaciation was southern Africa, where huge ice sheets coalesced during the transition between Carboniferous and Permian times. The location and timing of the inception of the Late Paleozoic glaciations is less clear, however, probably largely because later advances of ice sheets reworked and even obliterated earlier records (Crowell, 1978, 1981). As described above, Early Carboniferous glaciations are recorded in the Subandean region and in Brazil in the western part of the Gondwana supercontinent. The inception of glacial sedimentation in the Karoo Basin of southern Africa may go back as far as the Visean Stage (middle Lower Carboniferous) (Plumstead, 1964). Whether the climatic refrigeration came on gradually after inception, or whether there was an interval of warmer climate before the onset of the culmination toward the end of the Carboniferous Period remains to be documented.

In recent years, evidence has come in suggesting that Late Paleozoic glaciations were more widespread than previously recognized. In central Arabia and Oman, boulder beds and other facies, dated by palynology as Permo-Carboniferous, document a glacial environment (McClure and Young, 1981; Braakman and others, 1982), and widespread pebbly mudstone beds in Malaya, Thailand, and Burma, which are poorly fixed in time between Early Carboniferous and Late Permian, may have been deposited around the margins of Gondwana, and may have had a glacial contribution (Stauffer and Mantajit, 1981). A glacial interpretation for Upper Carboniferous and Permian beds in parts of Antarctica has been strengthened but their extent and age range has not been significantly expanded as the result of recent investigations (Nelson, 1981; Barrett and McKelvey, 1981; Laird and Bradshaw, 1981; Ojakangas and Matsch, 1981).

CHAPTER 14. DISCUSSION AND CONCLUSIONS

The long debate about existence or non-existence of the Late Devonian glaciation should come to an end. Plate tectonics causes world climatic changes because tectonism can bring land to poles producing worldwide refrigeration (Crowell and Frakes, 1970; Fairbridge, 1973) or it can bring ocean to poles generating world-wide heating. Important consequences of this glaciation are observed where interaction among climate, biota and tectonism play a role in facies changes in the world geological record.

Glaciation and associated events which occurred in Late Devonian (Mid-Famennian) time might have had a significant effect worldwide. For example, it was responsible for the rapid sea-level regression around the world, as well as for the inception of massive biotic extinction (Cooper, 1977). This extinction was very similar to the one that occurred in the Late Ordovician. In Late Devonian time less intensive evaporation, due to a worldwide refrigeratton, might have reduced the formation of large amounts of evaporites on platform interiors. The warm carbonate belt extending to 40-50° of latitude away from the Equator was reduced to about 20° (Figure 95) from Middle Devonian to Late Devonian time (Heckel and Witzke, 1979). An increase in the amount of clay and loess deposits in the world ocean might also have taken place. Glacial rather than non-glacial climates prevailed over Gondwana during the Paleozoic

Era, and the Mid-Famennian regression and unconformity may have been caused by the glaciation recognized in northern Brazil.

Over sixty years ago du Toit (1921, p. 223; 1927) and Wegener (1929) recognized that Permo-Carboniferous ice of united Gondwana first grew in the west and then migrated eastward with time. King (1958), with more data in hand, showed that the glaciation waxed in Argentina before it began in Australia, and that when Australia was cooling, western Gondwana was warming. Crowell and Frakes (1975) and Crowell (1978) added details to reconstructions of Late Paleozoic geography and climatology in Gondwana, and concluded that as the supercontinent moved across the south rotational pole, an intermittent glacial imprint followed its course.

Late Paleozoic glaciation began in western South America during Tournaision or Visean time (Early Carboniferous) and disappeared in eastern Australia and Antarctica during Early Kazanian time (early Late Permian). The ice age lasted for about 90 m.y., from about 330 m.y. ago until about 240 m.y. ago but at no time was there a single huge ice cap. Instead, ice centers waxed and waned across different sites during this long interval. The path of intermittent glacial centers across Gondwana has been traced and briefly documented back into early Paleozoic time, and compared with published data from paleomagnetic investigations (Crowell, 1981; Crowell, 1983).

This study documents the ice-center migration during early and middle Paleozoic times, with special emphasis on the Devonian record. Unfortunately, the direct stratigraphic record is fragmentary, and much glacial evidence has been eroded away and lost forever. At some places, glaciation is established in the vicinity, but the dating may be questionable; at others, the age of the sediments may be reasonably well known, but the glacial input controversial. At several places, however, both age and glacial environment are now known, and these are the data stressed in this synthesis. Using the stratigraphic data alone, this study has constructed a wide swath showing the path of ice centers as they migrated across the united Gondwana from Late Ordovician to Late Permian times.

In review, at the beginning of the Paleozoic Era, continental glaciation affected the northwestern African sector of Gondwana, apparently as the result of waning influences of Late Proterozoic glaciations in that region (Deynoux and others, 1978; Kroner, 1977}. From the Middle Cambrian to the Late Ordovician, there is no record of glaciation. From near the end of the Ordovician Period in Africa and migrating across adjoining South America into the Early Silurian, glaciations were marked, and the record is uncontestable. No glaciation is then recognized until the Late Devonian when the record is clear again in Brazil. Whether small ice caps developed in highlands during this long interval is unknown. From the Late Devonian onward in time glacial centers waxed and waned irregularly until the early Late Permian time when they disappeared. No further large ice sheets have grown upon Gondwana until the onset of the Late Cenozoic Ice Age, when the Antarctic fragment began to develop an ice cap in the Eocene or Oligocene (Harland, 1981; Barrett, 1981).

The drift of the united supercontinent across the Earth´s rotational axis during the Paleozoic Era is independently established by paleomagnetic data (Morel and Irving, 1978; Brock, 1981; Vilas, 1981; Pal and Bhimasankaran, 1972; McElhinny and others, 1974; Smith and others, 1981). Figure 96 shows the apparent polar wander path based on the premise that glacial centers develop on land in high latitudes. Published paleomagnetic apparent wander paths show similar patterns. The similarity suggests that Paleozoic glaciations, like those in the Late Cenozoic, are sited in high latitudes or polar regions, that the magnetic pole was dipolar and approximately coaxial with the rotational axis, and that the angle between the rotational axis and the plane of the ecliptic was not very different from what it is today. In contrast during the Late Proterozoic a body of data from paleomagnetism suggests that continental glaciation took place also in low latitudes (Harland, 1964a,b; Tarling, 1974, 1978; Williams, 1975; McWilliams and McElhinny, 1980; and Hambrey and

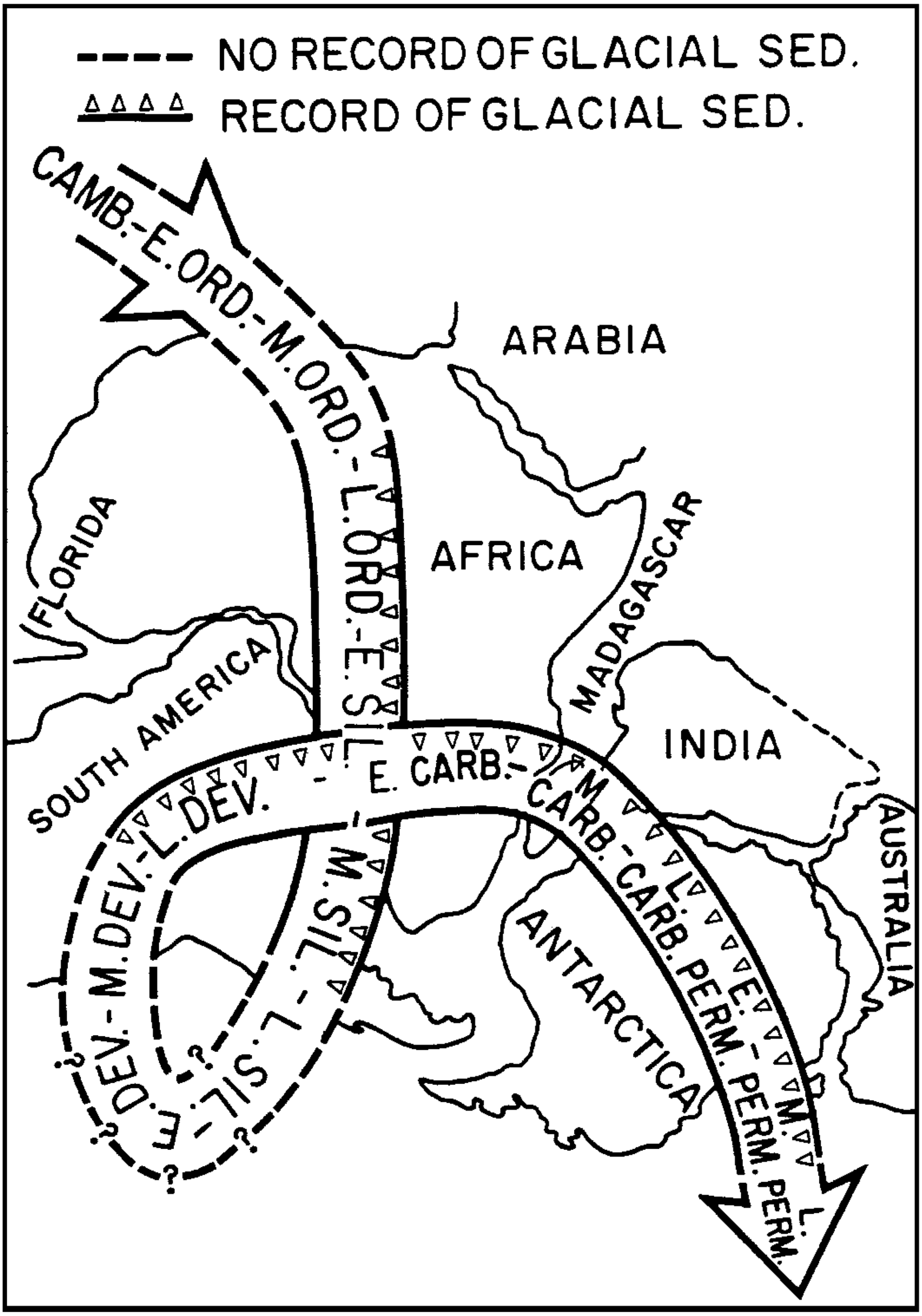


Figure 96. Migration of main ice centers across gondwana throughout paleozoic times based on tillites.

