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Hydrological Processes in 'undisturbed' forest

Climatic variability including ENSO events (droughts, extreme events)

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ABSTRACT

There are many uncertainties in the rainforests and climate variability relationships knowledge under tropical latitudes. This work proposes an up-to-date review focused on this subject through three axes : rainforests and tropical climate history, forests and climate relationships at low latitudes including vegetation modelling in GCMs and recent climate variability in forest areas. These three parts point out many interrogations and polemical aspects about an anthropogenic impact on vegetation and climate. A large knowledge about climate and vegetation interactions is indispensable for estimating human implications in climate and land cover changes, mainly through modelling studies.

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1 ANALYSIS OF THE HISTORY OF FORESTS AND ASSOCIATED CLIMATES

1.1 Methods and data

Previously, dense tropical forests were considered to be the most stable ecosystems on the planet, and their exceptional richness has often been, specifically associated with their resistance to past climate changes. Because of recent advances in the area of paleo-ecology it has been shown however that dense forests, such as those in Africa and in Amazonia, in fact have undergone profound changes in response to global climatic changes.

The history of dense forests and their dynamics can be reconstituted by the study of fossils such as pollens or, much rarer, wood or carbon fossils, within specific disciplines such as palynology, paleo-botany or anthracology. Reconstitution of paleo-vegetation is one method, among others, of reconstructing the paleo-climates of past eras. Nevertheless, even if much progress has been made in studying the history of intertropical forests and their associated climates, there are a number of problems that remain, mainly linked to the small amount of available data. This introduces subjectivity in describing the succession of paleo-environments.

Trees in dense tropical forests are practically all angiosperms. It would be logical to begin the history of these forests at the time angiosperms evolved, that is, during the lower Cretaceous era, to the Barremian and the Aptian eras, around 120 million BP (Maley, 1996a). Up until then, gymnosperms were the dominant form of plants but by the end of Cretaceous, dense tropical forests had become almost entirely made up of angiosperms. Pollen data have shown that African and South American dense forests were then quite similar and especially characterised by a great number of palm trees (Maley, 1990, 1996a). Palm tree species have remained abundant in South American forests while becoming relatively rare in African forests. It is since the upper Eocene (around 40 million BP) in particular, that the floral composition of these forests began to resemble their current state (Maley, 1996a).

To reconstruct the history of this vegetation requires the knowledge of many disciplines in the fields of botany and paleo-botany. The most often used of these disciplines is palynology and to a lesser degree, anthracology.

Palynology

Palynology is the study of pollen, which is the male component of the flower. Pollen is of microscopic size and is spread in great numbers by the stamen. The study of fossilised pollen in lakeside sediment or peat bog is often used to reconstruct the history of vegetation (Bonnefille, 1993). It is first necessary to extract sediment core samples. The various levels of a sample are dated using radiocarbon techniques. Then pollen is extracted using a physical chemistry procedure before it is placed between microscope slides. Because of its diverse morphology, pollen can be clearly identified under the microscope. By making a statistical count, the range of pollen types can be established, which thus give an idea of regional plant varieties over the course of time. Then, by comparison with current pollen samples and using special modelling techniques, scientists can develop a relatively precise reconstitution of the range and dynamics of plants which lived for thousands of years around the lakes or peat bogs studied (Jolly *et al.*, 1998).

Anthracology

Carbonisation preserves the fine structure of wood and carbonised wood can then be identified by botanists, because the ligneous species having an anatomical constitution specificity according to the make up and relative quantity of cellular elements (Tardy, 1998). The determination of fossil woods coming from radiocarbon dated soil samples allows the identification of specific taxas from flower trails. It also allows these taxas to be placed in the chronological context and so permit the reconstruction of the vegetation dynamics and the history of the plants (Vernet *et al.*, 1994).

Reconstructing past climates

In this area, the most spectacular advances have been made within the last twenty years in the polar and oceanic regions (Jouzel *et al.*, 1994). It is now possible to :

(i) determine the surface temperature variation of the Arctic and Antarctic polar ice caps and Greenland by analysing oxygen isotopes from deep ice core samples which date back over 400 000 years (Raynaud *et al.*, 2000),

(ii) measure the quantity of CO_2 in air bubbles trapped in ice samples and thus reconstitute the CO_2 component of the earth's atmosphere in parallel to the temperature variation over time (Raynaud *et al.*, 2000),

(iii) determine ocean temperatures through analysis of foraminifer plankton taken from marine core samples. These results have permitted scientists to show that major climatic variations are global in nature and follow patterns linked to variations in the earth's orbit around the sun.

Because of the complexity of the global climatic system, this data can not be extrapolated to the continents. For continental areas, such studies are concentrated on areas around lakes and river headwaters and on the analysis of pollen samples to determine the nature of past climates.

Problems in reconstructing the history of intertropical forests and associated climates

The continental climatic dynamics can only be deduced by comparing many pollen diagrams, accurately dated and studied in detailed resolution, in other words within a densely sampled study area (Bonnefille, 1986; Maley and Brenac, 1998; Jolly *et al.*, 1998). In Africa, this requirement causes some problems :

(i) the irregular distribution of test sites :

when there is only one available site for a very large region, it is difficult to extrapolate; also, it becomes necessary to use theoretical schematisation, which adds uncertainty to the reconstruction of paleo-environments;

(ii) the irregular time distribution of data :

there is much more information about the last 20 000 years than there is for the Tertiary era (Maley, 1996a).

The irregularity of data in space and time makes it difficult to propose a very precise distribution scheme for the various types of ecosystems in forests of the past. Nevertheless, the data we currently have for the past 20 000 years shows that main changes in the composition and distribution of tropical forests were synchronised with the major variations in the global climate, which leads us to think that this has been the case for millions of years.

1.2 The origin and history of intertropical forests

It would be difficult to use the current distribution of the major intertropical forest regions as a reference concerning the fluctuations of forest boundaries over the past millions of years. In the current study, a starting point could be when the continents that contain the forests of Southeast Asia, South America, and Africa attained their present proportions (*Fig.* 1).

The tropical forests of Southeast Asia

The tropical rain forests of Southeast Asia are, geographically speaking, relatively spread out, in particular, on numerous islands in Indonesia and New Guinea. It has been shown that these forests were at the outermost limits of their extension during the beginning of the Tertiary Period some time during the last 65 million years when they reached as far as Japan and China. Over the last 25 million years they have been progressively shrinking from their northern and southern boundaries (Heaney, 1991).

The rain forests of South America

The presence of a tropical forest closely resembling what we know today has existed since the Oligocene era, or about 36 million years ago. At this time the north-south boundaries of this paleoforest stretched much further, bearing witness to a hotter and more humid climate than exists currently. The range and diversity of this vegetation seem to have been the greatest during the Miocene era (Van der Hammen, 1991). Between 10 and 2 million BP (end of the Pliocene), there was the gradual formation of the Andes mountain chain, which reordered the continental drainage directions and gave the Amazon Basin its western facing contour. The morphogenesis of the South American continent, therefore, separated the forest into two zones (Choco and Magdalena Valley) and changed the zones of heavy rainfall in the western Amazon. During the upper Miocene era the forest progressively diminished, affected by the cooling of the global climate.

The African tropical forest

From the end of the Cretaceous to the Eocene, because of the shift of the African plate, the tropical forest extended well into what is now the Sahara desert and the northern Sudanian savannas. It was especially from the beginning of the Miocene that the forest attained its current position around the Gulf of Guinea (Maley, 1996a). The Ethiopian plateau was covered by forests from the Eocene until the Miocene period (Bonnefille, 1993). In eastern Africa, to the East of the Rift, the savannas began to expand in the Oligocene period (Hamilton and Taylor, 1991; Harris, 1993).

1.3 The historical and climatic framework of African tropical forests from the end of the tertiary period to the quaternary period

Starting in the Miocene period, the major variations in tropical forests can be interpreted in a global context of temperature variations and in particular, of cooling phases which were marked by the extension of the polar ice caps in the Arctic and Antarctic regions (Maley, 1996a).

The end of the Tertiary period

Between 15 and 10 million BP, following the increase of the Antarctic ice mass, the climate became dryer and cooler, with the ascent of the polar fronts and, progressively, a pattern of seasonal climates alternating from dry to humid. This aridification would have a direct impact on the African vegetation which opened and dried out (Bonnefille, 1993). A more humid period occurred between 8 and 6.5 million BP, associated with a new forest extension. Still, towards 5 million BP, came a significant expansion of the Antarctic ice mass which led to a drier period in tropical Africa accompanied by a new period of savanna expansion (Bonnefille, 1993). Then, once again, the climate became more temperate with oscillations between dry periods and humid periods but with a less overall effect than the preceding period, and a lot less effect than the period that was to follow, starting around 2.5 million BP.

The beginning of alternating ice age and interglacial periods : from 2.5 million BP to 20 000 BP

The study of pollen deposits from the Niger River delta (Morley and Richards, 1993) as well as the East African sequences (Bonnefille, 1993) show a new and significant expansion of savannas around 2.5 million BP. This major change occurred simultaneously with another important event : the first major ice age, which was marked by the extension of the polar ice caps into the arctic region of the northern hemisphere and, at the same time, by new glacial development in the Antarctic (Maley, 1996a). Then came a progressive increase in the magnitude of glacial variation, marked by two principle phases : the first occurred between 2.5 million BP and 800 000 BP, and was characterised by ice age/interglacial cycles of about 40

000 years; the second takes us up to the current era and is characterised by dominant cycles of about 100 000 years. These cycles are controlled by the main parameters of the earth's orbit as shown by Milankovitch (Mc Intyre *et al.*, 1989). This alternation of ice age and interglacial periods that began with an arid phase about 2.5 million BP would have a major impact on the forest ecosystems of the entire tropical zone.

The ice ages controlled the water level of the ocean; for example, the sea level was lower by 120 m around the year 18 000 BP, during the last glacial maximum. This entailed a concurrent change in the amount evaporating water surface. At the same time, global temperatures rose or fell (depending on which phase we were in, interglacial or ice age). This synchronous variation of water surface and temperature affects the quantity of water vapour in the air and leads to an increase or decrease of precipitation. On the continents, the resulting decrease (increase) in rainfall leads to an expansion (reduction) of the savannas and open areas or forests, depending on the conditions.

The periods of maximum forest expansion, comparable to the current situation, began towards the beginning of the Holocene about 9500 BP; and corresponded to the warmest climatic phases in which ice masses of both polar ice caps were reduced. The pollen study of marine core samples taken from the Gulf of Guinea show that the preceding phase of the African forest's maximum expansion occurred between 130 000 and 115 000 BP during the Isotopic stage 5e (Dupont *et al.*, 2000).

For about 800 000 years, it seemed, therefore, that the length of interglacial stages associated with the great expansion of dense forests corresponded approximately 10% of the time while the remaining 90% corresponded to the ice ages linked to the expansion of savanna areas (Maley, 1996a).

In tropical Africa, there are currently no pollen data that exist before 30 000 – 40 000 BP. On the other hand, in South America, in the Andes near Bogota, Columbia, there are long continuous pollen records dating back to the upper Pliocene (Hooghiemstra and Van der

Hammen, 1998). In Southeast Asia, 400 000 year old pollen records for the extreme south of China have been published (Zheng and Lei, 1999). In these two records, the fluctuations of forest vegetation and open areas were also in phase with global climatic changes.

