CENTRE FOR QUATERNARY STRATIGRAPHY AND PALEOCLIMATOLOGY

Contactgroups National Science Foundation (N.F.W.O. - F.N.R.S.)

CONTRIBUTIONS OF THE JOINT-MEETING HELD ON 18 - 19 - 20 APRIL 1983

Université Catholique de Louvain,
Institut d'Astronomie et de Géophysique

edited by

R. PAEPE and C. BAETEMAN

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GEOLOGICAL DATA IN PALEOCLIMATICAL PERSPECTIVES

by Roland PAEPE,

President CQS

An Introduction

Except for a few specialists dealing with the topic of stratigraphy, the majority of scientists having to judge about the extrapolation of geological data for paleoclimatical purposes, remain few.
Any new branch of development in a given field of investigation looks quite hasardous indeed especially in its very initial phases.

Wegener's Continental Drift needed more than 150 years to be proven correct by quite other ways of research than had been thought of originally.

True enough, marine or better deep-sea geological evidence clearly showed the double evidence of continuity in widening of the deep-sea-ocean floor at the same time as there was a continuous sedimentation, i.e. the paleoclimatical record.

Indeed thanks to the presence of appropriate marine fossil elements, deep-sea research allows determination of oceanwater temperatures along with an approximately elaborated timescale.

Despite the accuracy of measurements yielded in the marine geological field of investigation, number of problems still remain with regard to continental ones.

Indeed, it has been shown quite often that e.g. glacial movements do not necessarily occur simultaneously all over the world and even not in a given area. Moreover glaciated regions are no longer thought as to be the sole representative areas for setting up continental records.

The fossil periglacial regions now occurring mainly between 40° and 60° N yield a good number of sequences of alternating cold and warm cycles Via the periglacial area one finally links up with the subtropical and tropical areas.

There is a great need of developing appropriate methodologies in each of these regions concerned with special reference to each of the climato-sedimentological provinces and not at least with the landscape evolution. In the past quaternarists essentially were biased by research methods developed in the fossil periglacial and subpolar (actually periglaciated) regions. They exported methodologies from these regions towards the areas neighbouring the equator. Evidently this has conducted to wrong conclusions regionally as well as correlations on a global scale.

In the light of accurate measurements yielded from the deep-sea cores and the related mathematical treatment of these data in view of establishing paleoclimatic curves, there is a need of a more accurate investigation of the continental records all over the world.

This is the new task for geologists dealing with continental records in an attempt to meet with the requirements of mathematical processing of the geodata in a paleoclimatic perspective.

In creating CQS * as a contactgroup of NFWO-FNRS (National Fund for Scientific Research), it was aimed at a better development of an accurate standard methodology in the field of quaternary stratigraphic records. In creating Paleoclimatology * as a contact group, it was aimed at the better development of contacts between field geologists and climatologists.

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* : CQS : Belgian Geological Survey, President : R. Paepe
Secretaries : C. Baeteman, J. Thorez

* : Paleoclimatology : Institut d'Astronomie et de Géophysique, U.C.L.
President : A. Berger
Secretaries : R. Paepe, E. Van Overloop, P. Pestiaux.

PROGRAM

MONDAY 18 APRIL 1983 : CQS - MEETING

R. PAEPE	- Introduction : geological data in paleoclimatic perspectives.
P. HAESAERTS	- Climatic interpretation of the Upper Pleistocene Loess Record in Europe.
R. LANGOHR	- Soil studies as information source of paleo-climatic data, examples from Belgium.
K. BRUNNACKER & R. GRÜN	- Paleoclimatic indications given by speleothems, spring deposited travertines and marine terraces.
Ilunga LUTUMBA & R. PAEPE	- Sedimentary facies of the Ruzizi Plain deposits and their possible paleoclimatic implications.
R. BONNEFILLE	- Tropical pollenanalysis and paleoclimatology.
J. SOMME	- La séquence d'Aachenheim.
J. THOREZ	- Tentative paleoclimatic reconstitution through clay-mineral analysis.

TUESDAY 19 APRIL 1983 : PALEOCLIMATOLOGY I

A. BERGER	- Astronomical theory of Paleoclimates.
E. VAN OVERLOOP & R. PAEPE	- Climatic variations and geological data.
P. PESTIAUX	- PACDATA and Signal Processing.
J. NIHOUL	- Interactive ocean atmosphere models.
G. BRASSEUR	- Chemical and radiative models coupled to atmosphere.
G. SERET	- Paleoclimatic data of sedimentological analysis of the Grande Pile peat bog (Voges, France).
P. PESTIAUX	- The physical explanation of the typical spectral shape of the deep sea paleoclimatic records.
G. VAN KERSCHAVER	- Geomorphologie paleoclimatique dans le bas Zaire (Afrique Centrale).

WEDNESDAY 20 APRIL 1983 : PALEOCLIMATOLOGY II

Ph. GASPAR	- Modelling the upper ocean for climatic studies.
C. NICOLIS	- Coupled oscillator model in climate dynamics : synchronization, phase lag, and non periodic behaviour.
C. TRICOT	- Radiative approximations in climatological studies.
A. HENDERSON-SELLERS	- Modelling anthropogenic impact on climate.
J. PALUTIKOF	- The impact of CO_2-induced climate change on crop yields in England and Wales : methodology and results.
R. SNEYERS	- On the use of large values for the determination of short climatic variations.
C. TILL	- Etude dendrochronologique au Maroc.
Chr. GOOSSENS	- Principal component analysis of Mediterranean rainfall.
J.L. MELICE	- Desertification research in South of Tunisia.

SOIL STUDIES AS AN INFORMATION SOURCE OF PALEOCLIMATIC DATA. EXAMPLES OF BELGIUM.

by R. LANGOHR, J. SANDERS and T. LEROY

A brief review is made of the potential of various soil caracteristics and pedogenetic processes as an information source on paleoclimatic data.
For this interpretation the importance of the time factor as well as a good knowledge of 9 other factors of soil formation is stressed.
Two examples from Belgium are given, one about surface soils in the loess belt and one of buried soils in Tertiary Sediments.

PALEOCLIMATIC INDICATIONS GIVEN BY SPELEOTHEMS, SPRING DEPOSITED TRAVERTINES AND MARINE TERRACES

by R. GRÜN, K. BRUNNACKER

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The growth of secondary calcite formations (speleothems, spring deposited travertines) are strongly influenced by climatic factors. CO_2 and organic acids produced by humic soils on limestones increase the dissolution of $CaCO_3$ with seepage waters,