Harland, 1981, p. 944-947; 982-984).

From these two types of apparent wander paths, two observations may help to explain the intermittent history of continental glaciation during the Paleozoic upon Gondwana. First, during most of the Cambrian and Ordovician, when glaciation is not recorded, the supercontinent was not in a near-polar position. The apparent polar wander path determined in this study (Figure 96) shows an entry into north Africa in Late Cambrian and Ordovician time. Paleomagnetic paths also show a departure from Africa in Early Cambrian time, a time when glacial centers waned on that part of Gondwana. During the Late Cambrian, continents were located in middle and low latitudes and the poles in oceanic regions (Ziegler and others, 1979, Fig. 2; Scotese and others, 1979). Second, from Mid-Silurian time until Late Devonian (when there is another long break in the history of Gondwanan glaciation) the apparent wander path had moved from South America into the Paleopacific Ocean coast. This is also a time of gradual increase in the number of coral reefs and associated organisms and an increase in evaporites (Boucot, 1975). A sharp regression in sea level, detected by Johnson (1974) and Isaacson (1978), an abrupt decline in reefs (Boucot, 1975), a reduction in evaporites (Frakes, 1979) and biotic mass extinction (Copper, 1977) predated the Late Devonian glaciation. Between Late Devonian and

Mid-Permian time, the south pole always lay somewhere on the supercontinent. By Early Triassic time Gondwana had glided away from the pole and rotated so that the huge elongate continent was oriented latitudinally instead of longitudinally (Smith and Briden, 1977, Map 39; Crowell, 1978, p. 1363).

To understand the causes of continental glaciation, several additional factors will eventually need appraisal. Eustatic sea-level changes and CO_2 content markedly influence climate, along with other terrestrial changes such as the arrangement of continents which affect the flow of ocean currents and in turn the air. Feedbacks in the climate system are complex (Crowell and Frakes, 1970a; Crowell, 1978). During an ice age worldwide sea-level is lowered in view of the quantities of water contained in glaciers. On the other hand, factors, such as changes in sea-floor spreading rates and tectonics, may also bring about eustatic changes, and lower sea level and so reinforce a glaciation. At the climax of the Late Devonian worldwide regression, glaciation took place and during the Mid Silurian-Early Devonian regression a glacial record is unknown. At other times these same tectonic processes may operate oppositely and cause wide flooding of continental margins, and bring on climate amelioration. Among this mix of processes I speculate that the warm Cambrian and Early and Mid-Ordovician interval, and the end of glaciation in the Permian, may in part be related to the movement of the Gondwana supercontinent out of high latitudes and at the margin of the Pole and into a new orientation so that it no longer provided sites for glacier accumulation.

In Early Silurian time the ice-sheets changed their path and from Late Devonian to Late Permian times completed -an angle of about 90º with the Ordovician-Silurian path. This means a change in the direction of the motion of Gondwana. The ice-sheet path makes a loop, so that northwestern Gondwana was reoriented and moved toward the tropics and eastern Gondwana toward the South Pole. This change may have resulted in collision with Laurasia in Permian times, a collision (Hercynian Orogeny) that finally welded Gondwana to Laurasia making up the gigantic Pangea continent and resulting a worldwide climatic change.

In conclusion, glacial rather than non-glacial climates largely prevailed over wide reaches of Gondwana during the Pale-ozoic Era. Moreover, continental glaciation took place in high latitudes and was apparently controlled by complex terrestrial factors similar to those that operate today and during Late Cenozoic time. As the huge united continent drifted across the South Pole, and from time to time changed its latitudinal orien-tation, and polar regions no longer were oceanic, ice ages ensued. But many other factors guided climate change and challenge the historical geologist to document them and to explain how they operated. Among extraterrestrial factors, geometrical arrangement between the earth and sun (Milankovich effect), bolide events, and oscillation in solar output may also contribute to changes in the intensity of earth's heat budget and consequently in the generation of glacial, interglacial and non glacial times.

APPENDfX A

PALEONTOLOGICAL DATA

Fossils found in several formations of the Amazonas and Parnaíba basins are listed below for biostratigraphic studies.

PITINGA FORMATION

The following fossils were identified in the Pitinga Formation of the Trombetas Group. Paleontological data were obtained from Bouman and others (1960), Mendes and Petri (1971) Daemon and Contreiras (1971a,b) and Da Costa (1974):

Algae: *Tasmanite euzebioi* Sommer, *T. Lamegoi* Sommer, *T. mourai* Sommer, *T. salustianoi* Sommer.

Acritarchs: *Baltisphaeridium* sp., *Dictyotidium* sp., *Leiofusa* sp., *Micrhystridium* sp., *Polyedrixium* sp., *Veryhachium* sp.

Foraminifera: Arenaceous form, genera and species undetermined.

Chitinozoa: *Ancyrochitina erichseni* Sommer and Van Boeckel, *Ancyrochitina swani* Lange,~·aff. *filosa* Eisenack, *A. onniensis major* Costa, *Clathrochitina* Eisenack, *C. striata elongata* Benoit and Taugourdeau, *Conochitina brevis* Taugourdeau and Jekhowsky, *C. brevis conica* Taugourdeau and Jekhowsky, *C. djalmai* Sommerand Van Boeckel, *C. Pellucida* Taugourdeau, *C. caudata* Lange, *C* cf. dactylus

Coll and Schw., *C. filamentosa* Lange, *C. hadrocephala* Lange, *C. inflata* Lange, *C. oblata* Lange, *C.* cf. *proboscifera* Eisenack, *C.* cf. *simplex* Eisenack, *C. conulus* Eisenack, *C. lava herculi* Eisenack, *C. decipiens* Taugourdeau and Jekhowsky, *C. simmilis* Costa, *C. dolosa* Laufeld, *C. dolosa patula* Costa, *C. intermedia* Eisenack, *C. lagenomorpha* Eisenack, *C. lepida* Jenkins, C. *micra-cantha* Eisenack, *C. parvicolla* Taugourdeau, *C. seriespinosa* Jenkins, *C. trombetensis* Costa, *C. vasculiformis* Eisenack, *Cyatochitina conica* Taugourdeau, *C. baculosa* Costa, C. *cam-panulaeformis* Eisenack, *C. cylindrica* Taugourdeau and Jekhowsky, *C. dispar* Benoit and Taugourdeau, *C. dispar verrucata* Taugourdeau, *C. fistulosa* Taugourdeau and Jekhowsky, *C. granulata* Taugourdeau, *C. tenella* Costa, *C. costulata* Lange, C. Rhabdomorpha, Desmochitina *cingulata* Eisenack, *D. minor* Eisenack, *D. andrei* Lange, *D. balanomorpha* Lange, *D. diadema* Lange, *D. digitiformis* Lange, *D.* cf. *errata* Eisenack, *D.* cf. *sommeri* Lange, *D. spongiosa* Lange, *D. lata* Schallrenter, *D. minor amphorea* Eisenack, *D. urna*, *Hoegisphaera complanata* Jansonius, *Illichitina cuspidata* Lange, *Lagenochitina brevicolis* Taugourdeau, *L. cylindrica* Eisenack, *L. exigua* Lange, *L. esthonica* Eisenack, *L. ovoidea* Benoit and Taugourdeau, *L.* sp., *Linochitina cingulata* Eisenack, *L. erratica* Eisenack, *Pogonochitina spinifera inornata* Costa, *Pterochitina perivelata* Eisenack, *P. symetrica* Lange, *Rhabdochitina clavi-*

formes Taugourdeau, *R. magna* Eisenack, *R. truncata* Taugourdeau, *R. turgida* Jenkins, *R. virgata* Taugourdeau, *R. usitata* Jenkins, *Sphaerochitina lepta* Jenkins, *S. annulata* Lange, *S. setosa* Lange, *S. sphaerocephala* Eisenack, *Tanuchitina bergstroemi* Laufeld.

Worms(?): *Arthrophycus harlani* Conrad.

Annelids: Gen. Et sp.undetermined.

Scolecodonts: Gen. Et sp undetermined.

Brachiopods*:* *Lingula* cf. oblata Hall, *Lingula* sp., *Lingulops derbyi* Clarke, *Pholidops trombetana* Clarke, *Orbiculoidea hartti* Clarke, *Orbiculoidea* sp.; *Chonetes* cf. *novascotica* Hall, *Orthis callactis* var. *amazonica* Clarke, *Heterorthella freitana* Clarke, *Dalmanella smithi* Clarke, *Anabaia paraia* Clarke, *Austrocoelia* sp.

Gastropods: *Plectonatus trilobata* Conrad var. *viramundo*, *Murchisonia* sp.

Scyphozoa: *Tentaculites trombetensis* Clarke, *Conularia amazônica* Clarke.

Pelecypods: *Anodontopsis austrina* Clarke, *A. putilla* Clarke, *Clidophorus brasilianus* Clarke, *Tellinomya pulchella* Clarke, *T. subrecta* Clarke.

Cephalopods: *Cyrtoceras?* sp., *Orthoceras* sp.

Ostracodes: *Bollia lata* var. *Brasiliensis* Clarke, *Primitia minuta* Eichwald.

Radiolaria: Gen. and sp. undetermined.