The last ice age from 70 000 BP to the beginning of the Holocene

Although the study of the history of intertropical forests and associated climates during the Tertiary and the major part of the Quaternary is handicapped by a lack of paleo-data, the upper Quaternary, particularly from the last ice age, is relatively well documented.

From all tropical regions there is evidence that the Last Glacial Maximum was both drier and cooler than present climate. In some areas the desiccation is more obvious, and in others the cooling. There is clear evidence of depressions of forests limits, which may be interpreted as a reduction of mean annual temperature. According to Lezine (1998), during the Glacial-Interglacial transition at 15 000 then at 10 000 BP the regime was dominated by meridional exchanges. The Holocene is, to the contrary, characterized by low continent-ocean exchanges, with a dominance of the Atlantic monsoon fluxes.

Concerning the current forested regions of central Africa, pedological, geological and archaeological data shows that between 70 000 and 40 000 BP, this region was relatively dry and subject to intense erosion leading to the first generation of stone lines that are frequently found at the base of the soil. A second generation of stone lines appears in the period at the limit between the Pleistocene and Holocene around 11 500 years BP, coinciding with the first large increase of rains in the regions, until then, still dominated by open space vegetation (Maley, 1996b).

At the global level, the maximum cooling period occurred between 20 000 and 15 000 BP, characterised by major expansion of the polar ice cap into the northern parts of North America and Europe. In response to this global cooling, monsoons were dramatically reduced which entailed a severe reduction of the area of forests. Such reductions resulted in nothing more than a series of isolated forested areas, not far from the coast of the Gulf of Guinea and some others, near the centre of the Congo basin (riverine forests), and at the foot of the mountains of the African Rift (Maley, 1996a, 1997).

With regard to the second generation of stone lines mentioned above, it is interesting to observe that the driest phase, which was synchronous with the last major ice age, was not the most erosive phase, probably because the reduced rainfall was not of the erosive type (Maley, 1996b). It was calculated that the average temperature dropped by close to 4°C in one sector of the mountainous zone of the East African Rift (Bonnefille, 1991; 1993); in Barombi Mbo, in west Cameroon and in the region of Lake Bosumtwi, in Ghana, a temperature drop of about 3° C was estimated based on the lower altitude of certain mountainous taxas (Maley, 1991).

In South East Asia, West Pacific, New Guinea and Australia the general pattern is one that exhibits lowland rain forest in the Holocene, but some variation from this at the Last Glacial Maximum. Lower montane elements in some places suggest that the lowland climate at that time was both drier and cooler (Flenley, 1998).

In Africa the amount of mean annual temperature cooling is of 4±2°C, though a reduction of 5-8°C is still required by snow-line and forest limit data (Bonnefille *et al.*, 1990), if the latter are explained by temperature alone (Flenley, 1998).

In tropical Latin-America there are some evidences for desiccation during the Late Pleistocene in some places. In North-East of Brazil there are suggestions for an extension of the savanna between 22 000 and 11 000BP, at the expense of forest. The Amazon forest could therefore have been divided into blocks at the Last Glacial Maximum (Flenley, 1998).

The end of this last dry period initiated the beginning of a phase of forest regrowth, which, at the start of the Holocene, achieved its optimum range over the equatorial zone, including Southeast Asia and the Amazon.

The amount of such reduction is, however, of the order of 6-10°C, at least in Southeast Asia and the Western Pacific, and in Latin America.

The maximum forest extension during the Holocene. The chronological lag between the African and the Amazonian rain forests.

The start of the Holocene around 10 000 BP coincided almost exactly with the last phase of the maximum expansion of the rain forests in all the equatorial zones (Servant *et al.*, 1993). In South America, the history of the eastern part of the Amazon forest recorded at Carajas shows that after a first expansion between 10 000 and 8000 BP, the forest diminished considerably until about 4000 BP with the driest period occurring between 6000 and 5000 BP (Siffedine *et al.*, 1994). This drying of the climate created favourable conditions for recurring forest fires including in French Guyana until around 4000 BP (Tardy, 1998). In correlation to this, the level of Lake Titticaca in the Andes was relatively low in the middle Holocene, then towards 3800 BP, it rose abruptly by about 20 m to reach roughly its current level (Martin *et al.*, 1993). Shortly after 4000 BP began a new phase of forest expansion in Carajas and Guyana which lasted until modern times.

The situation was very different in central Africa. The expansion phase of forests began around 9500 BP in western Africa (Lake Bosumtwi) and central Africa (Lake Barombi Mbo) (Maley, 1991; Maley and Brenac, 1998). In western Africa the forest expanded continuously until present (Maley, 1991), but in central Africa, there has been a major interruption around 2800 BP in southern Cameroon and western Congo (Maley and Brenac, 1998; Maley *et al.*, 2000; Vincens *et al.*, 2000). Extremely dry conditions were present in these regions between 2800 and 2000 BP, facilitating the expansion of savannas and open spaces. At the same time there was a significant increase of pioneer taxas that would allow a rapid recovery of forests starting in about 2000 BP. According to the sites studied, the forest recovery and the succession of forest formations were not synchronous. For example, in the region of Lake Ossa near Edea, it was not until around 800 BP that the evergreen forest, rich in Caesalpiniaceae dominated again. (Reynaud-Farrera *et al.*, 1996). The configuration of the contemporaneous different forests types is largely the result of this former perturbation. The difference between the history of the South American and African forests are presented in *Fig.* 2.

The maximum extension of the African forest seems to have been synchronous with a sudden rise in the sea surface temperatures of the Gulf of Guinea (Maley, 1997). Monsoons pick up moisture from the eastern Atlantic and this rise in water temperature has the effect to sharply increase the water vapour pressure and ultimately increase rainfall in the neighbouring continent (see section 2). The variations occurring in South America, especially the expansion of open areas during the middle Holocene, were also connected to the sea surface temperatures variations which include such phenomenon as El Niño (Martin *et al.*, 1993).

The opposite behaviour between the western (Lake Bosumtwi in Ghana) and central parts (Lake Barombi Mbo in western Cameroon) of the forest domain is to be linked to sea surface temperatures variations and also to the particular monsoon features in the south Cameroon and Gabon areas (Maley and Brenac, 1998; Maley *et al.*, 2000).

For the period between 2800 and 2000 BP, rather than speaking of an "arid phase", the term "climatic pejoration" is used because this particular climatic phase appears to have resulted from an accentuation of the seasonality due to a shortening of the annual rainy season and, at the same time to an increase of disturbance lines with cumuliform clouds. This last feature is deduced from the heavy erosion which characterised this period (Maley, 1997).

So these large paleoenvironmental variations can not be imputed to the actions of Man but rather to global changes in the climate (Schwartz *et al.*, 2000). It would even seem that the perturbations which have occurred between 2800 and 2000 BP could have been the cause, or one of the main causes, of the Bantu migration through central Africa (Schwartz, 1992).

2 ANALYSIS OF THE MAIN RELATIONSHIPS BETWEEN FORESTS AND CLIMATE *VIS A VIS* CLIMATIC VARIABILITY

It is accepted that vegetation depends essentially on the climate. Its geographic distribution and its seasonal behaviour are quite largely influenced by rainfall, water being the main element of a plant. However, it has also been shown that the forest, within the earth-ocean-atmosphere system, has a significant impact on climate stemming from its use of the global system energy and its involvement in the water cycle.

2.1 Earth-ocean-atmosphere system energy

Earth radiation budget (ERB)

The earth-ocean-atmosphere system receives its energy from the sun in the form of short wave radiation. The sun's rays are reflected back into space in an infrared frequency bandwidth to maintain equilibrium of the radiation budget (*Fig.* 3).

The suns rays interact differently with the earth's atmosphere and with its surface (Polcher, 1994) :

(i) the atmosphere reflects a significant part of this radiation (26%) mostly because of clouds, and absorbs only a small amount (19%),

(ii) on the contrary, the surface of the earth reflects a small amount (4%) and absorbs a large portion (51%) of the short wave radiation received.

So the earth's surface absorbs over half of sun's rays and, to ensure its equilibrium, the earth reflects this energy in two different forms : a layer of moisture between the surface and the atmosphere engenders a flow of latent heat (linked to evaporation). This energy is called latent because it remains potential energy until the moment it condenses, at which point it is transformed into thermal energy, a flow of sensible heat ensures the diffusion of heat all along this thermal layer between the earth and its atmosphere (Polcher, 1994).

The heat dispersion that warms the lower atmosphere depends on climatic conditions but even more so on the surface conditions. The continental land surfaces are likely to have an immediate influence on the ERB (Fontaine and Janicot, 1993).

Climate sensitivity to surface

With the horizontal temperature gradients being weak in the Tropics, the atmosphere is very sensitive to land and ocean surface conditions (relief, albedo, temperature, humidity, vegetation), which influence the distribution and the intensity of heat sources and heat sinks, for which the atmospheric response is principally by vertical advection throughout the entire layer of the troposphere (Fontaine and al, 1998a, 1998b). These vertical movements develop notably within the deep convection systems, which with the associated atmospheric instability are capable of creating rain clouds and completing condensation of atmospheric water vapour in the convective systems (Bigot, 1997). Such deep convection is responsible for the majority of rainfall in tropical Africa and Amazonia. Contrary to conditions in the middle latitudes, the variations of surface conditions strongly influence vertical movement, with significant repercussions on diverging circulation systems such as those of Hadley and Walker (Fontaine *et al.*, 1998b).

The forest system, because of its great propensity for solar energy absorption and its capacity for evaporation, plays the role of an enormous energy converter : it absorbs more solar energy than any other plant surface. The forest system uses this energy to limit heating and to vapourise water that its root system extracts from the soil (Monteny, 1987). The resulting exchanges of energy that forests maintain with the atmosphere influence the physical air mass parameters of the atmosphere layer closest to the earth (Monteny and al, 1996). This role is linked to several properties, which Polcher (1994) catalogues as three characteristics that determine the sensitivity of the climate to surface processes :

(i) the density of the forest system is such that the *albedo* (the reflective power of a surface that measures the amount of solar energy given back to the atmosphere) is very weak compared to that of bare ground. Land clearing increases the portion of bare ground exposed to the sun's rays and therefore increases albedo,

(ii) the high rate of *evaporation*, comparable to that of oceans, is one of the main characteristics of forests whose leaf density allows them to intercept and re-evaporate a large part of rainfall. The root systems of trees allow them to extract water from a greater portion of the soil than could be done in any other surface system,

(iii) the *surface variation* caused by the different heights of trees that make up the forest generates turbulence, which is favourable to triggering precipitation.

The two key parameters seem to be albedo and moisture, closely linked, since wet ground, whether covered with vegetation or not, has a weaker albedo and greater evaporation capacity than the same ground, bare and dry (Fontaine and Janicot, 1993). The degree of surface variation is a more delicate parameter to study since it has been shown that on a smooth surface such as a pasture, the presence of a few isolated trees generate more turbulence than a whole forest. Therefore, we will focus on the analysis of the first two parameters, the albedo, determining the capacity of forests to absorb solar energy, and the involvement of evaporation in the water cycle.