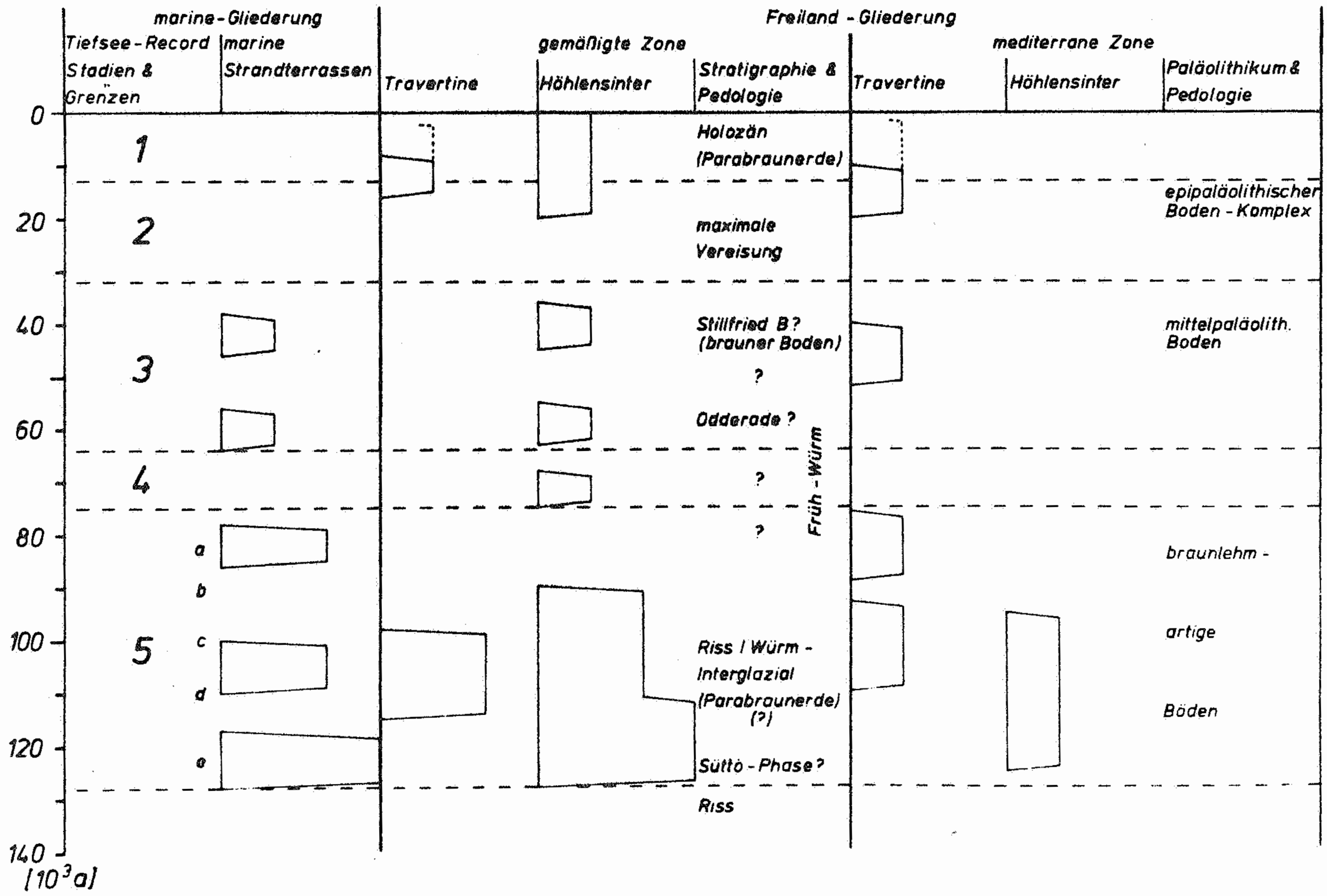

Fig. 1 : Land-Sea-Correlation on the base of U-series analysis.

which will hence reach a higher level in $Ca(HCO_3)_2$ when seeping through the limestone rock. This higher Ca-concentration of the karst-waters will lead to an increased formation of speleothems as well as spring deposited travertines.

On a Quaternary time scale it is expected, that secondary calcites were formed during warm and humic climates (interglacials - interstadials), whereas the growth ceased during permafrost or glacial climates because of the lacking of seepage waters.

A high number of about 7oo speleothems and about 15o spring deposited travertine age data given by U-series analysis were collected. These data were plotted as histograms and also as error-weighted frequency curves. Periods of most frequent speleothem growth turn out to be during approx. 9o,ooo - 13o,ooo yr and during the Holocene since about 15,ooo yr (fig.1). Periods beyond 15o,ooo yr cannot be yet recognized, because of the lack of sufficient results and the associated uncertainties of these in this age region.

A statistical evaluation of about 7oo age data correlated with marine terraces show a maximum at about 125,ooo yr.

A comparison of the frequency curves to the deep-sea core V28-238 oxigine isotope record (Shackleton & Opdyke 1973) shows a clear relationship for the secondary calcites as well as the marine terraces, especially for the 5e-stage which is classified as the last high sea level. Minor maxima of the frequency curves can be correlated to warmer climatic phases during the last glaciation (Würm).

LES FACIES SEDIMENTAIRES DES DEPOTS DE LA PLAINE DE LA RUZIZI ET LEURS POSSIBLES IMPLICATIONS PALEOCLIMATIQUES

par L. ILUNGA* et R. PAEPE**

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La plaine de la Ruzizi se situe au nord du lac Tanganyika dans le Western Rift entre les latitudes de 2°42' et 3°24' sud et les longitudes de 29° et 29°25' est.

L'étude de ces dépôts grâce à la photointerprétation, aux campagnes de terrain et aux analyses sédimentologiques nous a conduits à distinguer plusieurs faciès reflétant un environnement sédimentaire de type lacustre marginal.

En effet nous y avons observé :

-un faciès lacustre de séquence négative contenant des diatomées d'eaux douces à sa base;
-des faciès deltaïques de séquences négatives;
-des faciès de plaine deltaïque caractérisés par des cycles de séquences positives semblables aux dépôts de rivières de haute sinuosité;
-des faciès de rivières à méandre (point bar deposits);
-des faciès de rivières anastomosées divagantes (pebbly braided river deposits);
-des faciès de cônes alluviaux;
-des faciès des dépôts rouges d'épandage (rill wash,glacis)
-et des faciès de cuirasses ferrugineuses.

L'arrangement de ces différents dépôts sur le plan chronologique a abouti à l'établissement d'une lithostratigraphie dans laquelle nous avons distingué plusieurs Formations qui sont, des plus anciennes aux plus jeunes:

-la Formation de Muhira (± Pléistocène Inférieur);
-la Formation de Bu umbura (± Pléistocène Moyen);
-la Formation de Cibitoke (± Pléistocène Moyen);

-la Formation de Bwegera (± Pléistocène Moyen);
-la Formation de Gihungwe (± Pléistocène Supérieur);
-la Formation de Kamanyola correspondant au déversement du lac Kivu dans le lac Tanganyika daté de 9.500 ans BP (Hecky and Degens in Hecky, 1978) et donc holocène ;
-la Formation de Kadjeke ;
-et enfin les plages lacustres astuels, les cônes de déjection et les dépôts actuels des rivières, tous holocènes.

L'interprétation climatique faite sur base de la sédimentologie et de la géochimie (p^H et salinité) semblent refléter des oscillations climatiques allant du plus sec au plus humide ou inversement.

Le Pléistocène inférieur paraît humide (pollen, Sah 1969) mais devient moins humide vers la fin. Le Pléistocène moyen est dans l'ensemble sec mais devient relativement humide vers la fin. Le pléistocène Supérieur va du plus humide au plus sec, tandis que l'Holocène commence par une phase humide et présente un minimum sec.

Sur le plan tectonique, une phase active a dû exister vers la fin du Pliocène début Pléistocène inférieur puisque affectant les formations basaltiques datés d'environ 5,7 à $7,8.10^6$ ans BP (Pasteels, Communication orale).

Une deuxième phase tectonique importante a affecté la Formation de Muhira supposée du Pléistocène inférieur, y provoquant ainsi de petits horsts et graberes.

Une troisième phase a affecté la Formation de Cibitoke considérée comme datant du Pléistocène moyen et a par la même occasion affaissé la région zaïroise qui fut par la suite et pas nécessairement immédiatement après, envahie par les eaux du lac.

Une quatrième phase se serait produite à la fin du Pléistocène Supérieur et a consisté en l'affaissement de la Basse Plaine zaïroise d'une trentaine de mètres par rapport à la Burundaise.

Et enfin, une cinquième phase tectonique aurait affecté la Formation de Bujumbura au cours de l'Holocène.

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PALAEOCLIMATIC RECONSTRUCTION THROUGH CLAY MINERALS : A PROBLEM OF PARAMETERIZATION

by J. THOREZ

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Clay minerals seem to offer an attractive tool for deciphering palaeoclimatic variations and evolution within marine and continental post-Palaeozoïc (Tertiary and Quaternary) series, in particular when these are poor or devoid of any kind of fossils (pollen, Foraminifera). However the method - clay mineral analysis mostly achieved by X-Ray diffraction of usual complex clay assemblages - is not easy to applied. There appear many restrictions to it, and results must be handled carefully.

Surficial argillization of the Earth crust is a naturel, but obligatory geochemical pathway for any silicate-bearing geological material (igneous, metamorphic or sedimentary rocks; sediments) or soil that crops out at the surface through erosion and uplift, and becomes progressively submitted to a step-by-step weathering. The latter concerns a thermodynamic and physico-chemical readjustment that develops at the limit of the lithosphere and hydrosphere, that leads to the birth of clay mineral assemblages. The process of argillization is mainly a hydrolytic one governed by percolating (surficial drainage) and/or leaching (internal drainage) desaturated solutions escaping the landscape by gravity and moving uphill-downhill within the substratum. Beside the activity of the waters, other factors intervene in the process: These are to be considered as external climatic "stimuli". Temperature, when high (as in tropical areas) can speed up to height times the weathering of the silicates on the way of argillogenesis, comparatively to the temperature characterizing temperature areas. Rainfalls are also very important as they condition the action of the percolating solutions; but their action is mainly a matter of regularity through the whole year beside the volume reaching the surface.

Other physical parameters must be also taken into account: the nature of the substratum (mineral composition, grain-size, textures); importance of the tectonic which refreshes the Earth crust and offers to the weathering agents new silicate-bearing material unstable with the hydrosphere conditions; relief which governs the uphill-downhill escape of solutions; altitude with its consequent moisture variations; biological influence (vegetation cover, bioturbation); rate of weathering cupled with rate of accumulation (lack or weak influence of erosion). And duration of the process that develops within an open system.

Stratigraphy- litho and chronostratigraphy- is as well an important parameter because offering series of deposits which may be interrupted (erosive effects or lack of deposition, as in continental series) or somewhat "recycled" materials (mixing of several weathering products) developed in several areas on the continent, but that become mixed "downhill" when reaching their final settlement in the sedimentary basin.