Sponges: Hexactinellids, gen. and sp. undetermined.

Graptolites: *Climacograptus innotatus* Nicholson, var. *brasiliensis* Ruedemann, *Climacograptus* sp. *Monograptus* sp., Graptolites, gen. and sp. undetermined.

MAECURU FORMATION

Caster (1952), Bouman and others (1960), Lange (1967), Mendes and Petri (1971) have contributed to the list of fossils referred below:

Acritarchs: *Baltisphaeridium* sp., *Leiofusidae* sp., *Micrhystridium* sp., *Veryhachium* sp.

Chitinozoa: *Ancyrochitina ancyrea* Eisenack, *Angochitina devonica* Eisenack, *Conochitina edjelensis* Taugourdeau, *Desmochitina erratica* Eisenack, *D. Margaritana* Eisenack, *Lagenochitina* sp., *Cyathochitina* sp.

Sporomorphs: *Spizonotriletes* sp., *Leiotriletes* sp., *Azonotriletes* sp., *Retusotriletes* sp.

Antozoa: *Chaetetes carvalhoanus* Katzer, *Pleurodictyum amazonicum* Katzer.

Crinoids: *Ctenocrinus* sp.

Bryozoa: *Fenestella paralela* Hall, *Fenestella* sp., *Reptaria stolanifera* Rolle, *Rhombopora ambigua* Katzer, *Stictopora* sp.

Annelids: *Hicetes* cf. *Inexatus* Clarke.

Scolecodonts: Gen. And sp. undetermined.

Brachiopods: *Amphigenia elongata* Hall, *Anoplia nucleata* Hall, A. sp., *Anoplotheca flabellites* Conrad, *Australospirifer buarquianus* Rathbun, *A. laurosodrianus* Katzer, *Brachyspirifer pedroanus* Katzer, *B*. sp., *Camarotoechia cf. Sappho* Hall *C*. cf. *dotis* Hall, *Centronella wardiana* Hartt and Rathbun, *C. jamesiana* Hartt and Rathbun, Chonetes comstockei Hartt and Rathbun, *C. curuaensis* Rathbun, *C. freitasi* Rathbun, *C. herbertsmithi* Rathbun, *Cyrtospirifer* sp.; *Cyrtina maecuruensis* Rathbun, *Dalmanella net-toana* Rathbun, Gypidula sp., *Oriskania navicella* Hall and Clark, *Orthothetes agassizi* Hartt and Rathbun, *Orthothetes* sp., *Productella maecuruensis* Rathbun, *Protoleptostrophia perplana* Conrad, *Pustulina pustulosa* Hall, *Rhipidomelea hartti* Rathbun, *R. musculosa* Hall, *Schellwienella agassizi* Hartt and Rathbun, *Spirifer derbyi* Rathbun, *S. duodenarius* Hall, *S. hartti* Rathbun, *S. clarkei* Katzer, *S. laurosodreanus* Katzer, *S. buarquianus* Rathbun, var. *alata* and var. *contracta*, *S. coelhoanus* Katzer, *S. cf. granulosa* Conrad, *S. valenteanus* Hartt, *Stropheodonta* cf. *con-cava* Hall, *S. perplana* Conrad, *S. hoeferi* Katzer, *Terebratula rathbuni* Clarke, *Tropidoleptus carinatus* Conrad, var. *maecuruensis* Katzer, *Tropidoleptus carinatus* Conrad, *Vitulina pustulosa* Hall.

Pelecypods: *Actinopteria eschwegei* Clarke, *A. humboldti* Clarke, *Aviculopecten coelhoanus* Katzer, *Cimitaria karsteni*

Clarke, *Cimitaria* sp., *Cypricardella hartti* Clarke, *C pohli* Clarke, *Grammysia gardneri* Clarke, *G. lundi* Clarke, *G. pissisi* Clarke, G. burmeisteri, Guerangeria (or Nyassa) ortoni Clarke, *Leiopteria browni* Clarke, *L. sankinsi*, *Modiomorpha helmreicheni* Clarke, *M. sellowi* Clarke, *Nucula bellistriata* Conrad, *Nuculites smithi* Clarke, *Palaeoneilo orbignyi* Clarke, *Sphenotus bodenbenderi* Clarke, *Toechomya freitasi* Clarke, *T. rathbuni* Clarke.

Gastropods: *Bellerophon stelzneri* Clarke, *Bellerophon* sp., *Bucaniella reissi* Clarke, *Bucania freitasi* Clarke, *Diaphorostoma* (?) *agassizi* Clarke, *Diaphorostoma darwini* Clarke, *Murchisonia* sp., *Platyceras coutoanus* Katzer, *P. gracilis* Katzer, *P. hartti* Clarke, *P. hussaki* Clarke, *P. meerwarthi* Katzer, *P. steinmanni* Clarke, *P. subconicum* Katzer, *P. symmetricum* Hall var. *maecuruense* Clarke, *P. tschernyschewi* Katzer, *P. whitei* Clarke, P. (?) *agassizi* Clarke, *Plectonotus derbyi* Clarke, *P.* sp., *P*. (?) *salteri* Clarke, *Ptomatis forbesi* Clarke, *Strophostylus varians* Hall.

Pteropods: *Styliolina clavulas* Barrande, *Tentaculites crotalinus* Salter (?), *T. eldregianus* Hartt and Rathbun, *T. stubeli* Clarke, *T. tenellus* Katzer, *T. osseryi* Clarke.

Trilobites: *Anchiopella australis* Clarke, *A. goeldii* Katzer, *Metacryphaeus galea* Clarke, *Dalmanites gemellus* Clarke, *D. infractus* Clarke, *D. maecurua* Clarke, *D. paituna* Hartt and

Rathbun, *D. tumilobus* Clarke, *Homalonotus (Calymene) acanthurus* Clarke, *H. derbyi* Clarke, *Phacops brasiliensis* Clarke, *P. goeldii* Katzer, *P. (Dalmanites) macropyge* Clarke, *P. menurus* Clarke, *P. (?) pullinus* Clarke, *P. scirpeus* Clarke.

Ostracodes: *Beyrichia* sp.

Conodonts: *Ozarkodina* (?) or *Trichognathodus.*

ERERÊ FORMATION

The following fossils were listed in the Ererê Formation (De Oliveira and Leonardos, 1943; Mendes and Petri, 1971; Bouman and others, 1960; Caputo and Andrade, 1968; Daemon and Contreiras, 1971a,b):

Sporomorphs: *Archaeotriletes* sp., *Archaeozonotriletes* sp., *Auroraspora* sp., *Biharisporites* sp., *Calyptosporites* sp., *Dictyotriletes* sp. *Emphanisporites* sp., *Leiotriletes* sp. *Retusotriletes* sp., *Samarisporites* sp., *Spinozonotriletes* cf. *echinatus* Benoit, *Zonotriletes* sp., *Hymenozonotriletes*, *Calamospora* sp *Calyptosporites*.

Acritarchs: *Baltisphaeridium* sp. *Cymatiosphaera* sp. *Dictyotidium* sp., *Leiofusa* sp., *Micrhystridium* sp., *Navifusa bacillum*, *Polyedryxium* sp., *Veryhachium* sp.

Alga: *Tasmanites ericheseni* Sommer and Van Boekel

Chitinozoa: *Alpenachitina* sp., *Ancyrochitina* sp., *A.* cf.

multiramosa, *A. tumida* Taugourdeau and Jek, *Angochitina bifurcata*, *A. capillata* Eisenack, *A. devonica* Eisenack, *Lagenochitina sphaerica* Coll., *Sphaerochitina collinsoni* Dunn, *S. cuvillieri* Taugourdeau, *S. vitrea*.

Brachiopods: *Anoplotheca* (?) sp. *Brachyspirifer pedroanus* Hartt, *Camarotoechia* cf. *dotis* Hall, *Centronella jamesiana* Hartt and Rathbun, *C. wardiana* Hartt and Rathbun, *Chonetes comstockei* Hartt, *C. Freitasi* Rathbun, *C. herbertsmithi* Hartt and Rathbun, *C. onettianus* Rathbun, *Cyrtina* (?) *curupira* Rathbun, *Dalmanella nettoana* Rathbun, *Derbyina jamesiana* Hartt and Rathbun, *D.* sp. *Lingula ererensis* Rathbun, *L. gracana* Rathbun, *L. rodriguesi* Rathbun, *L.* cf. *spatulata* Hall, *L. stautoniana* Rathbun, *Mucrospirifer* sp., *Orbiculoidea lodiensis* Hall, *Orthis* sp., *Pustulatia pustulosa* Hall, *Rhynchonella ererensis* Rathbun, *Schellwienella agassizi* Hartt and Rathbun, *Schuchertella agassizi* Hartt, *Spirifer (Acrospirifer) pedroanus* Hartt and Rathbun, *S. cf. pedroanus* Hartt and Rathbun, *S.* cf. *granulosus* Conrad, *S. valenteanus* Hartt, *Terebratula derbyana* Hartt and Rathbun, *Tropidoleptus carinatus* Conrad.

Pelecypods: *Edmondia sylvana* Hartt and Rathbun, *Goniophora woodwardi* Clarke, *Grammysia ulrichi* Clarke, *Nuculana diversa* Hall, *Modiomorpha pimentana* Hartt and Rathbun, *Nucula kayseri* Clarke, *Nuculites branneri* Clarke, *N. ererensis* Hartt and Rathbun, *N. nyssa* Hall *var. majora* Clarke, *Palaeoneilo* (?) *pondiana* Hartt and Rathbun, *P. sulcata* Hartt and Rathbun, *P.* (?) *simplex* Hartt and Rathbun, *Pholadela parallela* Hall, *Sphenotus gorceisei* Clarke.