2.2 The role of the albedo in plant-atmosphere interaction

The threat of forest removal and eventually the destruction of all tropical forests has lead to climatologists to ascertain the climatic impact of such changes to the earth's surface (Gash and Shuttleworth, 1991), all the more so because computer models have affirmed the sensitivity of the climate to surface processes (Polcher, 1994).

Albedo is historically the first parameter to have been taken into consideration and connected to recent forest conversion. Charney (1975) proposed a now famous mechanism that illustrates the retroaction linked to albedo and its influence on the regional climate. Charney's work (1975) showed that an increase in albedo after a reduction in surface vegetation (through a greater amount of exposed bare ground) brought about a decrease in net ground back radiation (that is the preponderant term of the energy balance for the study of heat and mass exchange (Monteny *et al.*, 1996). Thus a decrease in the sum of sensible and latent heat flows results. Because of this, a column of atmospheric air would be cooled and this heat loss would be compensated adiabatically by subsidence (Fontaine and Janicot, 1993) and thus contribute to a reduction of precipitation. In turn, the latter will react positively on the first part of Charney's mechanism by decreasing the amount vegetation on the surface (*Fig.* 4).

Charney's work (1975) gave rise to numerous criticisms; for example that his mechanism did not in any way incorporate the role of water and especially humidity of the earth's surface. This role is essential in tropical regions covered by dense forest, where the forest is closely integrated in the water cycle. A diminishing albedo and therefore a higher loss of solar energy at the surface means a significant reduction in energy available for evapotranspiration; thus the recycling of water is disrupted, which in turn, causes a decrease in the water vapour contribution from the continental land masses (Brou Yao, 1997).

2.3 The role of water and the involvement of the tropical forest in the water cycle

Indeed, by evapotranspiration, vegetation recycles moisture locally and influences the regional distribution of precipitation (Bigot, 1997, Bonell, Callaghan and Connor, this volume). In the monsoon regions that characterise a large part of the tropical zone, the quantities of water precipitated on the continent come from the condensation of water vapour accumulated in the mass of ocean air as it passes over the ocean. The concentration of water vapour in this air mass which then crosses the continent depends on the evaporation process of the vegetation-atmosphere interface. It has been shown in central Africa (*Fig.* 5) that a large part of moisture transfer into the atmosphere, which is generated by evapotranspiration, contributes to the formation of cloud systems (Bigot, 1997) and that the rainfall associated with these convective systems depends not only on the monsoon flow but also on the recycling of moisture by the forest (Cadet and Nnoli, 1987).

Monteny (1987) described the importance of this water recycling in supplying water vapour to this monsoon flow through the study of favourable conditions for the creation of dense African forests, conditions which permit water recycling and movement towards regions further to the north. Brou Yao (1997), after Monteny (1987), writes that the tropical rainforests of southern Ivory Coast inject into the atmosphere the equivalent of 55-75% of the annual precipitation. His work on the Ivory Coast underscored the important role of the forest concerning potential rainfall. This role is twofold, since the forest system is both a receiver of precipitation (especially monsoon rains); and a generator of rainfall by means of evapotranspiration and fine scale processes (Bigot, 1997). In contrast Gong and Eltahir (1996) from modelling gave a lower estimate of recycled precipitation over West Africa : 27%. This ranges the question of our understanding of the precipitation process and its recycling over West Africa, which is still very difficult to measure and to take correctly into account in models.

The variation of available moisture at the surface is added to the effects of changes in albedo in the sense that it has a direct link with the ERB. Charney *et al.* (1977) reformulated his theory on the increase of albedo and the retroaction of vegetation; the new theory included the role of water and especially changes of the Bowen's ratio which establishes the relationship between sensible and latent heat flows (sensible heat flow density divided by latent heat flow density). The results of Charney *et al.* (1977) were similar to those formulated by the 1975 theory. Subsequently Mylne and Rowntree (1992) proposed a mechanism which linked Charney's process to ground moisture (*Fig.* 6). Essentially a decrease in precipitation induced by an increase of albedo (Charney's mechanism) also brings about a decrease in ground moisture. This surface drying leads to a reduction of evaporation through an increase in soil resistance. Bowen's ratio is also lowered since the latent heat flow is reduced, which means lower air humidity and consequently, less precipitation. The sensible heat flow and the soil temperature rise because of this surface humidity decrease, all of which leads to a decrease of net radiation and an amplification of Charney's mechanism (Mylne and Rowntree, 1992). This leads to a reduction in the total of heat flows (sensible and latent), leading to a diminished contrast between land and ocean. On the contrary dense forest increases this net radiation at the surface by the combined effect of weak albedo and high surface humidity, and increases the sum of sensible and latent heat flows. These flows supply the humid static energy (HSE) in the boundary layer and reinforce the HSE layer between earth and ocean which is the driving force behind the circulation of monsoons (Zheng and Eltahir, 1998).

Highlighting the forest system's contribution to the regional climate leads naturally to a concern about the possible impacts of the massive forest clearance that is the current trend, along with the increased density of agricultural production and the acceleration of land clearing activities. Nevertheless, the impact of such actions is extremely difficult to evaluate and gives rise to considerable uncertainty; because the cause and effect links between the forest and climate are not yet clearly understood. (Brou Yao, 1997).

2.4 Evaluating the climatic impact of forest conversion : modelling global terrestrial vegetation-climate interactions

Even if it is accepted that decreasing forestlands affects, in theory, the climate (Salati and Nobre, 1991), it is difficult to evaluate the impact of extensive forest clearance on the climate, thus we are obliged to use climate simulation models. But large scale modelling is not without its problems given the significant uncertainty that exists concerning the relationships between climate and vegetation. This topic is likely to be debated thoroughly within next years, as global or regional circulation models' resolution are improved.

According to Bigot (1997) the inadequate understanding of the relationships between climate and vegetation is principally due to : an incomplete knowledge of average rain fields in tropical forest regions (mainly in Africa); seasonal vegetation activity and the variability of these two items as a function of climatic anomalies. Accurate parameterisation of vegetation, that is, the resistance in heat and water vapour transfer between soil and atmosphere, the amount of surface variation, albedo, and the interception of a portion of precipitation, are indispensable to realistic modelling and require multiple field campaigns to gather specific measurements (Fontaine and Janicot, 1993).

Recent climate simulations

Predictions of global atmospheric models are highly sensitive to prescribed large-scale changes in vegetation cover, such as removal of tropical forests (Henderson-Sellers *et al.*, 1993; Polcher and Laval, 1994; Zheng and Eltahir, 1997).

The majority of recent forest clearance simulations by GCMs indicate what a considerable influence the disappearance of forest cover would have, eventually, on tropical regions. Polcher (1994) expands on Charney's hypothesis by showing that a reduction in sensible heat flow, which could be caused by forest conversion, seems to affect the number of convective events and thus to reduce local precipitation. Zheng and Eltahir (1998) support the hypothesis of the dramatic influence of forest clearance on the north coast of the Gulf of guinea, which would lead to reduced rainfall and a weakening of moisture convergence in all West Africa.

But simulation results from climatic models still yield contradictory conclusions and some of these simulations have shown that the complete disappearance of the Amazonian forest, while having significant consequences on albedo, would bring about only a slight decrease in local precipitation or evapotranspiration (Bonnefille, 1993; Bonell, Callaghan and Connor, this volume).

Past climate : ellipsoid, albedo and biospheric feedbacks

For validating the hypothesis, Foley *et al.* (1994) suggested the investigation of past environments such as the climate of the early to middle Holocene, some 6000-9000 years ago, for which strong differences in global vegetation pattern are amply documented. According to Claussen (2001), most researchers do not agree on the relative importance of biospheric feedbacks on climate. Moreover, Claussen (1994) discovered the possibility of multiple equilibria in the 3-dimensional atmosphere-vegetation system, which seems to be specific to the subtropics and particularly to North Africa, as the high-latitude climate-vegetation system is much more stable (Levis *et al.*, 1999). Two solutions seem to be possible : the arid, present-day climate and a humid solution resembling more the mid-Holocene, some 6000 years ago, with a Sahara greener than today. Simulations with mid-Holocene vegetation yield only one solution, the green Sahara (Claussen and Gayler, 1997), while for the Last Glacial Maximum (LGM), some 21,000 years ago, two solutions exist (Kubatzki and Claussen, 1998). Brovkin *et al.* (1998) show that for the present-day climate, the green equilibrium is less probable than the desert equilibrium, and this explains the existence of the Sahara desert as it is today. The difference in albedo between desert and vegetation cover appears to be the main parameter that controls an existence of multiple stable states. Claussen (1997), Claussen and Gayler (1997), and Claussen *et al.* (1998) explain this positive feedback by an interaction between the high albedo of Saharan sand deserts and the atmospheric circulation as hypothesized by Charney (1975), whilst Texier *et al.* (1997) suggest an additional feedback between sea-surface temperatures and land-surface changes. Claussen *et al.* (1998) found that the velocity potential patterns, which indicate divergence and convergence of large scale atmospheric flow, differ between arid and humid solutions mainly in the tropical and subtropical regions. It appears that the Hadley-Walker circulation slightly shifts to the west. For the mid-Holocene boreal summer, the large-scale atmospheric flow is already close to the humid mode, even if one prescribes present-day land surface conditions. This is cause by differences in insolation : in the mid-Holocene boreal summer, the Northern Hemisphere received up to 40 W/m^2 more energy than today, due to a change in the eliptoid orbit of the earth around the sun, thereby strengthening African monsoon (Kutzbach and Guetter, 1986). During the LGM, insolation was quite close to present-day conditions.

After using a coupled atmosphere-vegetation-ocean model, Ganopolski *et al.* (1998) conclude that in the subtropics, the biospheric feedback dominates, while SST adds only a little. Claussen *et al.* (1999) clearly show that subtle changes in the seasonal eliptoid orbital forcing triggered changes in North African climate. Such changes were than strongly amplified by biogephysical feedbacks in this region leading to a rather fast desertification within a few centuries starting around 5,500 years ago. This seems to be in agreement with palaeogeological reconstructions (Petit-Maire and Guo, 1996).

De Noblet *et al.* (2000) compared two GCM (LMD 5.3 and ECHAM 3) asynchronously coupled to an equilibrium biogeography model to give steady-state simulations of climate and vegetation 6000 years ago, including biogeophysical feedback. They found surprisingly different results of simulation of climate and vegetation 6000 years ago, neither of both GCMs being fully realistic and both being unaffected by the choice of green or modern initial conditions, due to an inadequate strength of the tropical summer circulation in the GCMs. Such results highlight the importance of correct simulation of atmospheric circulation features for the sensitivity of climate models to changes in radiative forcing, particularly for regional climates where atmospheric changes are amplified by biosphere-atmosphere feedbacks.

For Claussen *et al.* (2001) biogeochemical and biogeophysical processes do not operate independently, and being triggered by large-scale land cover changes oppose each other on the global scale. Tropical forest conversion tends to warm the planet because the increase in atmospheric CO_2 and hence, atmospheric radiation, outweighs the biogeophysical effects.

Thus the sensitivity of the tropical climate to forest conversion remains open to debate. The change in surface energy transfer exchange between latent energy and sensible heat flows, linked to this degradation of forest cover could have an impact on precipitation but at the present time we are not able to clearly evaluate its magnitude (Polcher, 1994).