The process of hydrolysis (degradation, desilication) with its subsequent processes (neoformation, aggradation) is neither simple nor direct. The geochemical pathways leading to the argillization of the substratum are generally highly complexed. They depend on the nature of the parent material wherein some minerals are more sensitive to weathering than the other associated ones. In other words the minerals are variable in their sensitivity towards the climatic, thermocynamic and physico-chemical "stimuli" quoted above. The step-by-step often interrupted degradation, therefore, naturally gives rise to polymineralic clay assemblages which associate for a certain duration: residual parent minerals, transitional products (particularly mixed layers, smectites, vermiculte, secondary chlorite, all of them produced by the degradation of parent illite, biotite, chlorite), and possibly some end-products (kaolinite, amorphous). Neoformed minerals can mix up the degradation sequence (i.e. neoformation of smectite), producing mimetism effects.

As a consequence, when a tentative approach about the palaeoclimatic environment and conditions is seeked through a clay mineral analysis of geological or pedological series, two necessary conditions must be encountered at the very outset of the data treatment.
First, the parent material and minerals must be traced back in time and space as to know their quality and eventually their residual occurrence in the weathered clay assemblages. Secondly, an accurate and complete analysis- qualitative and semi-quantitative - must be completed as to define all the associated minerals: parent, transitional and end-products or minerals. The degradation sequence of the polymineralic assemblages may, then, reconstructed even if intermediate terms are lacking actually because their labile of fugitive appearance. The nature of the end-products (smectite, vermicultite,

chlorite, kaolinite) can, afterwards, be indicative of the weathering trends in term of geochemical processes: respectively smectitization, vermiculitization, secondary chloritization. These are grouped within specific diagrams wherein the occurring and missing transitional phases (the mixed layers) clearly appear. Even if the step-by-step degradation embodies polymineralic clay assemblages, it is easy to reconstruct the different sequences: parent - transitional - end-minerals, and to number each step. From the occurrence of the minerals, and because their numbering within the sequence(s), a graphical device can be offered wherein the trend, rate, importance of the weathering are matched. These weathering numbers, coupled with the relative abundance each time of the occurring minerals, lead to a "weathering or aggression" index representing the importance of the hydrolysis that has affected the parent material following one or more geochemical pathways quoted above: smectitization, vermiculitization, chloritization, ...

The second step in the climatic reconstruction is to build up, in parallel with the lithostratigraphic sequence, three specific curves: one is redated directly to the "weathering index" while the two others refer to the inferred climatic (cold-temperate-warm) and pluviosity or moisture (dry-dry-wet-wet) conditions. Indeed the occurrence and relative abudance of the various clay minerals, in particular the mixed layers, smectite, vermiculite, secondary chlorite and kaolinite can be related to such climatic parameters.

These curves represent a tentative reconstruction of the successive climatic variations and evolutions that have accompagnied the building up the geological and/or pedological series. They are, of course, not in continuity as the actual supperposed layers or beds may have been the only ones which escaped erosion.

Hiatuses as well as the importance and quality of the "missing" deposits, and the fact that non-deposition may occur in reality, all these factors demonstrate that the pattern shown by the three curves is, in reality a "shaked" evolution. Such a situation is particularly true for continental series whilst for the marine ones, even if considered to have been deposited in "continuity", the problem of mixing or "cannibalism" of "uphill" materials complicate the pattern of the climatic curves.

There are other situations. For instance if there occurs a climatic shift from a mild climate (i.e. dry-temperate) to a drastic one (warm-wet), the climatic signature will only be shown by the latter conditions, "shading" the former ones. If a warm-temperate climate (inferred from the clay anamysis) shifts to a drier climate, no change will occur in the clay assemblage, generally (in reality other "signatures" may occur such as a caliche encrustment). Finally for a climatic signature to become printed in the clay assemblage, a certain span of geological time must have to be recorded within the deposit: a too short shift dry-temperate to wet-warm will not be recorded.

Examples of application are illustrated for the Quaternary series of Italy and Greece, offering possibilities to demonstrate the different interferences of the climatic parameters associated with the nature of the clay assemblages.

ASTRONOMICAL THEORY OF PALEOCLIMATES : INSOLATION SIGNATURES FOR THE LAST GLACIAL CYCLE AND ACCURACY OF LONG-TERM VARIATIONS OVER THE QUATERNARY

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Among all the indices against which the geological data are tested to validate the astronomical theory of paleoclimates, the most popular ones are certainly the orbital elements of the earth and linear or non-linear combinations of them.

However, as the climat system is thermally driven by solar insolation, there is a real interest to check whether or not there are some relationship between insolation parameters and the climate at the global scale. The difficulty wich then first arises is to determine which are the most realistec latitudes and during which periods of the year. The adjustment may be tested in the time and/or the frequency domains. This is why the different kinds of insolation which are supposed to be used for modelling the climate or for simulating the climatic variations, are carefully reviewed (computation, accuracy and spectrum).

These insolation parameters are : half year astronomical seasons (lenght, total and mean insolation), half year caloric seasons, astronomical and meteorological seasons (length, total and mean insolation), monthly mean insolation mid-month (solar date) and calender date insolations.

These insolations depend upon (I) 4 astronomical parameters which all affect the total energy received by the planet : the astrophysical solar constant, the lenght of the tropical year and of the day, the secularly varia mean distance from the Earth to the Sun (defined by the eccentricity, e),

(II) and 2 other ones which re-distribute differently the energy among the latitudes and months : the climatic precession, e sin $\tilde{\omega}$, and the obliquity, ε This dependance is summarized in table 1.

Table 1 - Insolations as a function of astronomical parameters (++ means stronger dependancy).

	ε	e sin $\tilde{\omega}$
Mid-month insolation at equinoxe		+
at solstice	+	++
Half-year astronomical seasons		
-total insolation	+	
-lenght		+
-mean in polar latitudes	++	+
in equatorial latitudes		+
Caloric seasons polar latitudes	+	
equatorial latitudes		+
Meteorological seasons (astronomical definition)		
-total insolation	+	
-lenght		+
Meteorological seasons (monthly mean)	+	++

The accuracy of the long-term variations of the astronomical elements and of the insolation values and the stability of their spectrum have also been analysed by comparing 7 different astronomical solutions and 4 different time spans (0-0.8 Myr BP, 0.8-1.6 Myr BP, 1.6-2.4 Myr BP and 2.4-3.0 Myr BP). The general conclusions are, for the accuracy in time, that improvements are necessary for periods further back than 1.5 Myr BP. About the stability of the frequencies, the fundamental periods (around 40, 23 and 10 Kyr) do not detoriate with time over the last 5 Myr but their relative importance for each insolation and even astronomical parameter is a function of the period considered.

The second part of this paper deals with the time behaviour of the insolation patterns over the last 150 000 years.

This analysis of the deviations of midmonth daily insolation from the mean state calculated over the last million years clearly indicates that variations in the annual cycle of the insolation and in the amplitude of the summer month insolation especially in high Northern latitudes are related to Quaternary climatic changes. It is not a linear real-time relationship, but it is their degree of steadiness in time which is thought to be responsible for the appearance or not of a full glacial or interglacial. This persistence is required because of the lag times for oceanic and cryospheric responses to atmospheric temperature changes, these lag times being proposed to be a function of the amplitude and of the speed of the insolation change.

Starting 129 000 YBP-5-6 stages boundary-where the positive deviation is the largest in June and amounts to 100 cal $cm^{-2}d^{-1}$, a strong warming of the summer months is observed, lasts steadily till 122 000 YBP at which a maximum deviation of 50 cal $cm^{-2}d^{-1}$ is observed in August and thus corresponds clearly to Barbados III coral reef terrace or substage *5e*
In 118 000 YBP, a deep minimym is already present in June (- 88 cal $cm^{-2}d^{-1}$) and is still present in August 109 000 YBP (- 31 cal $cm^{-2}d^{-1}$), announcing the low sea-level stand found in Barbados data and the low temperature found in Camp Century ice core.
The next maximum positive deviation arising in June (70 cal cm^{-2} d^{-1}) is located 105 000 YBP (Barbados II) and has changed in a weak minimum (- 48 cal $cm^{-2}d^{-1}$) starting 95 000 YBP, year for wich ice-rifted detrita have been found in deep-sea core V23-82.

From 94 000 YBP, the weak maximum insolation in April intensifies and turns towards a maximum in July-August in 82 000 YBP (Barbados I). During the same time, a minimum is created in winter and shifts towards a large minimum in April-May at around 75 000 YBP, the maximum in August-September being now very weak. The minimum is then entering the summer season(-72 cal $cm^{-2}d^{-1}$ in June 73 000 YBP). Remaining there during some 12 000 years, it can be regarded as responsible for the beginning of the last generally cold interval (Würm).