Gastropods: *Bellerophon morganianus* Hartt and Rathbun, *Bucaniella coutinhoana* Hartt and Rathbun, *Platyceras furmanianum* Hartt and Rathbun, *Diaphorostoma darwini* Clarke, *Pleurotomaria rochana* Hartt and Rathbun.

Pteropoda: *Tentaculites eldregianus* Hartt and Rathbun.

Trilobites: *Asteropyge paituna* Hartt and Rathbun, *Dalmanites Ulrichi* Katzer, *Homalonotus* sp., *Homalonotus oiara* Hartt and Rathbun.

Ostracodes: *Beyrichia* sp.

BARREIRINHA FORMATION

The fossils found in the formation are listed below, According to De Oliveira and Leonardos (1943) Mendes and Petri (1971); Bouman and others (1960) Caputo and Andrade (1968) and Macambira and others (1977).

Plants: *Lycopodites amazonica* Dolianiti, *Lepidodrendales* sp.

Tasmanaceae: *Tasmanites avelinoi* Sommer, *T. derbyi* Sommer, *T. euzebioi* Sommer, *T. finkii* Sommer, *T. hartti* Sommer, *T. lamegoi* Sommer,.*T. mourai* Sommer, *T. roxoi* Sommer, *T. salustianoi* Sommer, *T. sommeri* Van Boeckel, *T. tapajoensis*, *T. mosesi*, *T.* sp.

Sporomorphs: *Samarisporites* sp., *Calyptosporites* sp., *Biharisporites* sp., *Hymenozonotriletes lepidophytus*, *H.* sp. *Spinozonotriletes cf. echinatus*, *Ancyrospora* sp., *Hystricho-Sporites cf. corystus*, *Reticulatisporites* sp., *Acanthotriletes* sp., *Densosporites* sp., *Nikitinsporites* sp., *Verrucosisporites* sp., *Azonotriletes* sp., *Apicularella* sp., *Brochotriletes* sp., *Liotyotriletes* sp., *Emphanisporites* sp., *Apiculatisporites* sp., *Grandispora* sp., *Convolutispora* sp., *Geminospora* sp., *Punctatisporites* sp., *Archaeotriletes* sp., *Lophotriletes* sp., *Retusotriletes* sp.; *Endosporites* sp., *Calamospora* sp., *Pustulatisporites* sp., *Retraletes* sp.,

Acritarchs: *Baltisphaeridium* sp., *Cymatiosphaera* sp., *Hystrichosphaeridium* sp., *Navifusa bacillum*, *Leiofusa* sp., *Polyedrixium* sp., *Pterospermopsis* sp., *Pulvinosphaeridium* sp., *Veryhachium octoaster*, *Duvernaysphaera radiata*, *Maranhite brasi-liensis,* Gorgonisphaeridium sp., *Synsphaeridium* sp.

Chitinozoans: *Alpenachitina eisenacki*, *Angochitina gracana A capillata*, *A. devonica*, *A. crumena*, *A. mourai*, *Ancyrochitina spinosa, A. tumida*, *A. ancyrea*, *A. langei* Sommer and Van Boeckel, *A. torentosa*, *A.* cf. *multiramosa*, *A. striata*, *Cladochitina* sp.
Conochitina sp., *Cyatochitina* sp., *Desmochitina someri* Lange, *Hablochitina glabra*, *H. cumdonlensis*, *Illichitina* sp.; *Lagenochitina avelinoi* Lange, *L.* cf. *avelinoi, L. sommeri*,

Linochitina sp., *Plectochitina tapajonica*, *Ramochitina ramosi*, *Sphaerochitina collinsoni*, *S. cuvillieri*, *S.* cf. *setosa*, *S. vitrae* Taugourdeau, *Urochitina bastosi*.

Brachiopods: *Lingula gracana* Rathbun, *L. stautoniana* Rathbun, *Orbiculoidea lodensis* Hall, *Orthotes agassizi* *Schizobolus truncatus* Hall, Spirifer sp.

Pelecypods: *Nuculites parai* Clarke, *Palaeoneilo scuptilis* Clarke

Gastropods: *Loxonema* (?) sp.

Trilobites: *Phacops menurus* Clarke.

Conodonts: *Hindeodella* sp., *Lonchodus* sp., *Trichognathus* sp.

ORIXIMINÁ FORMATION

The palynomorphs found (Daemon and Contreiras, 1971a,b) in the unit are the following:

Sporomorphs: *Reticulatisporites* sp., *Vallatisporites* sp., *Convolutispora* cf. *vermiformis* Hugues and Playford, *Acanthotriletes* sp., *Ancyrospora* sp., *Pterospermopsis* sp., *Hymenozonotriletes* sp., *Densosporites* sp., *Maranhites brasiliensis* Brito, *Hystrichosporites* cf. *corystus* Richard, *Biharisporites* sp., *Samarisporites* sp., *Calyptosporites*, *Spizonotriletes* cf. *echinatus* Bernort, *Duvernaysphaera radiata* Brito.

Acritarchs: *Veryhachum* sp.

FARO FORMATION

The main palynomorphs identified by Daemon and Contreiras (1971a,b) in the interval are the following:

Hymenozonotriletes sp., *Reticulatisporites magnidictyus* Play and Hel, *Cristatisporites* sp., *Acanthotriletes* sp., *Vallatisporites* sp., *Raistrickia* cf. *baculosa* Hacquebard, *Densosporites* sp., *Waltzispora*, *Reticulatisporites*, Convolutispora cf. *vermiformis* Hughes and Playford.

MONTE ALEGRE FORMATION

The following fossils are known in the Monte Alegre Formation, according to Bouman and others (1960):

Foraminifera: *Ammovertella* or *Calcitornella* sp., *Globivalvulina* sp., *Millerella* sp., *Paramillerella* sp., Plectogyra sp., *Tetrataxis* sp. *Linoproductus* sp., *Neospirifer* sp. *Orbiculoidela* sp.

Briozoans: *Fenestella* sp.

Brachiopods: *Brasilioproductus chandlessi* Derby, *Buxtonicides* cf. *amazonicus* Katzer, *Composita* sp., *Cleiothyridina* sp., *Dictyoclostus* sp. *Kiangsiella* sp. *Lingula* sp., *Linoproductus* sp., *Neospirifer* sp. *Orbiculoideia* sp. *Punctospirifer* sp., *Spirifer oliverai* Mendes, *Spirifer* sp.

Conodonts: Cavusgnatus sp., Hindeodella sp., Streptognathodus sp.

ITAITUBA FORMATION

The following list of fossils is based on Petri (1952, 1956, 1958), Caster and Dresser (1955), Mendes (1956a, 1956b, 1956c, 1957a, 1957b, 1959, 1961), Bouman and others (1960), Fulfaro (1965) and Caputo and Andrade (1968):

Foraminifera: *Ammovertella* or *Calcitornella* sp., *Cribrospira* (?) sp., *Cribrostomum* sp., *Boschubertella* sp., *Endothyra* sp., *Fusulinella* sp., *Globivalvulina* sp., *Glomospira* sp., *Hyporamminoies* sp., *Millerella* cf. *marblensis* Thompson, *Millerella* sp., *Paleotextulana* sp., *Paramillerella* cf. *cooperi* Zeller, *Paramilerella* derbyi Petri, *Paramillerella* sp., *Plectogyra* cf. *pandoral* Zeller, *Plectogyra* sp., *Profusulinella* *silvai* Petri, *Pseudostaffela* sp., *Spirillina* sp., *Staffella* sp., *Tetratis* sp., *Wedakindellina* sp.

Corals: *Aulopora* (?) sp., Campophyllum (?) sp., *Fistulipora* nodulifera Meek (?), *Lithostrotion* (?) sp., *Lophophyllum* (?) sp., *Michenilea* (?) sp., *Monticulipora* (?) sp., *Policoela* (?) sp., *Stenopora* (?) sp., *Zaphrentis* sp., *Clisiophyllum* sp., *Londsdaleia* *rudis* White, *Rugosa* sp., single corals, gen. and sp. undetermined.

Echinoderms: *Archaeocidaris* sp., *Cyathocrinus* sp., *Eocidaris* *hallianus* Geinitz (?), *Erisocrinus loczyi* Katzer, *Erisocrinus* (?) sp., Holothurian plates, genus and species undetermined.

Brachiopods: *Athyris subtilita* Hall, *Athyris* cf. *pectimifera* Sowerby, *Avonia derbyi* Reed, *Avonea rhomeana* Derby, *Brasilia-productus chandlessi* Derby, *B. chronici* Mendes, *B.* cf. *chandlessi* Derby, *Buxtonioides amazonicus* Katzer, Buxtonioides aff. *amazo-nicus* Katzer, *B.* sp., Camarophoria sp., *Chonetes* sp., *Cleiothyridina casteri* Dresser, C.·derbyi Dresser, *C.* aff. *hirsuta* Hall, *C.* sp., *Composita subtilita* Hall, *Composita* cf. *subtilita* Hall, *C. reedi* Mendes, *C. trilobata* Dumbar and Condra, *Composita* sp., *Craneana* (?) sp., *Crania* sp., *Crurithyris granularis* Dresser, *C.* sp., *Derbyia* cf. *crassa* Meek and Hayden, *D. buchi* D'Orbignyi, *Derbyoides tapajotensis* Derby, *D.* sp., *Derbyina* sp., *Dictyoclostus* cf. *burtingtonensis* Hall, *D. inflatus* McChesney, *D* sp., *Dielasma itaitubensis* Derby, *Discina* cf. *missouriensis* Shumard, *D.* sp., *Duartea batesiana* Derby, *Echinoconchus katzeri* Mendes, *Eumetria punctulifera* Shumard, *Heteralosia cornelliana* Derby, *Hustedia mor-moni* Marcou, *H. amarali* Mendes, *H.* sp., *Harttina* (?) *coutinhoana* Derby, *Juresania amazonensis* Mendes, *J.* cf. *nebrascaensis* Owen, *J. paraensis* Reed, *J.* sp., *Kiangsiella halliana* Derby, *K.* sp., *Neospirifer* sp., *Orbiculoidea* sp., *Orthotichia morganiana* Derby, *Phricodothyris mourai* Mendes, *P.* sp., *Productus clarkeanus* Derby, *P. nebrascensis* Owen, *P. punctatus* Martin (?), *P.* sp. *Puncto-*