Indeed, even if 50 – 70% of the western African forest lands have been transformed into agricultural use or left fallow, the magnitude of this transformation does not seem to have yet affected the regional climate significantly (Bonnefille, 1993).

Elsewhere, according to Bonnefille (1993), the recent "desertification" has resulted more from a change in rainfall distribution, than from an reduction in the total amount of rainfall. This leads the author to deduce that the regional climatic variations that have been highlighted by the historical data are not only the result of human activities, but can be attributed to a large degree to *natural climatic fluctuations*.

3 CLIMATIC VARIABILITY IN TROPICAL FOREST REGIONS

3.1 Definition

Climatic variability is defined as being the distribution of climatic elements around their average values calculated over 30 years; this natural variability is an intrinsic character of climate (Janicot, 1995). A major objective of climatology is to better understand the variations in the succession of average states of the atmosphere and to forecast these climate variations, particularly in the tropics, where climatic variability has a profound impact on the lives of the people there and on the evolution of ecosystems (Fontaine *et al.*, 1998b).

3.2 Climatic variability and impact on rainfall and runoff in West and Central Africa

Bonell (1998) showed that only a small section of the hydrological community was studying climate change and its linkages with hydrology. This is mainly due to uncertainties in climate projections from GCMs, insufficient knowledge of land-cover changes, and inconsistency in scale between the too large resolution of GCMs and the small scale focus of experimental hydrology and management of water resources. Nevertheless, significant results has been obtained in West and Central Africa when concerning rainfall and runoff variability and links with climatic variability (Mahe and Olivry, 1995; Bricquet *et al.* 1997; L'Hote and Mahe, 1996; Servat *et al.*, 1998).

Mainly annual and monthly time series have been analysed, since the beginning of the century for hundreds of raingauge stations over regional areas (Wotling *et al.*, 1995; Servat *et al.*, 1997; Paturel *et al.*, 1998; Mahe et al, 2001), which showed that the recent drought period not only affected the Sahel, but was also responsible for a decrease of rainfall over the more humid parts of tropical and equatorial Africa along the Gulf of Guinea. Statistical tests for detection of discontinuities in times series have been applied (Hubert *et al.*, 1998) and the results showed that the discontinuities in rainfall time series were often observed around the year 1970, with some regional variability (*Fig.* 7).

Runoff series have been also studied : annual and monthly flows, and low flows (Servat and Sakho, 1995; Aka *et al.*, 1996; Servat *et al.*, 1997; Mahe and Olivry, 1999; Laraque *et al.*, 2001), are also decreasing since 1970. Seasonal floods are modified : in tropical West African basins their magnitude is lower and their rise and decrease are more rapid since 1970, which is partly due an increase of the recession coefficients (Bricquet *et al.*, 1997; Mahe *et al.*, 2000). In equatorial Africa one major impact of climatic variability on equatorial river hydrological regimes is to shorten and reduce the magnitude of the boreal spring flood since 1970, what is particularly evident in the case of the Ogooue and the Kouilou rivers (Mahe *et al.*, 1990; 2000) (*Fig.* 8). These studies showed that the runoff variability is greater than that of the rainfall, mainly in the case of tropical and Sahelian rivers and most likely due to a decrease of groundwater levels. The same observations apply also for a lot of equatorial rivers, except for special cases where the aquifer plays a major buffering role (Laraque *et al.*, 2001). Figure 9 shows rainfall and runoff decadal variability over eight regional areas. Over the period 1971-1989, the rainfall decrease was stronger (-25%) in the northwestern West Africa (Senegal, Guinea, Mauritania). Over Central West African areas the rainfall decrease was smaller (-15%). The rainfall deficit diminished towards the equator that is : -11% over Cameroon, -4% over

Gabon and Congo, and –3% over the Zaire River basin. The runoff decrease also decreased from the Sahel to Equator : from –60% for the Senegal region, to –11% for the Zaire River. In addition, good relationships between rainfall, low flows and groundwater levels inter-annual variability has been showed for the Bani tropical river in Mali (Mahe *et al.*, 2000).

Adaptations of river flow simulation models has been developed, mostly at the monthly time step, or sometimes at lower time scales (daily or decadal) (Servat *et al.*, 1990; Servat and Dezetter, 1991; 1993; Paturel *et al.*, 1995), which have helped study the variability of flow regimes on tropical basins. A recent study (Mahe *et al.*, 2002) showed that the river flows of a Sahelian basin have been significantly increased since the changes in land cover (increase converted areas for agriculture), despite the reduction of rainfall. Simulations of river flows were improved by modifying the annual values of the soil water holding capacity, by reference to the satellite-observed land cover changes during the period.

In Ivory Coast, Brou Yao *et al.*, (1998) observed that the coffee and cocoa productions areas shifted following the rainfall changes, and that the associated forest conversion might have had a significant impact on local precipitation occurrences.

The diminution of rainfall since 1970 has been well described. It concerns the whole West Africa, and Central Africa to a lesser extent. Runoff diminution is amplified by the groundwater level decrease, which causes less groundwater to participate in the surface runoff during the flood, and which causes also an increase of the speed of the flood recession (Mahe *et al.*, 1998). The changes in land-cover seem to have a great impact on the hydrological cycle and the rainfall-runoff relationships. The impact of the forest clearance on the rainfall/runoff relationships seems to be dependent on the type of climate/vegetation system. In Sahelo-Sudanian areas, the forest clearance, associated with an increase of agricultural activities, rapidly induces a destructuration of the top layer of the soil which infiltrability decreases, and runoff coefficients increase, even during a dry phase. Whereas in equatorial and tropical humid areas the major impact of the forest conversion is to reduce the local evapotranspiration, thus reducing the total amount of available water vapour for precipitation recycling.

3.3 Sensitivity of the tropical climate to surface conditions

In the lower latitudes, the seasonal variation of temperatures is slight. The organisation of average tropical and subtropical climates is therefore most often a function of total rainfall and seasonal rhythms creating alternate dry and wet seasons, and thus reflecting the activity and variations of the water cycle in the atmosphere. This water cycle activity is expressed in terms of evapotranspiration, water vapour flows and precipitation (Fontaine *et al.*, 1998a). The surface conditions, especially the plant cover of dense rainforests and the upper layer of oceans have a significant effect on the atmospheric water cycle and also on the vertical movements within the tropical atmosphere (see section 2.3).

The space-time variability of climate principally depends on the interaction between the surface conditions (temperature, albedo, humidity) and the atmosphere : linkage manifested by wind pressure and sensible and latent heat flows. Recent researches in climatology focus on the heat content of the mixed global ocean layer, in other words, the upper layer subjected to the action of surface winds and in which the thermal gradients are weak (Fontaine *et al.*, 1998a).

3.4 The ocean-atmosphere linkage

Covering 70% of the earth's surface, the oceans constitute its largest water reservoir and an important reservoir of energy in tropical latitudes, thus playing a major role in driving general weather circulation (Fontaine and Janicot, 1993). Moreover, and contrary to the continents, oceans possess a strong thermal inertia, which makes the Sea Surface Temperature (SST) the most influential variable on the atmosphere and at the same time, an indicator of land climatic variability (IPCC, 1996).

As noted by Bigot (1997), the three essential parameters controlling the ocean-atmosphere linkage are :

(i) SSTs (thermodynamic estimation of the ocean-atmosphere interface),

(ii) precipitation (physical estimation of one element of the atmospheric water cycle),

(iii) wind (dynamic estimation of changes in atmospheric movement).

Even if the ocean's annual thermal amplitude is slight, due to its weak specific heat, the linkage is intensified by the release of latent heat. This energy release influences surface winds, which then change upwellings, the ocean drift currents, and thus the SSTs (Bigot, 1997).

SST variations determine significant atmospheric responses (Fontaine *et al.*, 1998b) and moreover, the ocean's strong thermal inertia defines the characteristic time steps of the ocean-atmosphere linkage, and accordingly, those of low latitude climatic variability (Fontaine *et al.*, 1998a). This SST influence is also found seasonally, where the monsoon pattern is its most obvious manifestation (Fontaine, 1991;Fontaine *et al.*, 1991).

3.5 Interannual and multiannual scale : the Southern Oscillation

One of the most important expressions of the ocean-atmosphere linkage is the phenomenon El Niño-Southern Oscillation (ENSO) in the equatorial Pacific (Folland and al, 1998; Fontaine *et al.*, 1998a; Bigot, 1997). Although the consideration of oceans in the climatic equation is rather recent (since the beginning of the 1980s), the fact is that oceans are now the object of innovative research and have a special place in the global scientific arena, being the main element of the earth climate and playing a major role in the earth climatic variability. This is attributable to the increased intensity of ENSO related events, in particular, that of 1982-1983. Unusual in its intensity and singularity, this event was an "extraordinary catalyst" for research and permitted scientists to confirm ENSO as an important element in the ocean-atmosphere linkage on an interannual and multiannual basis (Fontaine and Janicot, 1993).

ENSO, dominant cause of interannual and multiannual climatic variability in the world

ENSO is defined as the dominant cause of interannual and long term climate variability in the world (Trenberth, 1997; Callaghan and Bonell, this volume), particularly in the tropics where it influences the rainfall in many countries : in India, eastern and southern and north-eastern Africa as well as in Australia (Janicot, 1999). Indeed, few tropical regions are not affected by ENSO (Fontaine and Janicot, 1993) and this phenomenon has a considerable and continuous impact on temperature ranges and precipitation, not only around the Pacific and Indian Ocean regions but also in the Atlantic regions where ENSO has been associated with SST and trade wind variability in the tropical Atlantic (Hastenrath *et al.*, 1987; Nicholson and Kim, 1997; Fontaine *et al.*, 1998a).

The relationship between ENSO and interannual climate variability in Central Africa was well documented by Bigot (1997) who associated ENSO with a decrease of cloud convection and precipitation, caused mainly by a slow decrease of westerly winds that transport water vapour. Studies which came to the same conclusions were also performed in Guinean Africa (Janicot and Fontaine, 1997), tropical Amazonia (Fu *et al.*, 1999), Uruguay and Brazil (Diaz *et al.*, 1998), and indicate a notable reduction of rainfall in tropical forest regions during the warm phase of the Southern Oscillation.

The evolution of ENSO events and their repercussions in the tropics since 1970

Since the 1970s, the intensity and frequency of ENSO events have changed, as well as their impacts on tropical regions. Indeed, we can observe, starting around 1975, an increase in the magnitude of positive anomalies in the Southern Oscillation Index (SOI), occurring with increasing frequency. Similarly, a number of studies have shown an evolution in the correlations between anomalies of SST fields in the equatorial Pacific and precipitation in tropical regions (Nicholson and Entekhabi, 1986; Moron *et al.*, 1995; Janicot *et al.*, 1996, Semazzi, 1996).

In Central Africa, we can see an increase in the amplitude of annual rainfall during ENSO events after the 1970s, notably in 1977, 1982 and 1987 (Bigot, 1997). A study of the precipitation impact in western Africa of ENSO events created some controversy because of the high inter-annual variability in the evolution of correlation of the SOI and western African rainfall (*Fig.* 10). Although these correlations were not significant before the 1970s, they have become important in the last 25 years, correctly indicating weak rains in the Sahel region after an ENSO warm phase and heavy rains after an ENSO cold phase (Janicot *et al.*, 1996).