The period 65 to 32 000 YBP is characterized by the total absence of large deviations in the midmonth insolation, allowing the climate to stay in its cold state, some stadials and interstadials being only superimposed on it. From 70 000 YBP, insolation in May-July starts growing up. The maximum shifts from May in 62 000 YBP to July in 58 000 YBP, thus defining the 59 000 YBP Port Talbot I interstadial. From 60 to 48 000 YBP (Port Talbot II interstadial) a slight maximum is persisting during summer months, fall and winter receiving a little bit less insolation than the mean. From around 48 to 36 000 YBP, the maximum is moving back to spring, another maximum of the same small amplitude being also present in winter. Then it starts returning to summer months and at 32 000 YBP (Plume Point interstadial), only spring has negative departures. From 27 000 YBP, a minimum is deepening rapidely in April and by 22 000 YBP it has reached May and June.

In fact, from 25 000 YBP (-55 cal $cm^{-2}d^{-1}$) till 18 000 YBP (-25 cal $cm^{-2}d^{-1}$), midmonth negative deviations are persistently largest in June-July, this feature being related to the maximum glacial advance of the Würm.

However, this summer minimum becomes to be relatively weak at 19 000 YBP, year in which a maximum in May is appearing. Then insolation in April-May-June amplifies and, in 11 000 YBP (beginning of the Holocene interglacial), reaches a maximum which left the summer season 3000 years ago (+ 10 cal $cm^{-2}d^{-1}$ in August). Since that time, the spring minimum is deepening regularly and is shifting towards June which insolation deviation amounts now to - 30 cal $cm^{-2}d^{-1}$.

This minimum is progressively moving towards August and weakens (- 8 cal $cm^{-2}d^{-1}$ at 8000 YAP), no well-shaped maximum being present in the annual cycle during all this period. From 9000 YAP, when midmonth deviations from the mean are almost zero, a new minimum appears in March-April, shifts slowly towards summer and deepens (- 23 cal $cm^{-2}d^{-1}$ at 17 000 YAP). This minimum disappears in August 22 000 YAP, a year in which, for the first time after present, a spring maximum develops, moves to summer months and reaches +18 cal $cm^{-2}d^{-1}$ in June-July 28 000 YAP. It is not before 50 000 YAP that another significant deviation will appear. Located in May, it shifts towards August by 60 000 YAP, at the same time as when an insolation signature is being developed (June minimum of -40 cal $cm^{-2}d^{-1}$ in 51 000 YAP, June maximum of + 38 cal $cm^{-2}d^{-1}$ in 64 000 YAP).

Thus the very flat pattern of the annual cycle observed at 60°N between 70 000 YBP and 30 000 YBP must indicate the maintenance of the triggered glaciation around 75 000 YBP. The deep minimum starting in April around 30 000 YBP and moving towards June and July between 25 000 YBP and 20 000 YBP must be responsiblefor the last glacial maximum that has occurred around 20 to 18 000 YBP. When the speed of schange of the annual-cycle pattern from one thousand years to the next is fast, going from a large maximum summer insolation to a deep minimum and back to a maximum, in a very limited time span, short cooling or "abortive" glaciation has to be expected. It is the case from 127 000 YBP (maximum) to 118 000 YBP (minimum) and to 107 000 YBP (maximum again).

As far as the present Holocene interglacial is concerned, the 11 000 YBP July maximum is clearly related to its beginning. It is also evident that the insolation during summer months is decreasing since 300 YBP, will reach its minimum in July-August around 3000 YAP and will not be significantly larger than the mean state before 24 000 YAP.

CLIMATIC CHANGES FROM EPI-PLEISTOCENE TILL RECENT IN THE EASTERN MEDITERRANEAN, AFRICA AND NORTHERN SOUTH AMERICA : A REVIEW

by E. VAN OVERLOOP

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EEC - European Economic Community, Project Climatology, 05/002, CLI/024/B

Abstract

Vita-Finzi's theory concerning the "Older Fill" and "Younger Fill" fluviatile deposits covering the whole eastern Mediterranean is subject neither to many detailed neither to precise comparable climatic subdivisions.
The "Older Fill" starts at about 40.000 BP. Its end would coincide with the transition from constant climatic conditions, suggesting drought with short rainfall to major climatic fluctuations which change the aspect of the Fill deposits. However this end is situated at several moments of the Holocene (after 10.000 BP), varying from the Atlantic to the Subboreal and even the Subatlantic, depending on the author.
The "Younger Fill" at the contrary would start at about 1.600 BP, on one hand ending ± 200 years BP, on the other hand still going on nowadays.

The studies of the Holocene in Greece, E. Attica, reveal a rhytmic recurrence of dry climatic conditions with an average interval of 1.000 years.
As from the transition from the Subboreal to the Subatlantic (2.700 BP) on, several dry phases seem to have occurred : the Geometric drought of about 2.700 BP (Agora), the Roman drought of about 1.800 BP, the 700 BP drought (Akominatos) and the present-day one.
Twelve dry-wet cycles represented by as many soil formations proceed from these studies. Furthermore, the Atlantic (7.500 BP - 5.000 BP), up till now undifferenciated, now seems to yield in E. Attica four climatic subdivisions, bringing the number of cycles in Greece upon sixteen.

When comparing the results from E. Attica with other data from the eastern Mediterranean and especially with data from Africa and South America, the good correspondence between the Greek curve and these from Africa and South America appears (Table 2).

This fact suggests the existence in some deposits in the whole eastern Mediterranean of a stratigraphically very detailed subdivision, of the Holocene. In E. Attica, Greece, such subdivision is reached by means of the combination of various disciplines, such as stratigraphy, archaeology, archaeogeology, clay mineralogy, palynology, sedimentology and palaeomagnetism applied on the same sections. This fact is argument to encourage a profound and interdisciplinary approach of climatologic problems everywhere.

I. Introduction.

Many multidisciplinary data exist about changing environmental conditions and palaeoclimates over the world. In this respect the Eastern Mediterranean, Central and East Africa and Northern South America are some of the most interesting and intensively studied regions. A study of palynological and some geomorphological data made the comparison of climatic curves from these areas possible, covering the Late Glacial and the Holocene i.e. the Epi-Pleistocene.

II. The eastern mediterranean and N. Africa.

II.1. The Late Glacial.

Just before the start of the Late Glacial, the end of the Upper Pleniglacial (< 13.000 BP) has generally been marked by an increase of moisture and/or rainfall.

The area around the Dead Sea points at an increase in rainfall until 12.000 BP (*Bar-Youssef, 1974*). The same occurs in Lebanon - Syria until 13.000 BP (*Leroi - Gourhan, 1973*) after which dry conditions become dominating.

Palynology in Northern Israël (*Van Zeist & Bottema, 1980*) points at a recovery of the forest as from 14.000 BP. Hereafter a short dry phase (12.500 - 12.000 BP) seems to have interrupted the moistening in the Levant (*Leroi - Gourhan, 1980 ; Goldberg, 1980 ; Henry, 1980*).
Optimal climatologic conditions prevail between 12.000 and 11.000 BP as shown by pollen analysis (*Van Zeist & Bottema, 1980*). This period coincides with the flourishing of the Natufian culture in the Levant and Syria.

The Nile area and Egypt give evidence of decreasing floods and local winter rainfalls, which last from 17.000 BP until 8.500 BP. A dry interstade occurs at about 11.500 BP (*Butzer, 1975*).
For other regions, except for Greece, no precise datable information exists except for some broad statements about the Late Glacial as a whole.

The Sahara and North Africa also show a generally moister and even warmer climate, however, with more local rainfall and locally dryer areas (*Rognon & Williams, 1977 ; Jäkel, 1977 ; Rognon, 1979*).
Humidity is furthermore better expressed in the Central Sahara than in its northern part (*Pias et al., 1975*).

In Iran, the existing Artemisia-steppe slightly changes into a beginning Quercus- Pistacia woodland, which points again at increasing moisture (*Van Zeist & Bottema, 1978 ; Van Zeist & Woldring, 1978*). Geomorphology in Iran shows brief and very violent rainfall for this period. The relatively low temperatures are responsible for the increasing moisture.

The situation in Greece was almost the same ; herbs slowly left their places for trees (Quercus-Pistacia), and here too an increase of moisture was due to low evapotranspiration (*Van Zeist & Bottema, 1977 ; Bottema, 1979 ; Wijmstra, 1960*). Turkey shows a long lasting drought due to higher temperatures (*Van Zeist & Woldring, 1978 ; Van Zeist et al., 1975*).
The Post-Glacial therefore may generally be identified by a slight amelioration of the climatic conditions due to an increase of humidity.

II.2. Preboreal and Boreal (10.500 - 7.500 BP).

After the drier period in the Levant (11.000 - 9.000 BP), pollen analysis, archaeology, and geomorphology indicate again wetter environment during the 10th millennium BP.
The climate is even moister than at present (*Bar-Youssef, 1975 ; Goldberg, 1980 ; Tchernov, 1980 ; Aurenches, 1980*). From 9.000 BP on, one meets with an increasing drought with a climax of aridity around 8.500 BP.