spirifer leinzi Mendes, *Koslowskia petrii* Mendes, *Krotovia walla-ciana* Derby, K. Derbyi Reed, *Lingula* sp. *Linoproductus derbyi* Mendes, *L.* aff. *Derbyi* Mendes, *L.* sp., *Lissochonetes amazonicus* Derby, *L.* sp.; *Marginifera oddoni* Mendes, *Neospirifer dresseri* Mendes, *Rhipidomella penniana* Derby, *R.* sp., *Rhynchonella pipira* Derby, *Seminula argentea* Shepard, *Spirifer* cf. *increbescens* Hall, *Spirifer oliverai* Mendes, *S.* cf. *oliveirai* Mendes, *S. piracanensis* Katzer, *S.* sp., *Spiriferella* sp., *Spiriferina spinoza* Norwood and Pratten, *Spirigera* sp., *Streptorhynchus correanus* Derby, *S.* aff. *correanus* Derby, *S.* sp., *Waldheimia coutinhoana* Derby.

Bryozoans: *Fenestella intermedia* Prout, *F.* Shumardi Prout (?), *Fenestella* sp., *Glauconome trilineata* Meek(?), *Monticulipora brasiliensis* Barbosa, *Polypora submarginata* Meek (?), *P. derbyi* Barbosa, *Ptilodicta carbonaria* Meek (?), *Rhombopora lepido-dendroides* Meek (?), *Septopora katzeri* Barbosa, *Synocladia biserialis* Swallow(?), *Trepostomata* sp.

Pelecypods: *Acanthopecten* sp., *Allorisma subcuneata* Meek and Worthen, *A.* sp., *Astartella* (?) sp., *Avicula* cf. *longa* Geinitz, *Bakewellia bicarinata* King, *Bakewellia parva* Meek and Worthen, *Aviculopecten carboniferous* Stevens, *A.* cf. *carboniferus* Stevens, *A. coxanus* Meek and Worthen, *A. hertzer* Meek, *A. lyelli* Dawson, *A. neglectus* Geinitz, *A. occidentalis* Shumard, *A. subquadratus* Bell, *A.* sp., *Aviculopinna americana* Meek, *Bithophagus* sp., *Chaenomya* (?) sp., *Conocardium* sp., *Edmondia rudis* McCoy, *Edmondia* (?) sp., *Entolium aviculatum* Swallow, *Euniorotis hawni* Meek and Worthen (?), *Leptodesma occidentalis* Girty (?), *Lima retifera* Shumard, *L.* cf. *retifera* Shumard, *Lithophagus poolii* Dawson, *Macrodon tenuilineatus* Meek and Worthen, *Macrodon* sp., *Modiola* sp. *Myalina kansasensis* Shumard, *Myalina keckuki* Worthen, *M. swalloni* MacChesney (?), *M. wyomingensis* Lee (?), *M.* sp., *Nuculana* sp., *Nuculites* (?) sp., *Palaeoneilo* (?) sp. *Pinna peracuta* Shumard, *Pleurophorus tropidophorus* Meek (?), P. sp., *P. subcostalus* Meek and Worthen, *P. similis* Girty, *Posidonomya* (?) sp., *Pseudomussium* (?) sp., *Pseudosonotis* aff. *aequistriata* Beede, *Pseudomonctis* sp., *Pteria* sp., *Pteronites* (?) sp., *Schizodus rossicus* De Verneuil (?), *S. wheeleri* sp Swallow, *S.* sp., *Sedgwickia* sp., *Solenomya* sp., *Solenopsis* sp., *Volsella* (?) sp., *Yoldia* (?) sp.

Gastropods and Scaphopods: *Aclis* sp., *Bellerophon carbonarius* Cox, *B. crassus* Meek and Worthen, *B.* (?) sp., *Capulus* sp., *Dentalium* sp., *Euomphalus luxus* White (?), *E. sulcifer* Girty, *E.* sp., *Loxonema* sp., *Murchisonia* sp., *Naticopsis nana* Meek and Worthen, *N.* sp., Pharkidonotus sp., *Platyceras nebrascensis* Meek, *Pleurotomaria conoides* Meek and Worthen, *P.* cf. *depressa* Cox, *P. marconana* Geinitz, *P. speciosa* Meek and Worthen, *P.* cf. *subdecussata* Geinitz, *Polyophemopsis* sp.

Cephalepods: *Orthoceras* cf. *cribrosum* Geinitz.

Trilobites: *Griffithides tapajotensis* Katzer, G. (?) Phillipsia (Amerura) *duartei* Kegel, *P.* (*Ameura*) *plummeri* Kegel.

Pisces: Ctenacantideae.

Ostracodes: *Gavollina* sp., *Gytherellina* sp.

Conchostraca: *Asmussia* sp. Cycloosthericides sp. *Lioestheria* sp., Estheriae genus and species undetermined.

Conodonts: *Cavusgnathus* cf. *lanta* Gunnel, *Cavusgnathus* sp. *Euprioniodina* sp., *Gnathodus* sp. *Hindiodella* sp., *Ideognathodus* sp., *Ligonodina* sp., *Lonchodus* sp., *Ozarkodina* cf. *delicatula* Tauffer and Plummer, *Polygnathodella* sp., sp., Pteronites (?) sp., Schizodus rossicus De Verneuil (?), S. *Streptognathodus* sp. *Symprioniodina* (?) sp.

Scolecodonts: genus and species undetermined.

NOVA OLINDA FORMATION

The following fossils list is based on Bouman and others (1960), Caputo and Andrade (1968) and Lima (1982, written communication) from the Petrobras paleolab. The fauna of the Nova Olinda Formation shows some similarities to that of the Capacabana Group of Peru and Bolivia and both formations seem to have a similar derivation (Mendes, 1959):

Sporomorphs: *Plicatipollenites indicus* Lele, *Protohap-Loxypinus* sp., *Striomonosaccites* sp., *Potonieisporites norvicus*

Bhardwaj, *Hamiapollenites* sp., *Guthaerlisporites* sp., *Taeniae-sporites* sp., *Laevigatosporites* sp., *Striatosporites major* Bhardwaj.

Foraminifera *Ammovertella* or *Calcitornella* sp., *Globivalvulina* sp., *Millerella* sp., *Paramillerella* sp., *Fusulinella* sp., *Plectogyra* sp., *Tetrataxis zelleri* Petri, *Wedekindellina* sp.

Corals: *Syringopora* sp.

Brachiopods: Derbyoides (?) sp., Duartea cf. batesiana Derby, *Lingula* sp., *Linoproductos derbyi* Mendes, *Orbiculoidea* sp., *Productus* sp., *Rhipidomella penniana* Derby, *Rhipidomella* sp., *Spirifer* sp., *Streptorhynchus correanus* Derby, *S.* sp.

Pelecypods: *Allorisma* sp., *Aviculopecten* sp., *Myalina* (?) sp., *Pleurophorus* (?) sp., *Pteria* sp., *Pteronites* sp., *Volsella* sp.

Crustaceans: *Asmussia* sp., *Estheriina* sp., Estheriae genus and species undetermined.

Conodonts: *Cavusgnatus* sp., *Gnathodus* sp., *Hindeodella* sp., *Idiognathodus* sp., *Prioniodina* sp., *Strepto-gnathodus.*

Sphenophyta: *Calamites* sp.

ALTER DO CHÃO FORMATION

The palynomorphs mentioned by Daemon (1975) are the following:

Gnetaceaepollenites diversus Stover, Incertae sedis sp. (?), *Elaterosporites protensus* Stover, *Classopoleis classoides* Pflug, *Chomotriletes* sp., *Ephedrites irregularis* Herngreen, Elaterocolpites castelaini Jardine and Magloire, *Elateroplicites africaensis* Herngreen, *Steevesipollenites binodosus* Stover.

ITAIM FORMATION

The following fossils were reported by Kegel (1953), Mesner and Wooldridge (1964) and Moore (1963).

Plants: Psilofitale, Psilopsid.

Conidaria: *Conularia undulate* Conrad, *Conularia sp*.

Cephalopods: *Bellerophon* sp.

Pelecypods: *Pterinopecten* sp., *Nuculites* sp., *Tentaculites* cf. *eldredgianus* Hartt and Rathbun, *T. stubeli* Clarke, *T*. sp.

Brachiopods: Amphigenia sp., *Chonetes* sp.; *Derbyina smithi* Derby, *Eodevonaria* sp., *Spirife*r sp., *Orbiculoidea* sp.

Ostracodes: Bairdia sp., *Bythocypris* sp., *Primitia* sp.

Crinoids: Genus and species undetermined.

PIMENTEIRA FORMATION

The following list of fossils found in the Pimenteira Formation is based on Caster (1952), Kegel (1953), Kräuse and Dolianiti (1957), Mesner and Wooldridge (1964), Moore (1963), Mendes and Petri (1971), Beurlen (1970) and Campanha and Mabesoone, (1976).

Plants: Psilopsids, *Archaeosigillaria picosensis* Krausel and Dolianiti, *Haplostigma* sp., *Palaeostigma* Seward Krausel and Dolianiti, *Protolepidodendron* Kegeli Krausel and Dolianiti, *Spongiophyton* sp.

Cnidaria: *Mesoconularia* cf. *africana* Sharpe.