This evolution in relationships between ENSO events and Sahelian climate variability can be explained first, by an observed increase in ENSO's inter-annual variability, then by an increase in the number of major warm events over the past twenty years and lastly by the positive correlation between rainfall and the ten year global SST variations.

The relationships between the Southern Oscillation and precipitation in tropical regions tends to show the links between SST variations in the Pacific and those recorded in the Atlantic Ocean. During an ENSO event, there is a positive correlation in SST anomalies in the north of the tropical Atlantic with a delay of a few months (Uvo *et al.*, 1998) and there are also some complex interrelations, not yet well understood, between sea surface temperature variability of the two oceans, Atlantic and Pacific, and rainfall in tropical Africa (Bigot, 1997; Janicot and Fontaine, 1997) as well as tropical South America (Fu *et al.*, 1999; Diaz *et al.*, 1998).

The particular geometry of the Atlantic Ocean, which is very wide near the equator and narrower in high longitudes introduces other phenomena linked to ocean dynamics and gives special importance to meridional thermal gradients. Tropical Amazonia, like western and central Africa, situated in the Atlantic perimeter, are subject to the double influence of Atlantic and Pacific phenomena which affect both the Hadley and Walker circulations (Fontaine *et al.*, 1998a). Recent studies have also showed signs of multiannual and decadal variations specific to SST anomalies in the Atlantic Ocean (Janicot, 1999; Fontaine *et al.*, 1998b), separate from the impact of ENSO.

3.6 An approximate ten-year cycle : the tropical Atlantic

Numerous studies (Bigot, 1997; Fontaine *et al.*, 1998a, 1998b, 1999) have distinguished two regional modes of SST variability : in the north Atlantic and in the southern equatorial Atlantic The natural variation of these two modes induces almost ten year meridional gradient fluctuations in the tropical Atlantic SST anomalies which dominate one portion of the long term surface thermal structure of this Ocean (Fontaine *et al.*, 1998b).

The Atlantic thermal dipole has a significant impact on the variability of precipitation on neighbouring continents. It is associated with the atmospheric circulation in the equatorial and south Atlantic and has a major influence on the positioning of the monsoon shearline (see Callaghan and Bonell, this volume), which modulates a large part of tropical precipitation and defines the seasonal rain cycle in regions such as western Africa. When the tropical north Atlantic is abnormally warm and the south Atlantic abnormally cool, the northern monsoon shearline tends to migrate farther to the north during the rainy season. Conversely, when the tropical north Atlantic is cooler and equatorial and south Atlantic are warmer, the southern monsoon shearline is the most active (Janicot, 1999).

This modulation and meridional movement of the monsoon shearlines influences precipitation in tropical countries that have monsoon seasons. In western Africa, weakening of south and equatorial Atlantic circulation, characterised by a weak St Helene's anticyclone -sometimes shifted towards Brazil-, a warming of south Atlantic waters and reduced equatorial and coastal upwellings, results in a particularly dry season in Sahel (Fontaine, 1991; Mahe and Citeau, 1991; Mahe, 1993). Similarly, studies on central Africa by Bigot (1997) affirmed a major influence of this meridional Atlantic thermal gradient on annual rainfall in the Congo and Gabon, via its influence on the convergence zones meridional position and the associated conveyance of moisture over central Africa. Wotling *et al.* (1995), found good relationships between West Arican rainfall variability and Atlantic SST's.

Moreover, this decadal variability of the SST has amplified ENSO's impact in different areas of the globe, by increasing the correlation between precipitation and ENSO events in central Africa. The same phenomenon has also affected the cumulative rainfall in tropical Amazon's rainy season (Fu *et al.*, 1999). However, the relationship between the Pacific and Atlantic Oceans' SSTs are still fuzzy, on the one hand, because the cause and effect links remain difficult to discern (Diaz *et al.*, 1998); and on the other, because the Atlantic falls more into an intrinsic mode of SST variability more identified with multi-decadal intervals (Fontaine and Janicot, 1993).

3.7 The multi decadal scale : the inter-hemisphere reversal

On a larger area scale, studies based on recorded SSTs all over the globe during the past 20 years show that the thermal structure of the surface is dominated by a third type of variability, a multi-decadal type that affects the gradient between the two hemispheres and, at the same time, the zonal gradients on the equator through the relative warming of the southern and equatorial basins of the Atlantic Ocean as well as the Indian Ocean (Fontaine *et al.*, 1998a). This mode of low frequency variability describes a structure that associates warm (cold) anomalies in Southern hemisphere oceans and the Indian Ocean with cold (warm) SST anomalies in the north Atlantic and north Pacific oceans (Fontaine and Janicot, 1993). The same mode of variability, dominated by a slow inter-hemispheric reversal, is statistically independent of the two other regional modes, described above. The latter defines fluctuations of the Atlantic meridian thermal gradients in cycles of almost ten years. (Fontaine *et al.*, 1998b).

When the large, dipolar, multi-decadal system experienced warm anomalies in the south or cold anomalies as far as 30° north latitude, a correlation was made with dry periods in western Africa (Folland *et al.*, 1986, Fontaine *et al.*, 1998b; Fontaine and Janicot, 1993). The large scale low frequency oceanic influence is then associated with other regional fluctuations, which thus explain in large part the variability of rainfall in western Africa (Fontaine *et al.*, 1998b). Studies in Guinean Africa (Janicot and Fontaine, 1997) and central Africa (Bigot, 1997), linked the general influence of tropical forest region rainfall to the change in SSTs since 1945. Such changes corresponded to ocean warming in the Southern hemisphere and Indian Ocean, and simultaneous cooling of oceans in the Northern hemisphere. The effect of this trend is to strengthen the inter-hemispheric gradient in the boreal winter which is conversely reduced in the boreal summer (Janicot and Fontaine, 1997). There is a corresponding significant impact on the seasonal northerly migration of the northern monsoon shearline, which has caused a slow reversal of inter-hemisphere SST anomalies in the world's oceans since the 1970's. As a result, all of Africa is often subject to the same fluctuations in precipitation anomalies, despite some modifications from regional anomalies (Mahe, 1993; Mahe and Olivry, 1995; Bigot, 1997).

The recent climatic variability observed during the last 30 years lead us to focus on more recent studies that have dealt with a trend detection in climatic variables.

3.8 Recent fluctuations in climate : the trend towards increasing temperatures and decreasing rainfall over West Africa

Since the end of the 19th century, that is, the end of the Little Ice Age, the earth's climate has been affected by a large scale warming trend that has resulted in an average global air temperature increase of 0.5°C (Janicot, 1995). However, this warming trend has been evolving at different rates in the two hemispheres over the last few decades, with the Northern hemisphere warming more slowly than the Southern hemisphere (*Fig.* 11). Consequently, the air temperatures as well as the surface water temperatures are lower in the Northern Hemisphere.

This temperature change differential between the two hemispheres could be compared using the bias of recorded cumulative rainfall in intertropical regions of the Northern hemisphere since the 1960s (*Fig.* 12). A dry period is especially pronounced over the last thirty years in Guinean and central Africa as well as in the Sahelian region, where dependence on rain-fed agriculture by the population makes it all the more dramatic (Janicot and Fontaine, 1997; Bigot, 1997; Brou Yao *et al.*, 1998). A temporal drying phase of a tropical climate also has a major impact on forest regions where rainfall is a critical factor for the monitoring and growth of vegetation (Janicot, 1995), which is further increased by the intensification of human activities causing land degradation. Even if it seems clear that such anthropogenic disturbances have certainly aggravated this global climatic trend; it is currently very difficult to separate the relative contributions of natural variations and human activities to long term climatic variability.

The variations in thermal structure of the ocean's surface explain to a large degree the variability of tropical climate, especially regarding precipitation : over 41% of rainfall variation in central Africa due to deep water convection, according to Bigot (1997); between 25 and 40% variation in rainfall in Guinea (Janicot and Fontaine, 1997); 50% of yearly variability in Sahelian rainfall (Fontaine and Janicot, 1993). Nevertheless, more than half of the rainfall in these tropical regions is not explained by SST variations. The increase of anthropogenic activities over the last 50 years could contribute significantly to the current climatic trend.

Indeed, the increase of carbon dioxide, relative to the increase of industrial activities is often evoked as a cause of global warming (Houghton, 1994; Le Houerou, 1993; Duglas, 1993). But in tropical forest regions that cover a lot of territory, such as in central and western Africa and the tropical Amazon, recent studies have been more focused on the impact of forest conversion on climate fluctuation (see Costa, this volume). In these regions sensitive to surface conditions where the atmospheric humidity has land mass origins (Fontaine, 1991), an increase in dry lands would provoke warmer air temperatures. This is firstly, because there is a greater warming of air by heat transfer and secondly, because the reduction of forest cover, which would naturally absorb carbon dioxide in the atmosphere, will contribute to the increase in greenhouse gases in the atmosphere (Janicot, 1995; Myers, 1991; Myers and Goreau, 1991; Keller *et al.*, 1991).

Recent work by Brou Yao *et al.* (1998) on rain forests in Ivory Coast underscores the similar evolution of : rainfall and forest surface area; a marked albedo increase after 1970 and massive forest convers; in turn, a decrease of forest cover has reduced the land based component of the water cycle, and therefore diminised the amount of water that gets recycled in the cycle's atmospheric phase. All of these factors seem to have contributed to the reduced rainfall of the 1970s. Nevertheless, there is still significant uncertainty as to the relative contributions of anthropogenic effects as against natural variations in climate in the context of the current trend of global warming and decreased rainfall (Janicot, 1995).

4 CONCLUSIONS

Recent findings in paleoclimatology, coupled with reconstructions of the past interactions between the climate and the vegetation cover in low latitudes (on the scale of the geologic era), has stimulated recent research evaluating the connections between surface processes and the regional climate. Schemes of these surface processes are now being integrated within General Circulation Models (GCMs). Most of the results of these simulations support paleoclimatic evidence of the major role of the vegetation cover on GCMs simulations of climatic variability. There remains however, difficulties in realistic representation of the biosphere dynamics of the climatic modelling, due to the inadequate representation of the summer tropical atmospheric circulation in GCMs, and to their too large resolution. There remain also an uncertainty about biogeochemical and biogeophysical processes interactions, which seem to oppose each other at the global scale, and recent studies indicate that the atmospheric feedback on climate maybe more important at a global scale than the biogeophysical feedback.

There are some clear examples of strong hydrological impact of forest clearance on runoff. For example the diminution of rainfall and runoff since 1970 concerns the whole West and Central Africa. The changes in land-cover seem to have a great impact on the hydrological cycle and the rainfall-runoff relationships. The impact of the forest clearance on the rainfall/runoff relationships seems to be dependent on the type of climate/vegetation system. In Sahelo-Sudanian areas, the forest clearance, associated with an increase of agricultural activities, rapidly induces a destructuration of the top layer of the soil which infiltrability decreases, and runoff coefficients increase. But in more humid tropical and equatorial areas such a correspondence is not observed. In equatorial humid areas the major impact of the forest conversion is to reduce the local evapotranspiration, thus reducing the total amount of available water vapour through local recycling for monsoon rainfall. Due to lack of direct measurements, it is very difficult to estimate the impacts of a massive forest conversion on the climate dynamics as well as on evapotranspiration.