The same is attested in Syria and Lebanon, where palynology on settlements (*Leroi - Gourhan, 1978 ; Van Zeist & Bottema, 1980 ; Van Zeist, 1977 ; De Contenson, 1980*) shows a steady drop in arboreal pollen percentage.

In Iran and Turkey (*Van Zeist & Woldring, 1978 ; Van Zeist et al., 1975*), the inner area gradually is invaded by a forest - steppe during this time-span.

On the contrary Greece already is covered by a dense oak forest (*Van Zeist & Bottema, 1977 ; Bottema, 1974, 1979*) i.e. the climax vegetation, under increasing temperatures and dry environmental conditions, which turn to an increase in precipitation towards the end of the Boreal (8.000 BP).
At this very moment too, *R. PAEPE* and *M. HATZIOTIS* locate the Marathon Soil, coinciding with the beginning of the Neolithicum, and as determined by *J. THOREZ* on basis of clay mineralogy testifying of warmer and more humid conditions as well.

As a general conclusion the eastern part of the Mediterranean has been favoured by a moistening of the climate during the Preboreal, with a steady drying up throughout the Boreal followed by a humidity-peak around 8.000 BP. (*See also Paepe, Hatziotis, Thorez, Van Overloop, 1982*) (Table 1).

In the Central Sahara, increased rainfall lasts until 9.000 - 8.000 BP interrupted by dry phases (*Jäkel, 1977*).
For the Northern Sahara, a wet phase is known to occur between 8.500 - 7.000 BP (*Rognon & Williams, 1977*).
In Egypt, the local rainfall remains up till 8.500 BP (*Butzer, 1975, 1978*).

The southern part of the Mediterranean testifies of regional variance of rainfall-dispersion, but the general tendency as described for the eastern part of the Mediterranean still exists.

II.3. Atlanticum (7.500 - 5.500 BP).

The Atlantic is known as being a period of intensive human settlement in all regions.

In the Levant, settlement during the first part of the Atlantic remains in isolated areas whereas the most intensive period of dispersion of sites occurs between 6.000 and 5.500 BP as a result of greater agriculture possibilities due to higher humidity (*Moore, 1973 ; Price-Williams, 1973*).
Also pollen analysis in this region points at warmer and moister climatic conditions during the Atlantic (*Horowitz, 1974*). Furthermore the Dead Sea clearly shows a higher level at the end of the Atlantic (5.500 - 4.300 BP), due to more humid conditions.

Northern Israël climate between 7.500 and 5.500 BP is moister and comparable to the today's one. The pollen evolution is comparable to the Western European one (*Horowitz, 1975*), i.e. the installation of the climax vegetation.

Syria and Lebanon also deals with climatic conditions during the Atlantic period similar of the today's ones (*Leroi - Gourhan, 1973*). A highly moist phase at 5.000 BP is shown, whereas geomorphologic evidence points at higher rainfall between 7.500 - 5.800 BP (*Besançon, 1980*).

In Arabia, investigation concerning the Atlantic reveals pluvial periods until about 6.000 BP (*Mc Clure, 1976*).

North African climate in the Atlantic period first crosses an aridity phase, followed by a period of higher moisture between 6.500/6.000 and 5.000 BP (*Rognon & Williams, 1977 ; Jäkel, 1977*).
Lakes develop in the western Sahara from 8.000 till 4.000 BP, due to optimal climatic conditions as well (*Petit - Maire, 1977*).

Iranian palynological investigations yield the final establishment of the Quercus climax forest at the end of the Atlantic period, favoured by a warm and wet climate (*Van Zeist & Bottema, 1977*).

Evenso in Turkey, the climax vegetation comes to growth, also in mountaneous areas.

In Northern Greece, palynological studies (*Wijmstra, 1969 ; Bottema, 1974, 1979*) give prove of the final installation of the climax deciduous oak forest at the end of the Atlantic, as a result of an increase in temperature and humidity during this period. The archaeogeological evidences collected by *R. Paepe, M. Hatziotis, J. Thorez* and *E. Van Overloop*, in East-Attica (Greece), point at warm and intermittent dry climatic conditions during the Atlantic, a tendency which lasts until 3.900 BP, i.e. the end of the Protohelladic period coinciding with the middle of the Subboreal. For this period, four dry-wet cycles are up till now discovered (*M. Hatziotis, oral communication*).

The general trend during the Atlantic is in every part of the Mediterranean one of continuous moistening, with a humidity peak differing in time from area to area.
The intermittent dry-wet situation in East Attica (*R. Paepe, J. Thorez, M. Hatziotis, 1980*) makes us believe in more existing oscillations during the Atlanticum in all concerned regions, which are to be envisaged by deeper investigations on soils, pollen and archaeology on more detailed sections, if available.

II.4. The Subboreal (5.500 - 2.700 BP).

Results obtained from disciplines as varied as pollen analysis, geomorphology and archeology point at more humid and cool climate for the beginning of the Subboreal (5.500 - ± 4.000 BP) in the Levant (*Price-Williams, 1973 ; Oates, 1976 ; Vita-Finzi, 1978 ; Crown, 1972 ; Margaritz, 1973*).

Syrian palynology (*Bottema, 1977*) gives prove of a semi-arid climate for the Subboreal.

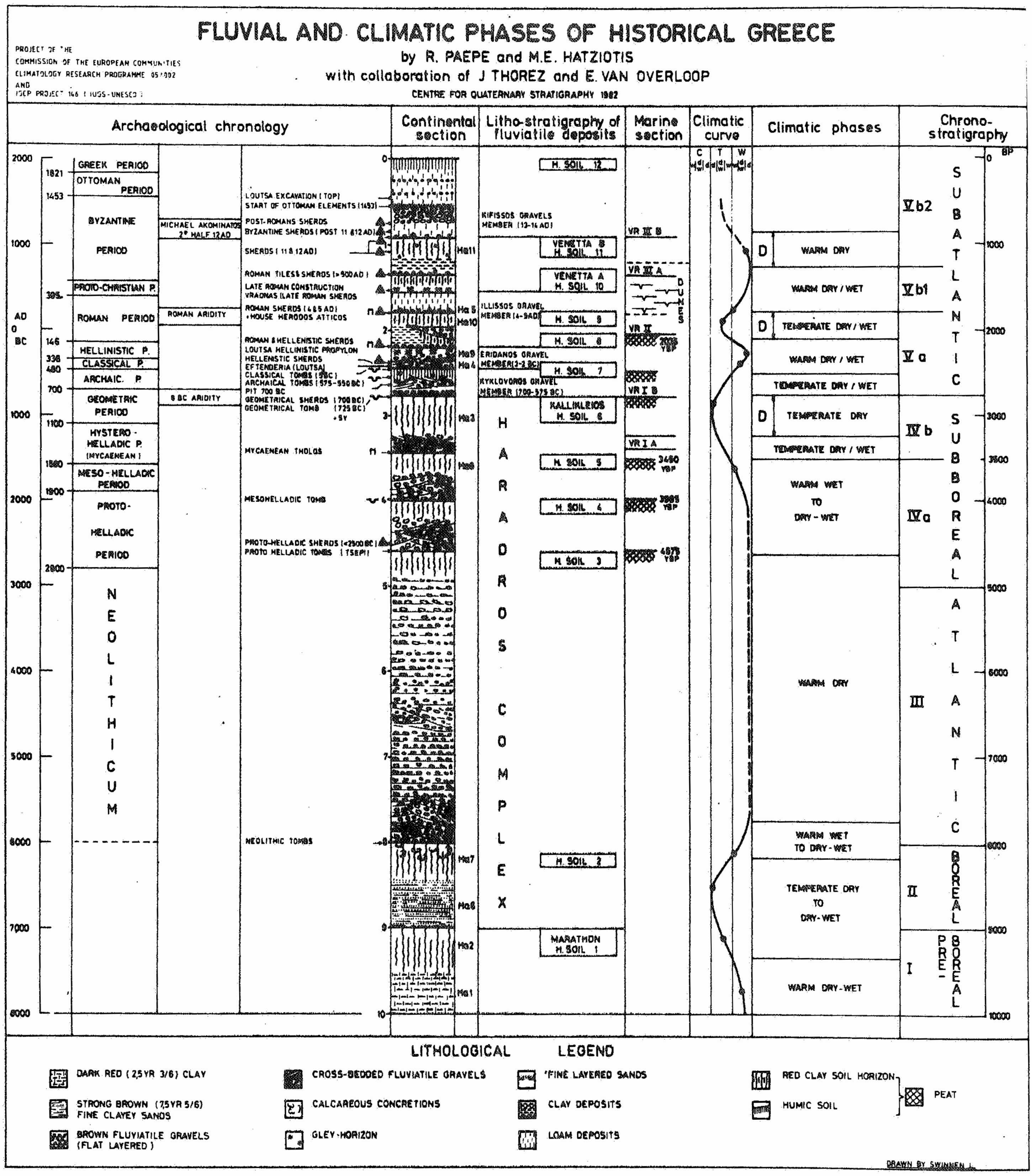
Pias et al. (*1973, 1974, 1975*) show the existence in the NW part of the Sahara of a cooler and more humid climate around the transition from Atlantic to Subboreal, changing into a warm and dry climate towards the end of the Subboreal.