Brachiopods: *Acrospirifer* sp., Amphigenia aff. *curta*, *A*. sp., *Athyris* aff. *spiriferoides*, *Atrypa* sp., *Brachyspirifer* sp., Camaratoechia sp., *Chonetes arcey* Ulrich, *C.* cf. *syrtalis* Conrad, *C. falklandicus* Morris and Sharpe, *C.* sp., *Clarkeia antisiensis* d'Orbigny, *C*. sp., *Criptonella* sp., *Cyrtina* sp., *Cyrtopirifer* sp., *Derbyina smithi* Derby, *D. whitiorum* Clarke, *D.* aff. *whitiorum* Clarke, *D.* sp., *Frimbrispirifer* sp., *Leptocoelia* aff. *flabellites, Leptostrophia* sp., *Lingula keideli* Clarke, *Orbiculoidea* sp., *Paranaia* sp., *Schizobolus* sp., *Schuchertella agassizi* Hartt and Rathbun, *S. sancticrucis* Clarke, *S*., sp., *Russelaria* (?) sp., *Rhipidomella* (?) sp., *Rhyncospirina* formosa (?), *Spirifer* aff. *inheringi*, *S*. aff. *clarkei*, *S*. aff. *subcuspidatus* Schnur, S. sp., *Tropidoleptus* cf. *carinatus* Conrad.

Gastropods: *Bucanella* cf. *derbyi* Clarke, *B. direisi* Knod, *B*. sp., *Diaphorostoma allardycei* Clarke, *Hyolithus* sp., *Tentaculites* sp.

Pelecypods: *Grammysia oviformis* Knod, *G. rara* Kozlowski, *G*. sp., Modiophorma sp., Nuculites cf. branneri Clarke, Nuculites sp., *Nucula* sp., *Palaeoneilo magnifica* Clarke, *P. sancticrucis* Clarke, *P.* sp., Panenka sp., *Solenomya* (?) sp.

Conularids: *Mesocunularia* aff. *africana*, *Conularia* aff. *huntiana.*

Trilobites: Euripterideae, *Burmeisteria notica* Clarke, *B.* sp., *Mecryphaeus australis* Clarke, *Asteropyge* cf. *paituna* Hartt and Rathbun.

Ostracodes: *Bairdia* sp., *Kloedenia* sp.

Chitinozoans: *Alpenachitina*, *Angochitina*, *Ancyrochitina ancyrea* Eisenack, *A. langei* Sommer and Van Boekel, *Cladochitina* sp., *Cladochitina* sp., *Ramochitina ramosi* Sommer and Van Boekel, *Sphaerochitina*, *Lagenochitina.*

Acritarchs: *Duvernaysphaera tesela* Deunf, *Evittia sommeri* Brito, *Dictyotidium* aff. *polygonium* Staplin, D. sp., *Maranhites brasiliensis* Brito, *M. Mosesi* Sommer, *M. pulcher* Brito, *Navifusa brasiliensis* Brito and Santos, *N. cylindrica* Brito and Santos, *N. eisenacki* Brito and Santos, *N. multistriata* Brito, *Polydrixium* sp., *A*, *P*. sp., *B*, *P.*·sp. *C*, *P.* sp., *D*, *Pseudolunulidia imperatrizensis* Brito and Santos.

Sporomorphs: *Tasmanites mourae* Sommer and Van Boekel, *T.* cf. *roxoi* Sommer and Van Boekel, *T.* sp., *Tapajonites* cf. *roxoi* Sommer and Van Boekel, *Grandispora* sp., *Samarisporites* sp., *Bihari-sporites* sp., *Auroraspora* sp., *Nikintioporites* sp.

LONGÁ FORMATION

According to Sommer and Van Boeckel (1964), Mendes and Petri (1971), Lima and Leite (1978) the fossils found in the Longá Formation are:

Chitinozoans: *Ancyrochitina langei* Sommer and Boeckel, *A. ancyrea* Eisenack, *Lagenochitina* sp., *Angochitina* sp., *Conochitina* sp., *Sphaerochitina* sp., *Ramochitina ramosi* Sommer and Boeckel.

Acritarchs: *Maranhites brasiliensis*, *M. mosesi*, *M. pulcher*, *Duvernaysphaera* sp., *Veryhachium* sp., *Umbellasphaeridium* sp.

Sporomorphs: *Hymenozonotriletes* sp., *Vallatisporites* sp., *Convolutispora* sp., *Acanthotriletes* sp., *Knoxisporites* sp., *Reticulatisporites* sp., *Ancyrospora* sp., *Samarisporites* sp., *Waltzispora* sp.

Plants: Undetermined fragments.

Equinoids: *Protaster* sp.

Brachiopods: *Chonetes* sp., *Lingula* sp., *Orbiculoidea* sp., *Schellwienella* sp., *Clarkeia antisiences.*

Pelecypods: *Janeia* sp., *Tentaculites* sp., *Palaeonelus* sp.

Trilobites: *Metacryphaeus* sp.

Ostracodes: *Kloedenia* , sp., *Primitia* sp.

Fish: Undetermined fragments.

POTI FORMATION

The following fossil list is based on Kegel (1954), Dolianiti (1954), Mesner and Wooldridge (1964), and Lima and Leite (1978):

Palynomorphs: *Reticulatisporites absimilis*, *Reticulatisporites* cf. *magnidictyus* Play and H., *Hymenozonotriletes dolianitii*, *Knoxisporites* sp., *Convolutispora* sp., *Phyllothecotriletes* sp.

Pelecypods: *Edmondia index*, *Edmondia celebris*, *Edmondia acclina*, *Edmondia obliquata*, Edmondia dequechi, *Edmondia corpulenta*, *Edmondia* sp., *Aviculopecten* sp.

Brachiopods: *Derbyoides* sp.

Plants: *Adiantites gothanica*, A. oliveiranus, *A. santosi*, *A. alvaro-albertoi*, *Cardioperidium* sp., *Sphenopteridium* *Kegelidium* sp., *Paulophyton* sp., *Rhodia.*

PIAUÍ FORMATION

According to Mabesoone (1977) and Oliveira and Leonardos (1943) The fossils found in the unit are:

Trilobites: *Phillipsia plummeri* Kegel, *P. duartei* Kegel.

Cephalopods: Genera and species undetermined.

Brachiopods: *Linoproductus* sp., Spirifer sp., Orbiculoidea sp., *Lingulidiscina* cf. *missouriensis*, *Lingula* sp., *Lingula* cf.

carbonaria, *Spirifer* cf. *opimus* Hall, *Productus* sp., *Productus* cf. *americanus*, *Orbiculoidea* sp., *Derbyoides* sp., *Derbya* sp.

Pelecypods: *Astartella* sp., *Edmondia* sp., *Nucula* sp., *Aviculopecten* sp., *Leiopteria* sp., *Allorisma* sp., *Schizodus* sp., *Pleurophorus* sp.

Gastropods: *Bellerophon* sp.

Plant: Genera and species undetermined.

PEDRA DE FOGO FORMATION

In the unit the following fossils were found (Mabesoone, 1977), Mesner and Wooldridge (1964), and Lima and Leite (1978).

Sporomorphs: Laevigatosporites sp., Punctatisporites sp., Virkkipollenites sp., Vestigisporites sp., Verrucosiporites sp.

Plants: *Psaronius brasiliensis*, *P. arrojadoi*, (Pecopteris and Gymnosperms).

Fish: *Ctenacanthus* sp., *Pleuracanthus* sp.

Anphibian (labirinthodont): Prionosuchus sp.

Conchostracans: *Estheria* sp.

Pelecypods, **ostracodes** and **algae** were also recovered.

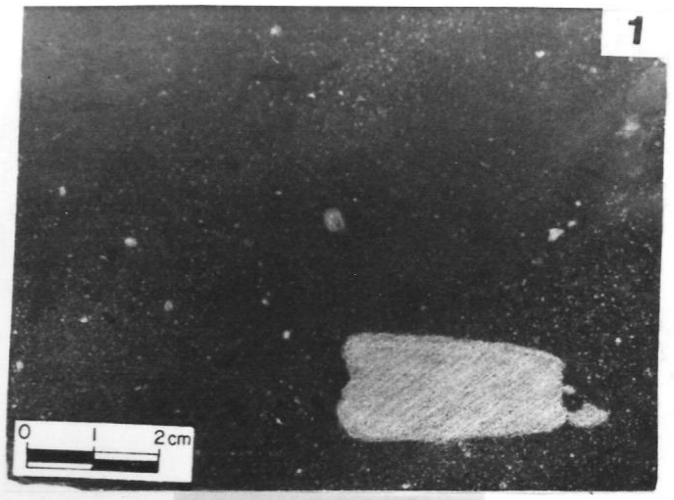
APPENDIX B

PHOTO CAPTIONS

Photo1. Dark gray diamictite containing angular sand-, granule- and pebble-sized clasts of great petrographic variety. Core from the well 1-RBB-1-AM (Jaraqui Formation) of Late Devonian age (Mid Famennian Stage), Solimões Basin

Photo 2. Dark gray diamictite containing subrounded and angular sand-, to cobble-sized clasts of great petrographic variety. Core from the bore hole 1-AM-1-AM (Nhamundá Formation) of Earliest Silurian age (Earliest Llandoverian Stage), Amazonas Basin.