Numerous uncertainties still remain about the knowledge of vegetation-climate relationships and about the part of the anthropogenic forcing on the current climatic evolution at the global scale. Nevertheless the human impact on local surface conditions is obvious at least at the regional scale of forest conversion, where hydrological changes are clearly connected to an increase of agricultural activities.

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Figure captions

Figure 1. Tropical forests and associated climates from the Lower Tertiary era to the beginning of the Quaternary era.

Figure 2 : Changes in South American and Atlantic Central African tropical forests over the past 20 000 years (Vincens *et al.*, 1996).

Figure 3 : Average and overall atmospheric energy (Polcher, 1994).

Figure 4 : Charney's Theory (1975) on retroaction of vegetation in dry and sub dry regions (Polcher, 1994; Fontaine and Janicot, 1993).

Figure 5 : Main interactions between water cycle and ocean-atmosphere-forest interface (Bigot, 1997).

Figure 6 : Retroaction of soil humidity on Charney's mechanism presented by Mylne and Rowntree (1992), Polcher (1994).

Figure 7 : Regional rainfall interannual variability over the period 1951-1989. Humid Africa is taken south of 8°N, including Western coast of the Gulf of Guinea and Zaire basin. Dry Africa is north of 8°N from Senegal to Chad.

Figure 8 : Dedadal variability of the monthly flows of the Kouilou basin (Rep. Congo Braz.).

Figure 9 : Decadal variations of precipitation (P) and runoff (Q) over eight regional areas in West and Central Africa : deviation from the 1951-1989 average in %. The rightmost column gives the cumulative deviation over the last two dry decades 1971-1989.

Figure 10 : Evolution of correlation between ENSO index and Sahelian rainfall during the period 1955-1984 (Janicot *et al.*, 1996.)

Figure 11 : Changes in air temperature differentials since the middle of the 19^{th} century, (Janicot, 1995).

Figure 12 : Change in cumulative rainfall in intertropical regions of the northern hemisphere, Janicot (1995).

Mahé G., Servat E., Maley J. (2004). Climatic variability including ENSO events (droughts, extreme events). In : "***Forests - Water – People in the Humid Tropics : Past, Present and Future Hydrological Research for Integrated Land and Water Management.***" ***Cambridge University Press***, 267-286.

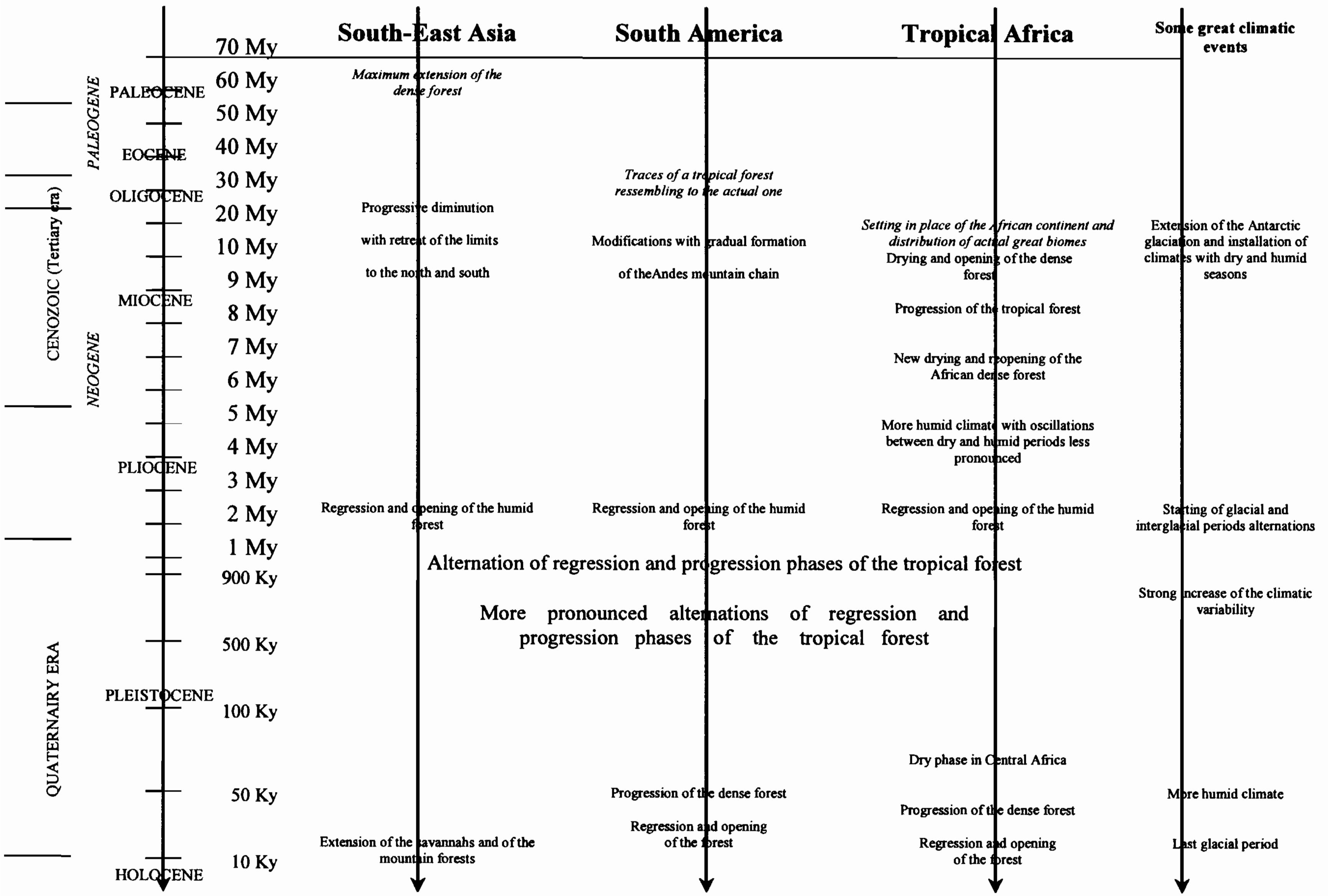
South-East Asia
South America
Tropical Africa
Some great climatic events
70 My
60 My
50 My
40 My
30 My
20 My
10 My
9 My
8 My
7 My
6 My
5 My
4 My
3 My
2 My
1 My
900 Ky
500 Ky
100 Ky
50 Ky
10 Ky
CENOZOIC (Tertiary era)
QUATERNAIRY ERA
PALEOGENE
NEOGENE
PALEOCENE
EOCENE
OLIGOCENE
MIOCENE
PLIOCENE
PLEISTOCENE
HOLOCENE
Maximum extension of the dense forest
Progressive diminution with retreat of the limits to the north and south
Regression and opening of the humid forest
Extension of the savannahs and of the mountain forests
Traces of a tropical forest ressembling to the actual one
Modifications with gradual formation of theAndes mountain chain
Regression and opening of the humid forest
Progression of the dense forest
Regression and opening of the forest
Setting in place of the African continent and distribution of actual great biomes
Drying and opening of the dense forest
Progression of the tropical forest
New drying and reopening of the African dense forest
More humid climate with oscillations between dry and humid periods less pronounced
Regression and opening of the humid forest
Dry phase in Central Africa
Progression of the dense forest
Regression and opening of the forest
Extension of the Antarctic glaciation and installation of climates with dry and humid seasons
Starting of glacial and interglacial periods alternations
Strong increase of the climatic variability
More humid climate
Last glacial period
Alternation of regression and progression phases of the tropical forest
More pronounced alternations of regression and progression phases of the tropical forest

Figure 1

Mahé G., Servat E., Maley J. (2004). Climatic variability including ENSO events (droughts, extreme events). In : *"Forests - Water – People in the Humid Tropics : Past, Present and Future Hydrological Research for Integrated Land and Water Management." Cambridge University Press*, 267-286.

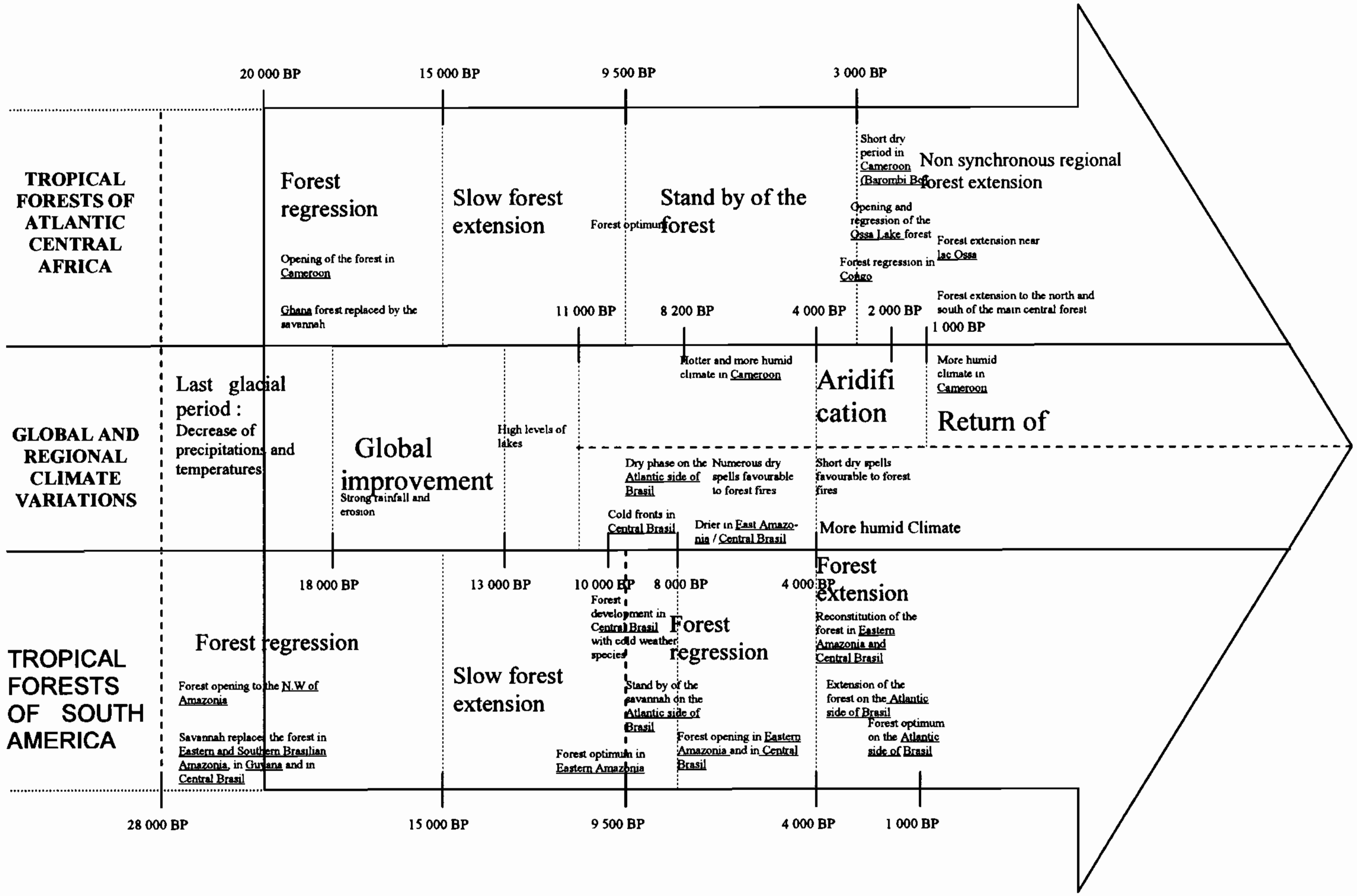

Figure 2

Mahé G., Servat E., Maley J. (2004). Climatic variability including ENSO events (droughts, extreme events). In : "*Forests - Water – People in the Humid Tropics : Past, Present and Future Hydrological Research for Integrated Land and Water Management.*" *Cambridge University Press*, 267-286.