Palynological investigation in Turkey (*Van Zeist & Woldring, 1978 ; Van Zeist & Bottema, 1977*) prove the definitive installation of the climax forest coinciding with a further moistening of the climate due to lower evapotranspiration as a result of lower temperature.
Also in NW Greece (*Bottema, 1974 ; 1979*) the already existing climax woodland remains almost unchanged, despite cooling and moistening of the climate. *Bintliff* (*1980*) concludes that climatic changes are only observed in the so-called "marginal" woodlands i.e. the ones growing in temporarily favoured areas where desertification is able to extend very rapidly due to local environmental factors.

Iraq deals for the same period with two big flood-periods (*Paepe & Baeteman, 1977*) respectively dated at 4.850 - 4.300 BP and 3.800 - 3.450 BP.
The last mentioned flood coincides with a humid period in Greece (*Paepe, Hatziotis, Thorez, Van Overloop, 1982*).

In eastern Attica (*Paepe, Hatzoitis, Thorez, Van Overloop, 1982*), these floods are also represented by the upper part of the Haradros Gravel complex, a series of continental gravels and/or peat formations in the coastal plain of Marathon. They correspond to three cycles of sedimentation each of which is coinciding with respectively the Proto-Helladic, Meso-Helladic and Hystero-Helladic (Mycaenian) Periods separated by a series of Holocene Soils : H.S.3, H.S.4, H.S.5.
A change from warm-wet to temperate dry climate at the end of the Subboreal is found, a drought which probably is at the origin of the fall of the Mycaenian culture (*Carpenter, 1966 ; Bryson, 1974 ; Laub, 1977*). Then dry conditions enhance the development of the "Kallikleios Soil", H.S.6 (*Paepe, Hatziotis, Thorez, Van Overloop, 1982*).

The Subboreal is thus clearly dominated by a change in climate which has marked almost every area of the Mediterranean in a similar way : i.e. cooling-off and the transition from wet to dry climate, especially in Greece, resulting in a soil development.


FLUVIAL AND CLIMATIC PHASES OF HISTORICAL GREECE
by R. PAEPE and M.E. HATZIOTIS
with collaboration of J THOREZ and E. VAN OVERLOOP
CENTRE FOR QUATERNARY STRATIGRAPHY 1982
Archaeological chronology
Continental section
Litho-stratigraphy of fluviatile deposits
Marine section
Climatic curve
Climatic phases
Chrono-stratigraphy
GREEK PERIOD
OTTOMAN PERIOD
BYZANTINE PERIOD
PROTO-CHRISTIAN P.
ROMAN PERIOD
HELLINISTIC P.
CLASSICAL P.
ARCHAIC P.
GEOMETRIC PERIOD
HYSTERO-HELLADIC P. (MYCAENEAN)
MESO-HELLADIC PERIOD
PROTO-HELLADIC PERIOD
NEOLITHICUM
MICHAEL AKOMINATOS 2nd HALF 12AD
ROMAN ARIDITY
8 BC ARIDITY
LOUTSA EXCAVATION (TOP)
START OF OTTOMAN ELEMENTS (1453)
POST-ROMANS SHERDS
BYZANTINE SHERDS (POST 11 & 12 AD)
SHERDS (11 & 12 AD)
ROMAN TILES & SHERDS (>500 AD)
LATE ROMAN CONSTRUCTION
VRAONAS (LATE ROMAN SHERDS
ROMAN SHERDS (4 & 5 AD)
+HOUSE HERODOS ATTICOS
ROMAN & HELLENISTIC SHERDS
LOUTSA HELLINISTIC PROPYLON
HELLENISTIC SHERDS
EFTENDERIA (LOUTSA)
CLASSICAL TOMBS (5 BC)
ARCHAICAL TOMBS (575-550 BC)
PIT 700 BC
GEOMETRICAL SHERDS (700 BC)
GEOMETRICAL TOMB (725 BC)
MYCAENEAN THOLOS
MESOHELLADIC TOMB
PROTO-HELLADIC SHERDS (<2500 BC)
PROTO HELLADIC TOMBS (TSEPI)
NEOLITHIC TOMBS
KIFISSOS GRAVELS MEMBER (13-14 AD)
ILLISSOS GRAVEL MEMBER (4-5 AD)
ERIDANOS GRAVEL MEMBER (3-2 BC)
KYKLOVOROS GRAVEL MEMBER (700-575 BC)
HARADROS COMPLEX
H. SOIL 12
VENETTA B H. SOIL 11
VENETTA A H. SOIL 10
H. SOIL 9
H. SOIL 8
H. SOIL 7
KALLIKLEIOS H. SOIL 6
H. SOIL 5
H. SOIL 4
H. SOIL 3
H. SOIL 2
MARATHON H. SOIL 1
VR III B
VR III A
DUNES
VR II
VR I B
VR I A
2005 YBP
3460 YBP
3985 YBP
4875 YBP
C T W
D WARM DRY
WARM DRY / WET
D TEMPERATE DRY / WET
WARM DRY / WET
TEMPERATE DRY / WET
D TEMPERATE DRY
TEMPERATE DRY / WET
WARM WET TO DRY-WET
WARM DRY
WARM WET TO DRY-WET
TEMPERATE DRY TO DRY-WET
WARM DRY-WET
Vb2
Vb1
Va
IVb
IVa
III
II
I
SUBATLANTIC
SUBBOREAL
ATLANTIC
BOREAL
PRE-BOREAL
LITHOLOGICAL LEGEND
DARK RED (2,5YR 3/6) CLAY
STRONG BROWN (7,5YR 5/6) FINE CLAYEY SANDS
BROWN FLUVIATILE GRAVELS (FLAT LAYERED)
CROSS-BEDDED FLUVIATILE GRAVELS
CALCAREOUS CONCRETIONS
GLEY-HORIZON
FINE LAYERED SANDS
CLAY DEPOSITS
LOAM DEPOSITS
RED CLAY SOIL HORIZON
HUMIC SOIL
PEAT
DRAWN BY SWINNEN L.

II.5. The Subatlantic (2.700 BP till Present).

During this period the climate in the Levant becomes almost the same as the present one, although in some cases less extreme.
The Sinai and surrounding area's show evidence for settlement from ± 2.800 BP until ± 2.000 BP (*Evenari, 1971*), probably as a result of increased humidity favouring agriculture which is responsible for the human occupation.

Moreover, stratigraphic evidence in the Levant witnesses of a "Younger Fill", which represents very thick fluvial deposits, dated by *Vita-Finzi* (*1969*) between 1.600 and 200 BP or by *Goldberg* (*1980*) between 1.700 and 600 BP.

The "Younger Fill" is also dealt with in Syria at the very same period as in the Levant (*Besançon, 1980*), and further on in Iran (*Vita-Finzi, 1969 b, 1975 a ; Brookes, 1977*) and the Gulf (*Diester - Haass, 1973*) where fluviatile "Younger Fill" deposits seem interrupted by dry intervals.

In Turkey, the "Younger Fill" is largely represented as well (*Vita-Finzi, 1969 ; Bintliff, 1975 - 1977 ; Dufaure, 1976*) dated however from ± 1.600 BP till Present.

As for the Subboreal, pollen-analytical diagrams show little if any change in their pattern during the Subatlantic. The climate changes are said to be too short to give general modifications in the wood cover, whereas from 3.000 BP on, human influence plays an important role in the vegetation assemblage.

The same vegetational historical problem exists for Greece (*Bottema, 1974, 1979 ; Wijmstra, 1969*) where too small time intervals of climatic changes and too, intensive human occupation occurred. Anyhow, *Bottema* (*1974*) mentiones cooler and relatively moist Subatlantic conditions in Greece.

It has been possible thanks to transdiciplinary research between archeology, stratigraphy, clay mineralogy and sedimentology to subdivide the Subatlantic in even more detailed climatologic events (*Paepe, Hatzoitis, Thorez, Van Overloop, 1982*) (Table 1).

Four main fluviatile gravel-beds are identified belonging to the last 2.700 years, each time coinciding with a change of human cultures ; "Kyklovoros Gravels" (700 - 575 BC), coinciding mainly with the Archaic Period and resting on the higher mentioned "Kallikleios Soil" (H.S.6) ; "Eridanos Gravels" (3 - 2 Century BC) comprising Classical and Hellenistic Periods overlying H.S.7 ; "Ilissos Gravels" (4 - 5 Century AD) overlying H.S.8 and 9 and finally "Kifissos Gravels" (13 - 14 AD) overlying Holocene Soils 10 and 11 i.e. "Venetta Soils". These evidences allow to subdivide the Subatlantic into five climatologic divisions (*Paepe, Hatziotis, Thorez, Van Overloop, 1982*) namely a temperate dry-wet phase, a warm dry-wet phase, followed by another temperate dry-wet phase of the end of the Roman Period ; a warm dry-wet one coinciding with the Proto-Christian and the beginning of the Bryzantine Period marked by the "Venetta Soil" development. The middle of the Byzantine Period is marked by a warm dry environment which is also mentioned by *Michael Akominatos* (2nd half 12 AD).
The sixth climatic subdivision shows a jump into the warm and wet conditions lasting during the Ottoman Period, characterised by sandy gravel deposits, until the Present-day conditions.