Photo 3. Dark gray diamictite containing angular to subrounded sand grains, granules, pebbles and cobbles.Upper Devonian (Mid-Famennian Stage), Curiri Formation. At Cuiabá-Santarém Highway, about 7 km northward from the intersection with the TransAmazonian highway, Amazonas Basin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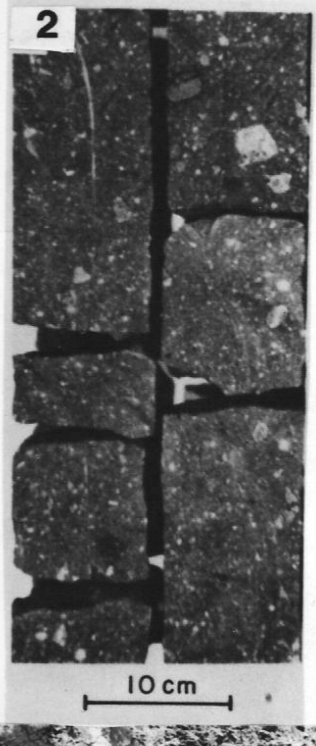
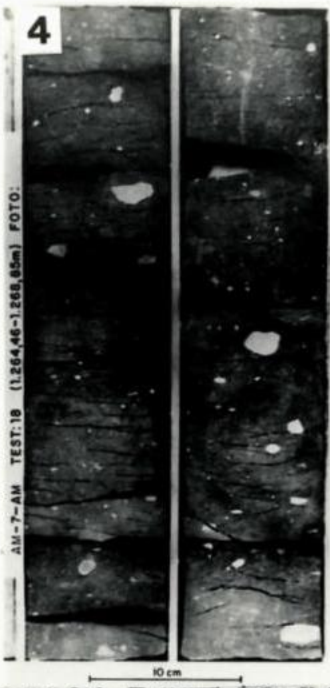

2
10 cm

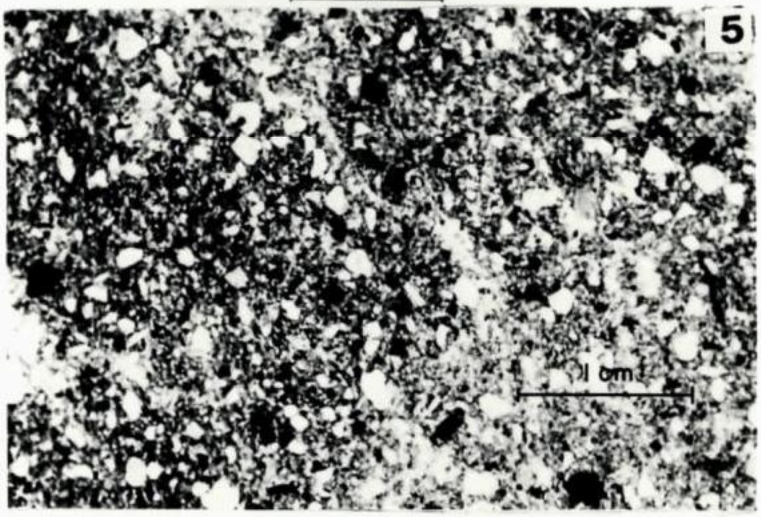

3

Photo 4. Dark gray diamictite, compact, micaceous rock with incipient bedding and with sparsely distributed sand-, granule- and pebble- sized clasts floating in a silty and clayey matrix. Core 18, bore hole 1-AM-7-AM. Curiri Formation (Mid-Famennian age), Amazonas Basin.

Photo 5. The thin-section from core 44 of the bore hole 2-PC-1-AM shows fairly fresh angular grains, many are corroded, and ranging from silt to coarse sand in size. They "float" in a silty and clayey matrix and are rarely in contact with each other. Curiri Formation (Mid-Famennian Stage), Amazonas Basin.

Photo 6. Faceted and striated stone composed of dark carbonate rock found within the Curiri Formation, Upper Devonian (Mid-Famennian Stage), Cuiabá- Santarém highway, about 7 km northward from intersection with the Trans- Amazonian highway.


4
AM-7-AM TEST:18 (1.264,46-1.268,85m) FOTO:
10 cm


5
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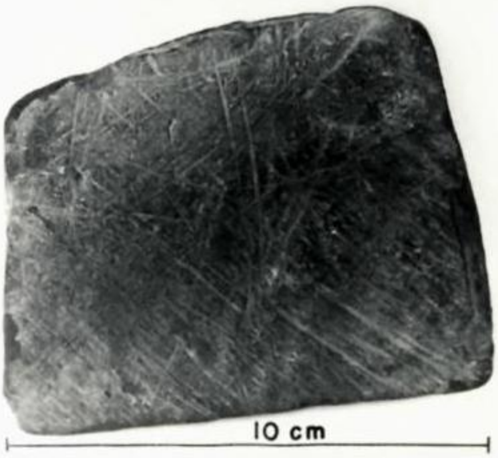
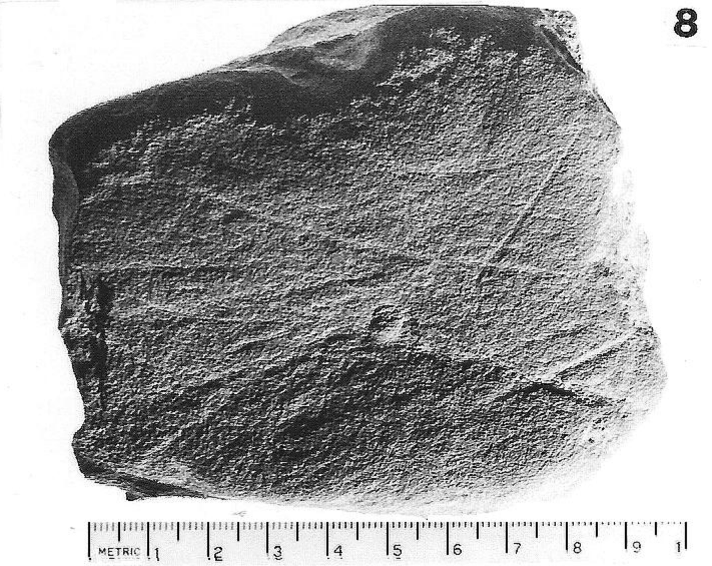

10 cm

Photo 7. Striated clast composed of dark carbonate rock with a flatiron shape from Curiri Formation, Upper Devonian. Cuiaba~Santarém highway, about 7 km northward from the intersection with the Trans-Amazonian highway

Photo 8. Non-parallel glacial striae on clasts from the Curiri Formation, Upper Devonian. Located at the Cuiabá-Santarém highway, about 7 km northward from the intersection with the Trans-Amazonian highway. Amazonas Basin

Photo 9. Fine-grained sandstone lens with strong soft-sediment deformation. It was possibly deposited in englacial channels (eskers). Bore hole 1-AM-1-M (Core 7). Curiri Formation Upper Devonian (Mid-Famennian Stage), Amazonas Basin.


METRIC 1 2 3 4 5 6 7 8 9 1
7


METRIC 1 2 3 4 5 6 7 8 9 1
8

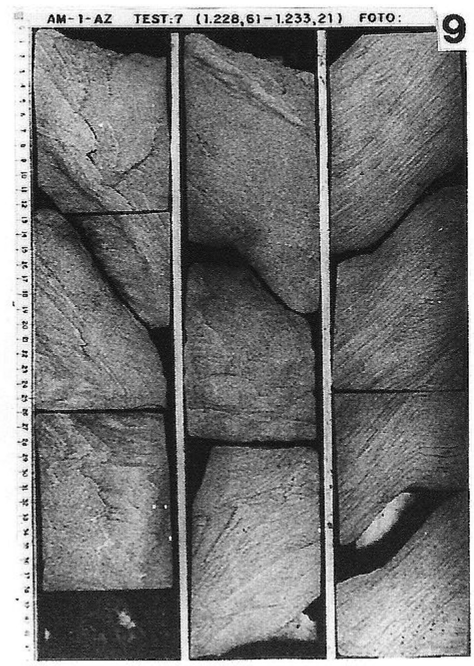
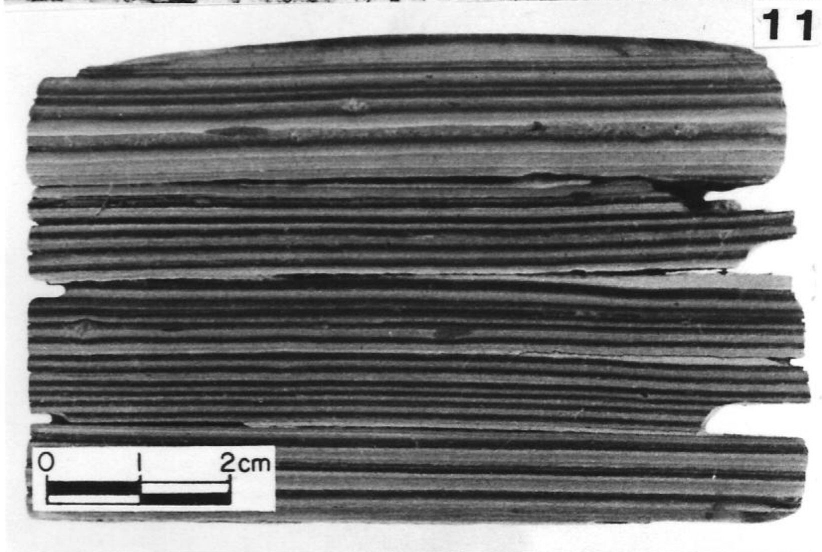

AM-1-AZ TEST:7 (1.228,6)-1.233,21) FOTO:
9

Photo 10. Cabeças Formation dark gray diamic-tite with very angular quartzitic cobble. Tocantins river cutbank, at Tataira, up river From the town of Carolina, Upper Devonian (Mid-Famennian Stage), Parnaíba Basin.

Photo 11. Rhythmically thin bedded varve-like sedimentary rock. Core 16 from the bore hole 1-TM-1-MA, Cabeças Formation (Mid-Famennian Stage) Parnaíba Basin.