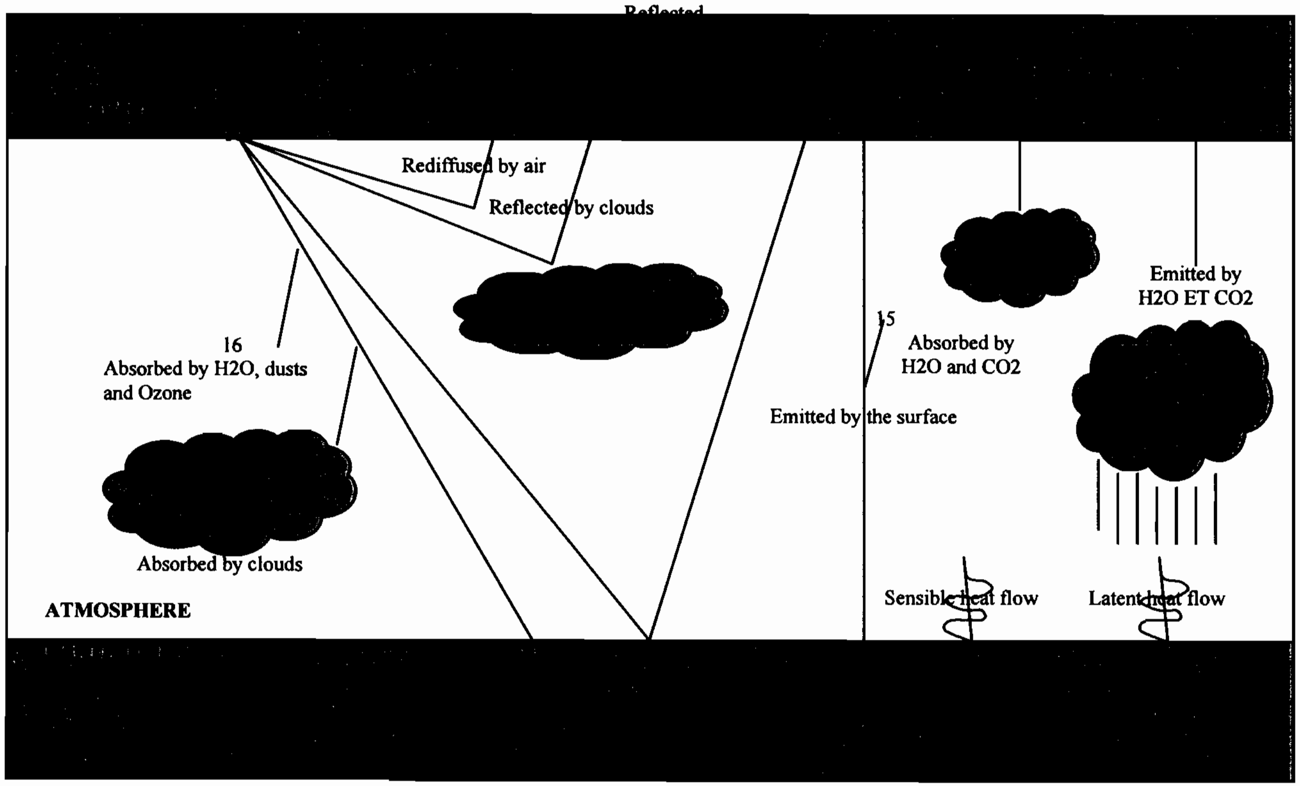

Figure 3

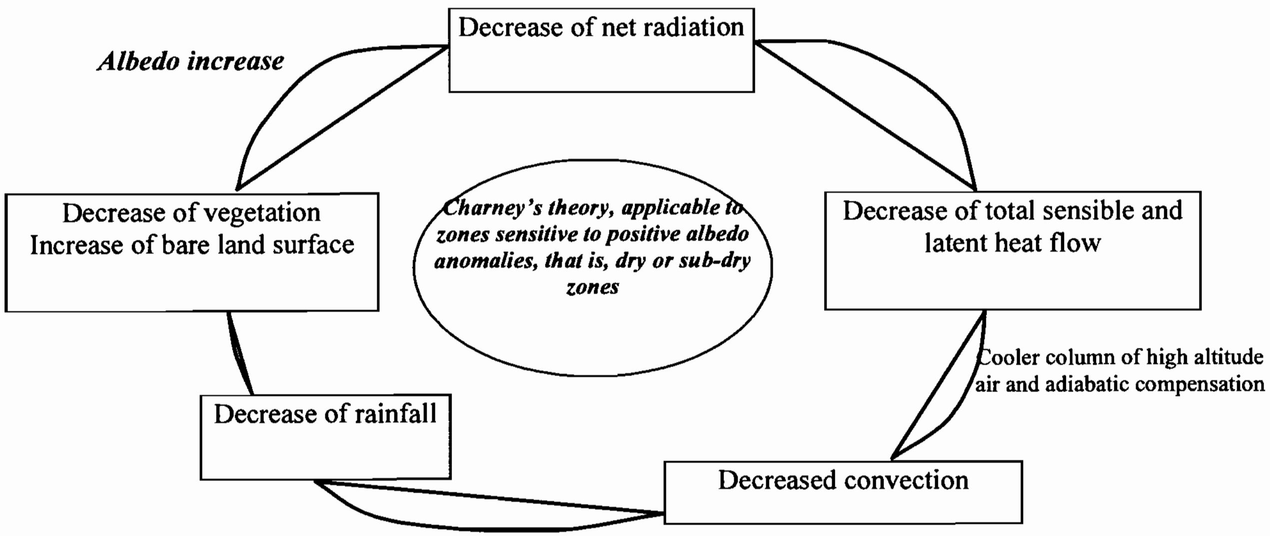
Decrease of net radiation
Albedo increase
Decrease of vegetation
Increase of bare land surface
Charney's theory, applicable to zones sensitive to positive albedo anomalies, that is, dry or sub-dry zones
Decrease of total sensible and latent heat flow
Cooler column of high altitude air and adiabatic compensation
Decrease of rainfall
Decreased convection

Figure 4

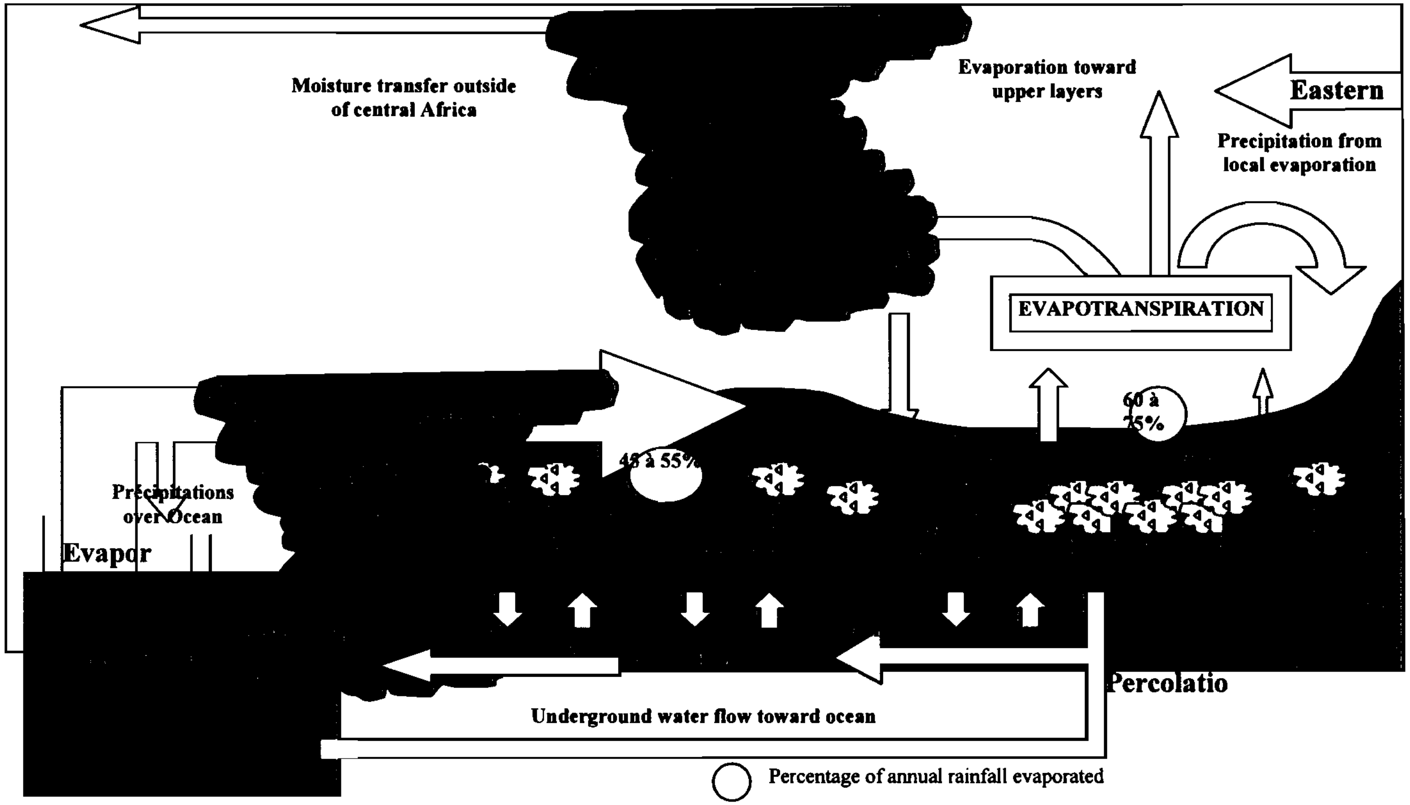
Moisture transfer outside of central Africa
Evaporation toward upper layers
Eastern
Precipitation from local evaporation
EVAPOTRANSPIRATION
60 à 75%
45 à 55%
Precipitations over Ocean
Evapor
Percolatio
Underground water flow toward ocean
Percentage of annual rainfall evaporated