Thus, the so-called "Younger Fill" deposits comprise only the "Ilissos" and "Kifissos Gravels" of the detailed subdivision of *Paepe, Hatziotis, Thorez* and *Van Overloop (1982)*. They incorporate three desertification periods i.e. the Roman, the Byzantine and the Present-Day Periods.

III. Central Sahara - Tchad.

Diatom studies and palynological investigations on the lacustrine sediments of lake Tchad (*M. Servant & S. Servant, 1970, 1972 ; J. Maley, 1977*) provided the scientific world with some of the most detailed information on climatic fluctuations in the lake Tchad basin. After compilation of all the data of these works, the Late Glacial and the Holocene seem to be composed of not less than seventeen humidity-drought cycles, the oldest one occurring at ± 14.000 BP.

Between 11.000 BP and 10.000 BP, three cycles are noticed with a peak in humidity at the base of the Preboreal.

An 800 years drought persists around 9.700 BP whereafter five humidity peaks have been discovered up until the boundary between the Boreal and the Atlantic (7.500 BP).

The very Atlantic can be compared to the "climatic optimum" of the Western Sahara (*H. Petit-Maire, 1978*) since a very wet and warm climate coincides with this Period.

As from 5.000 BP on till recent, seven humid dry cycles occur.

Remarkable is the similarity between the course of the cycles for Greece and for the Tchad region, especially during the Subatlantic.
Although lake Tchad ranges under the so-called "marginal" climatic regions and one could therefore to some extent expect detailed investigation results, the similarity with Greece again proves that precise results can also be obtained in less "marginal" or sensible regions of the earth, by means of collaboration between several disciplines.

IV. Eastern Central and East Africa.

Eastern Central and East Africa yield detailed studies on palynology from high mountaineous lakes in the western part of the great Rift lakes and on diatoms from eastern Rift lakes. The palynological diagram from lake Mahoma, mount Ruwenzori, Uganda, elaborated by *Livingstone* (*1967*) and reinterpreted by *Hamilton* (*1972*), together with the general overlook on palynological data from several mountains in the same region (*Van Zinderen-Bakker & Coetzee, 1972*) have been selected in order to establish a temptative construction of a climatic curve considering the changes in humidity and temperature in eastern Central Africa. As for East Africa, the same procedure has been followed, using the data from diatom studies on the lakes of the Afar-region, Ethiopia (*Gasse, 1975, 1977, 1978 ; Gasse & Street, 1977*).

IV.1. Eastern Central Africa.

The curve on summarized data which are described above (*Livingstone, 1967 ; Hamilton, 1972 ; Van Zinderen-Bakker & Coetzee, 1972*) yields a very dry Late Glacial, colder than today. In comparison with the other discussed regions, the upwarming and mountaineous deglaciation in Eastern Central Africa takes place very early, namely as from ± 14.000 BP on. A phase of increased humidity is situated at about 11.500 BP. The today's climatic conditions are established quite directly after this upwarming.

A more humid phase is mentioned, lasting during the Preboreal and the Boreal, changing into a period of still more humidity which coincides with the end of the Atlantic and the beginning of the Subboreal, this last one accompanied by a lowering of the temperature.

The very "climatic optimum" is for eastern Central Africa situated in the second half of the Subboreal, with high humidity and temperature.

The Subatlantic so far has not been subdivided into detailed oscillations of drought and humidity. As for the pollendiagram of *Livingstone (1967)*, reinterpreted by *Hamilton (1972)*, a possible subdivision seems to proceed out of the curves of certain well specified taxa of humidity-related plants, of which the pollen sediment locally or over long distances (*Personal interpretation*). Remarkable is the fact that these curves do proceed or receed at nearly the same time, and that the curve of the taxon indicating "human influence" behaives quite independantly of the curves of the humidity indicators. Hence the human influence on these moist-indicating taxa seems to have been of less importance.
Four dryer periods seems to occur as from the end of the Subboreal (2.700 BP), interrupted by periods of increased humidity, each oscillation covering about 500 years.

It would be of very much interest to put the palynology of the Subatlantic on the Central African mountains into more detailed investigation, as the features of human disturbance are well known in botany and did in historical times not affect the whole natural, especially the mountaneous vegetation.

IV.2. Eastern Africa.

F. Gasse (*1975 ; 1977 ; 1978*) and *E. Street* (*1977*) worked out, by means of diatom studies, several detailed oscillation curves of lake levels in the high mountains of Ethiopia.

F. Gasse evenmore pointed out, especially for the Tardiglacial and the Holocene, which were the specific palaeoenvironments in which the diatoms were living. Compiling these data, a temptative climatic curve has been elaborated.

The deglaciation took place before 11.500 BP, since at this very date, the present environmental conditions were already established.

Four levels of high humidity are recognized during the Lower and Middle Holocene (up till 3.500 BP) ; the first one occurs at ± 10.000 BP. The second one covers the end of the Preboreal and the beginning of the Boreal (up till ± 8.200 BP). The third one starts with the Atlantic, after a dry and colder oscillation, and ends with a short dryer phase around 6.000 BP. The fourth one coincides with the base of the Subboreal, after which a longer drought lasts (from ± 4.500 BP - ± 3.500 BP) at the end of which a cooling-off is registered.

As from 3.500 BP on, *F. Gasse* describes an hypothetic cyclicity of 450 - 500 years, expressed by the alternation of wet alcaline and dry acid environmental lacustrine conditions (*Mount Badda, F. Gasse, 1978*), cyclicity which has been confirmed by the diatom and lake level studies on other lakes of lower altitude (*F. Gasse, 1977 ; F. Gasse & E. Street, 1977 ; Butzer et al., 1972 ; F. Gasse, 1975*), probably pointing at a cyclic variation in local rainfall pattern.
For the whole Holocene, an eleven-cycles curve of dry-wet situations is suggested, of which four occur in the Lower and Middle Holocene and eight in the Upper Holocene.

As for the results from Lake Tchad area, here again the similarity with the Greek curve is striking, especially when looking at the number of cycles and their timespan.

V. Northern South America.

Both the palynological investigations on the "Sabana de Bogota", situated on the high plateau's of Columbia (*T. Van Der Hammen, 1961, 1963, 1966, 1974*) and the Holocene palynologic investigations in the Amazone basin, Brasil (*M.L. ABSY, 1978*) yield sufficient evidence for the construction of a tentative curve for the Lower and Middle Holocene to which the curve for the Upper Holocene from the work of *M.L. Absy*, is added.

The Tardiglacial is again very dry in northern South America and the upwarming starts at about 13.000 BP, with a new cooling off between 11.000 BP and 10.000 BP.

A humid phase covers the end of the Preboreal and the beginning of the Boreal (9.000 BP). Moreover, this humid phase is also found back in Minas Gerais (Brazil), by means of palynological investigation on a lacustrine peat bog, indicating a transition from high mountaineous vegetation towards the installation of the tropical-atlantic rainforest. This transition coincides with a high lake-level phase around 9.000 BP. (*E. Van Overloop, 1981*).

At the boundary between the Boreal and the Atlantic, a short dryer phase is recorded, following a period of intermediate dry-wet climate.

High temperature and humidity caracterise the Atlantic.

Two dry oscillations occur at the base of the Subboreal (± 5.500 BP and ± 5.000 BP) accompanied by a lowering of the temperature.

As from 3.000 BP on, three main cycles of higher humidity are discovered, interrupted by dryer phases. The second cycle shows three peaks and three periods of somewhat dryer conditions.

The first dry phase occurs from 2.700 BP up until ± 2.000 BP, after having passed through a maximum drought around 2.100 BP. The next dry phases are situated at ± 1.500 BP, ± 1.200 BP and ± 1.000 BP, and are much less expressed.
Moreover, two dry periods occur at ± 700 BP and ± 400 BP (*M.L. Absy, 1978*) ; the last one coinciding with the Western European "Little Ice Age".

VI. General overlook and comparison.

(See Table 2)

VI.1. Comparison of the curves.

Great similarities exist between both the curves from Central Sahara and Greece. Both of them are as well comparable to the curve from East Africa. Even the number of cycles is comparable for these three regions.

On the other hand, the curves from northern South America and from eastern Central Africa show many coincidences, which are on themselves again comparable to the East African curve.