Photo12. Rhythmically thin bedded core (varve-like), from well 1-TM-1-MA. Cabeças Formation (core 16) Upper Devonian (Mid- Famennian Stage). Note small dropstone (F). After Carozzi and others (1975). Parnaíba Basin


10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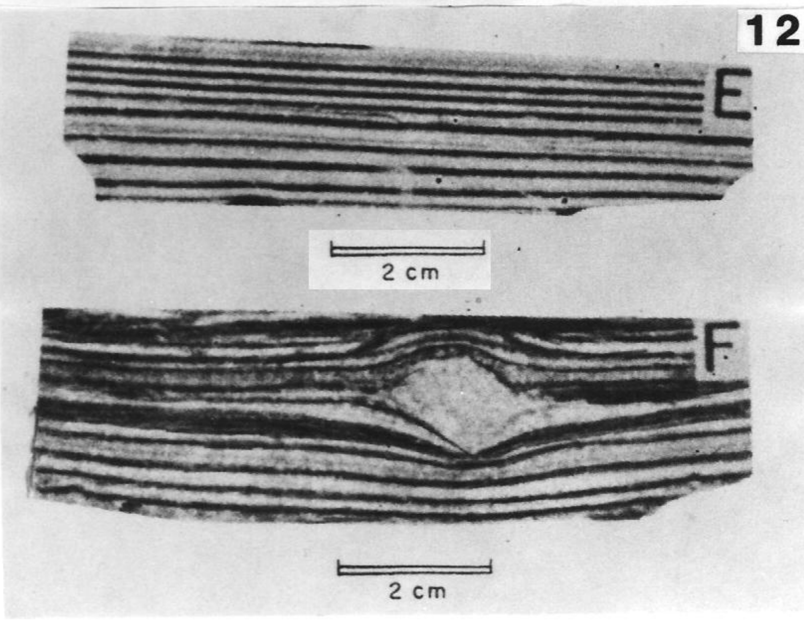
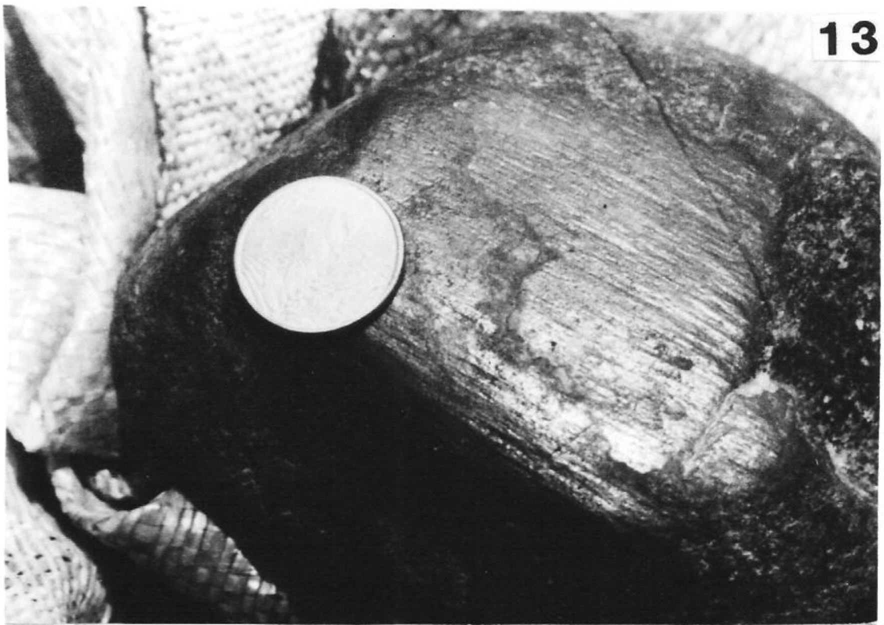

12
E
2 cm
F
2 cm

Photo 13. Glacially faceted, striated and polished cobble from Cabeças Formation (Upper Devonian, Mid-Famennian Stage), Western Parnaíba Basin, from diamictite in cutbank of the Tocantins River, about 20 km N of the town of Pedro Afonso. Coin is 2.5 cm in diameter.

Photo 14. Poorly stratified sandstone beds with scattered pebbles and boulders throughout, interpreted as deposited under glaciofluvial conditions. From Retiro, 5 km northward from the town of São Francisco do Piauí, in the Cabeças Formation, Parnaíba Basin.

Photo 15. Far-travelled pre-Silurian silicified conglomerate erratic 1.25 m long left by glaciers, at Morro Comprido, close to a striated pavement on the Cabeças Formation, observed in the next


13


14


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Photo 16. Glacial striations oriented N27ºE on the Cabeças Formation sandstone beds at Morro Comprido, at about 26 km from the town of Canto do Buriti, on the old road to the town of São Raimundo Nonato in the Piauí State. Parnaíba Basin

Photo 17. Detail of glacial striations on the Cabeças Formation at Morro Comprido, Parnaíba Basin

Photo 18. Glacial striae (oriented N15ºE) upon sandstone of the Cabeças Formation, Upper Devonian (Mid-Famennian Stage), Parnaíba Basin. Outcrop located about 1 km SE of road in the field and about 4.3 km along the E-W road from Canto do Buriti to Santa Iria.


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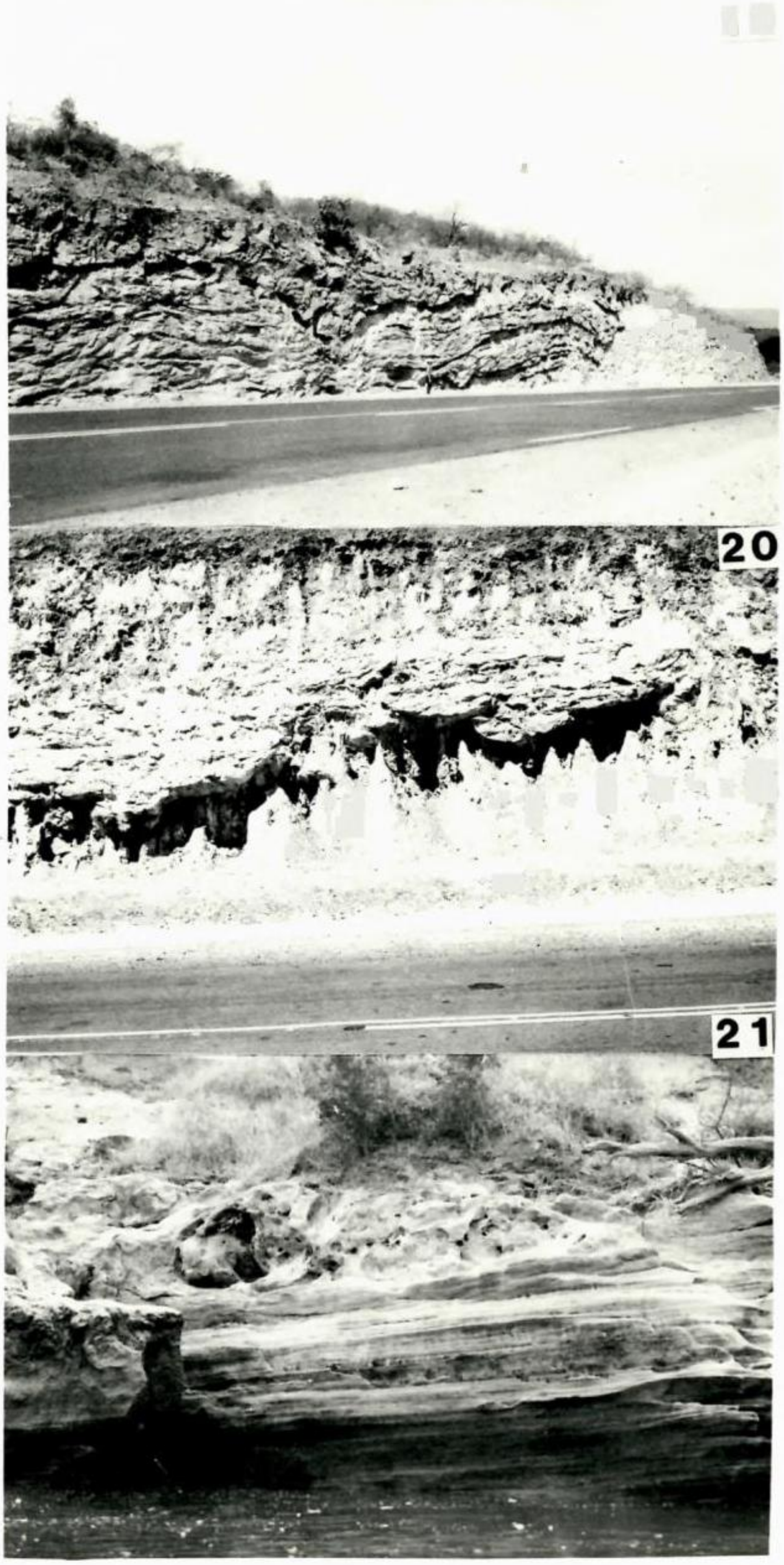
Photo 19. Slumped sandstone outwash beds. They may have resulted from collapse when ice underneath melted. In the middle level of the outcrop, a diamictite bed is present. These deformed features are characteristics of the Cabeças Formation. Two kilometers north of Oeiras, Piauí. Parnaíba Basin

Photo 20. Channel-like sandstone beds with "load" cast structures overlying argillaceous massive sandstone beds. These features are common in the Cabeças Formation. Road from Gaturiano to Valença (Piauí), Parnaíba Basin.

Photo 21. Chaotically deformed sandstone beds interpreted as a result of slumping when stagnant ice underneath melted. Cabeças Formation, Upper Devonian. Tocantins River cut-bank in front of the town of Pedro Afonso, Parnaíba Basin.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