Figure 5

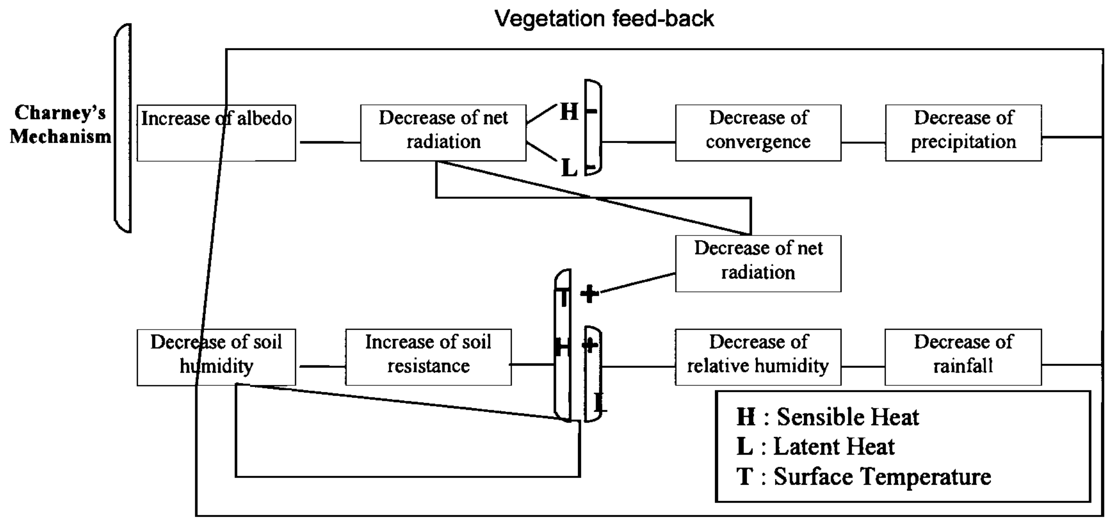
Vegetation feed-back
Charney's Mechanism
Increase of albedo
Decrease of net radiation
H
L
Decrease of convergence
Decrease of precipitation
Decrease of net radiation
T
H
Decrease of soil humidity
Increase of soil resistance
Decrease of relative humidity
Decrease of rainfall
H : Sensible Heat
L : Latent Heat
T : Surface Temperature

Figure 6

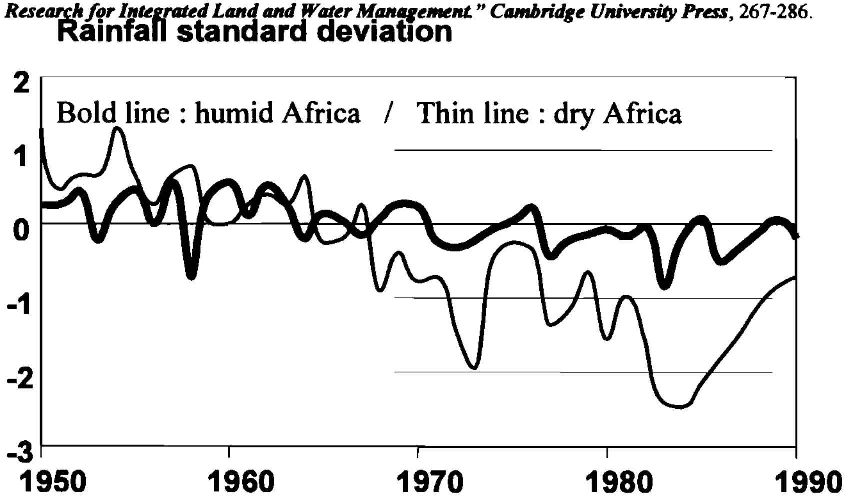

Figure 7

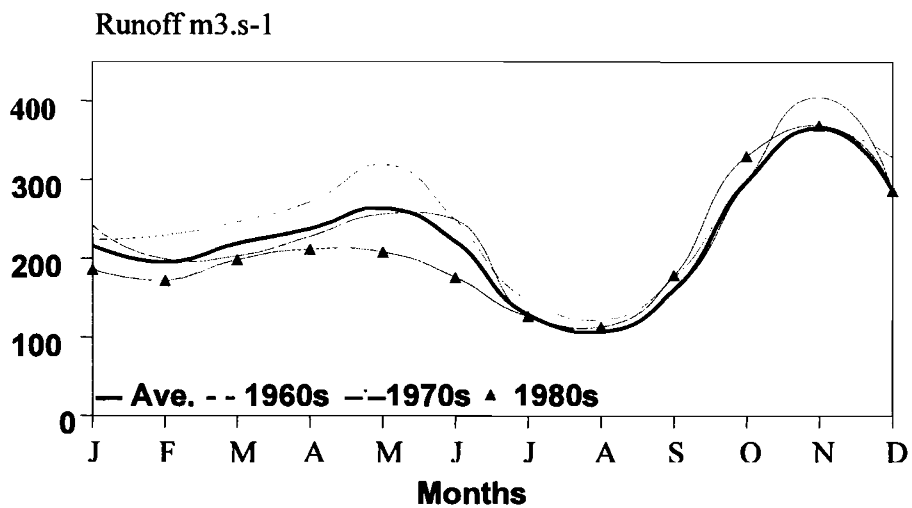

Figure 8

		1951-1960	1961-1970	1971-1980	1981-1989	Cumul 71-89
Sénégal/Gambia Rivers North Guinea : Riv. Corubal, Konkoure	P	+23,0	+13,0	-8,5	-16,5	-25,0
	Q	+32,6	+23,6	-24,1	-35,7	-59,8
South Guinea, Sierra Leone and Liberia Rivers	P	+10,3	+5,2	-3,5	-13,3	-26,8
	Q	+19,6	+15,7	-9,3	-28,8	-38,1
Niger River	P	+11,3	+3,1	-4,2	-11,2	-15,4
	Q	+14,8	+13,4	-8,7	-21,5	-30,2
North Coast of Gulf of Guinea : Ivory Coast, Ghana, Togo, Benin	P	+9,3	+4,6	-5,5	-9,4	-14,9
	Q	+23,4	+21,8	-18,4	-29,9	-48,3
Coastal Rivers of Nigeria, Central Cameroon : Mungo, Wouri, Sanaga	P	+3,1	+7,4	-1,4	-9,6	-11,0
	Q	+10,5	+12,6	-9,3	-15,3	-24,6
Angola Country, incl. Cubango & Cunene Rivers, and except Zaire River Basin	P	+2,6	+8,3	-5,2	-6,1	-11,3
	Q	+1,2	+8,7	-6,9	-4,0	-10,9
South Cameroon : Nyong & Ntem River Gabon/Congo Koui-Lou/Ogooue/Nyanga	P	+1,7	+3,6	-3,2	-1,4	-4,6
	Q	-1,2	+11,5	-6,9	-3,9	-10,8
Zaire/Congo River	P	+1,3	+3,2	-2,9	-0,6	-3,5
	Q	-4,0	+14,7	-1,8	-9,9	-11,7

Figure 9

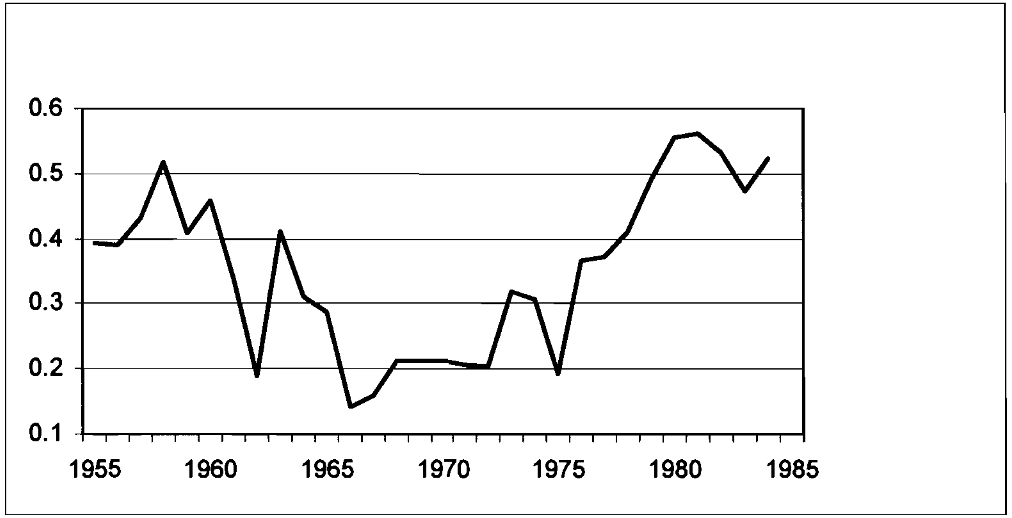

Figure 10

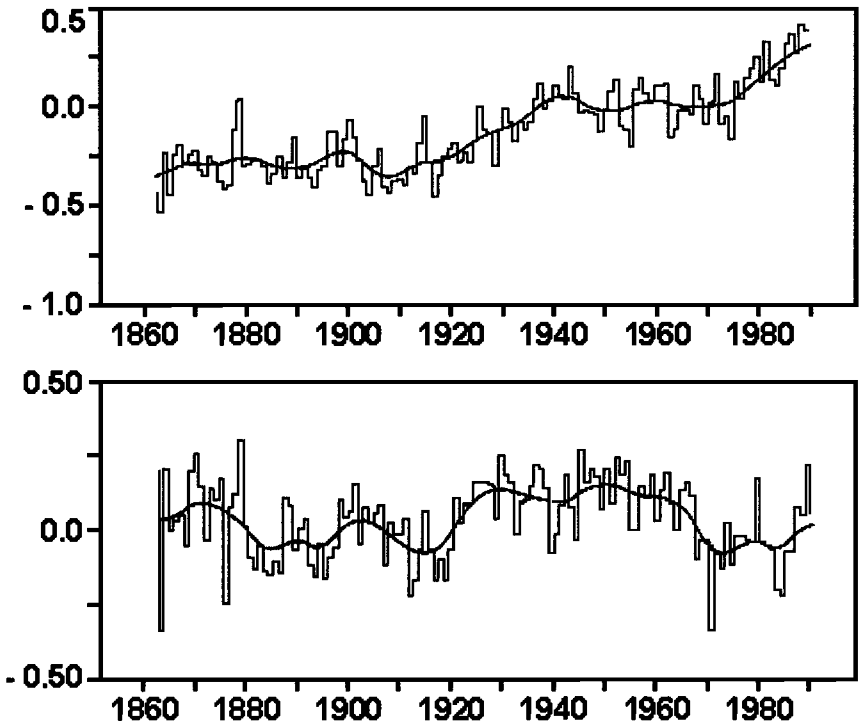
0.5
0.0
- 0.5
- 1.0
1860 1880 1900 1920 1940 1960 1980
0.50
0.0
- 0.50
1860 1880 1900 1920 1940 1960 1980

Figure 11

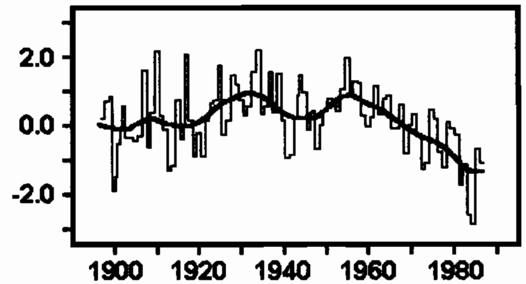

Figure 12