Since data from the Levant area and Egypt are, taking into concern the difficult field material (few fossils, complicated stratigraphy), not very abundant, the curve from these regions does not appear sufficiently detailed as to compare it to other curves. Unless this fact, some of its events may be found back in the Greek curve as well.

The curve which behaves quite independently is the one from the Atlas Mountains, say north-western Sahara, where climate has always been rather particular.

The Late Glacial throws light upon a general upwarming of the atmosphere in all the discussed regions. Only the Levant and Central Sahara show a more humid phase around 14.000 BP, whereas the humidity period between 11.000 BP and 12.000 BP is registered in the Levant, Central Sahara (at 11.000 BP) and eastern Central Africa.

Around 10.000 BP, during the Lower Preboreal, another humid phase has been mentioned, a two-fold one for Central Sahara and one for northwestern Sahara, Greece, the Levant and East Africa. In eastern Central Africa, a more humid phase starts at about 10.000 BP which will last until ± 7.000 BP.

At the end of the Preboreal, a humid phase occurs : a very short one in Central Sahara ; further in Greece, East Africa and northern South America. This phase lasts in the last two mentioned regions until the middle of the Boreal (± 8.500 BP).

Also in Greece, the Boreal is caracterised by one well expressed humid phase around ± 8.200 BP. At the end of the Boreal, eastern Africa suffers drought and cold. The same dry period occurs in Greece, and with some delay and very short in northern South America and Central Sahara. In this last region, this dry oscillation is the last one of a cycle of five dry periods, interrupted by humid ones, covering the end of the Preboreal, the Boreal and the beginning of the Atlantic.

The Lower Atlantic again announces humidity and high temperatures, especially pronounced in northern South America, East Africa and Central Sahara. In eastern Central Africa, humidity heighers but does not reach its maximum.

During the Atlantic, the northwestern Sahara withnesses of a dry climate, whereas Greece yields stratigraphic evidence of intermittent dry-wet conditions. (*Paepe et al., 1982*). Around the transition from Atlantic to Subboreal (5.500 BP), a dry oscillation is well marked for northern South America and for East Africa.
A second dry period occurs in the Subboreal, around 5.000 BP in northern South America and Central Sahara ; around 3.600 BP in East Africa ; at about 4.300 BP in the central Sahara and towards the end of the Subboreal (2.700 BP) in all discussed regions except the northwestern Sahara (no sufficient data), before which a humid oscillation is again noticed in northern South America and hypothetically in East Africa and eastern Central Africa.

Remarkable is the fact that the curves from northern South America, East Africa (hypothese), Central Sahara and Greece yield five periods of drought during the Subatlantic, each of which is accompanied by a cooling off in Central Sahara and Greece.

As from about 3.500 BP on, these cycles roughly repeat each 500 years. The today's drought, coinciding with extention of deserts, is going on everywhere except in the tropics, where the today's phase is a humid one. Desert extention is hitherto not only a human factor.

VI.2. Discussion.

The deglaciation during the Late Glacial took generally place in very early times (before 13.000 BP) in the tropical regions and appearingly much later in the subtropical regions (from 12.000 BP on).

Remarkable is the fact that the period of the so-called "climatic optimum", in some regions takes place during the Atlantic, in other regions (i.e. eastern Central Africa, and north-western Sahara) during the Subboreal. Whereas the minor climatic oscillations during the climatic optimum in the other regions are not found back, Greece yields evidence of a four-fold intermittent dry-wet climate for that period. Similar oscillations could possibly exist in other regions as well, which encourages more detailed study.

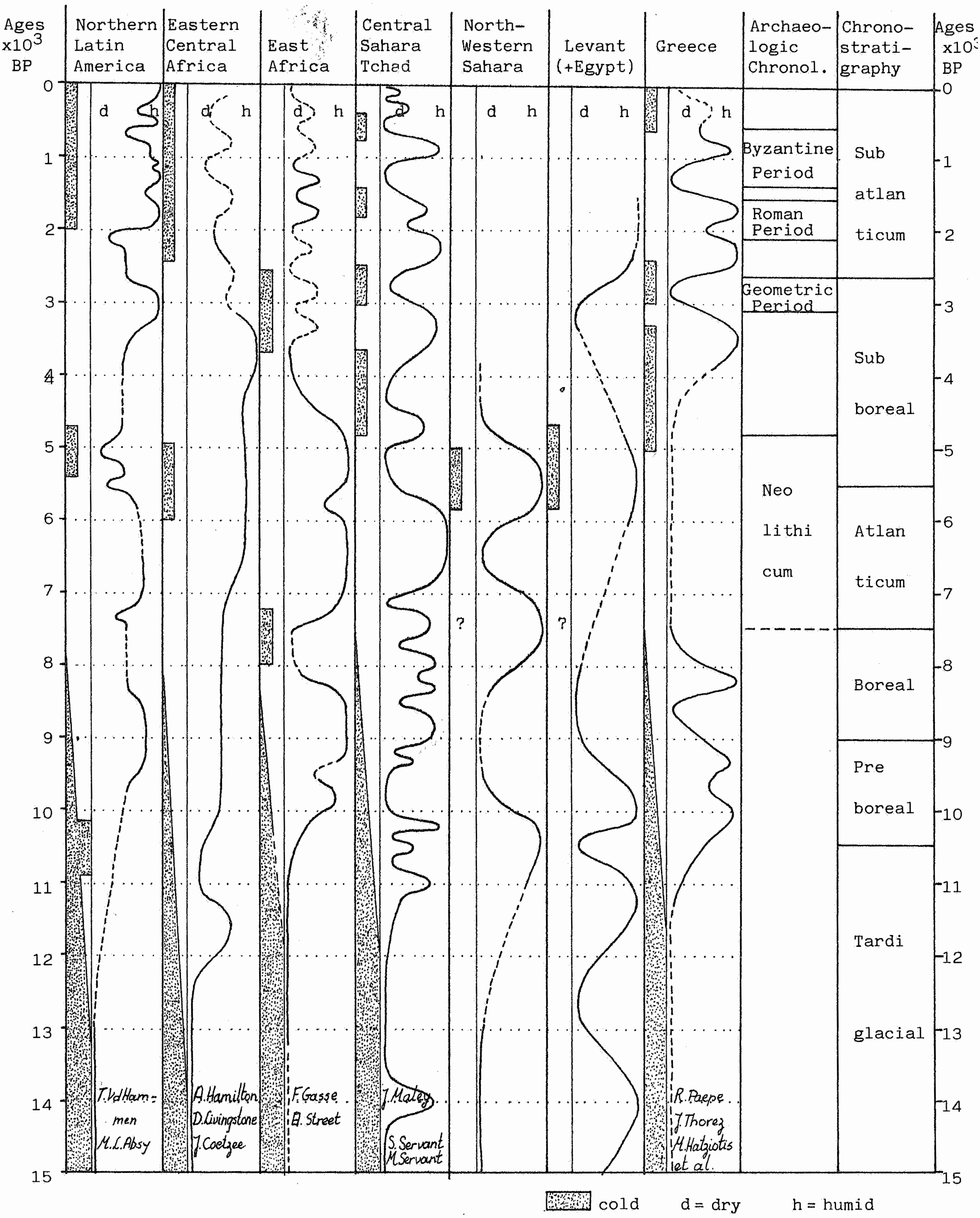
As for the subdivision of the Holocene, cycles from one region clearly do not always coincide with cycles from other regions.
The possibility exists to make long-distance correlations with the Western European Holocene chronostratigraphy (*Blytt - Sernander ; Livingstone & Van Der Hammen, 1978 ; Coetzee, 1973*), but attention should be made whether local chronostratigraphic subdivisions and names are not primarily preferable, taking into account the locally very different climatological features (*Ab Saber, 1978 ; Rognon, 1977, 1980*) which define the climatic variations in different regions.

Whereas the most detailed curves (Greece and Central Sahara) yield at least fifteen dry-wet cycles, the curves from northern Latin America and East

Africa at least procure eight of them (hypothetically eleven in East Africa). As there are for northwestern Sahara and the Levant not enough data available, and as the data from tropical Africa are subject to carefull interpretation because of the pioneer-character and the complexity of the investigations, the possibility can be suggested of the existance in the sediment records over the world of still more detailed data and of more cycles, encouraging very detailed and transdisciplinary investigation.

Many authors discuss about the fact that the Lower and Middle Holocene would cover relatively less oscillations than the Upper Holocene. Many times, only an "Older" and "Younger Fill" is mentioned, without any precise subdivision or climatic implication.
On the contrary, almost all the discussed curves show a major oscillation of 1.000 - 1.500 - 2.000 years, which generally becomes more detailed in the Subatlantic, i.e. with a 450 - 500 years rate (*F. Gasse, 1978 ; R. Paepe et al., 1982, 1983*). This 500 years oscillation can easily be derived from the curves of northern Latin America and Central Sahara as well. When the minor oscillations (450 - 500 years) covering the Late Glacial, the Preboreal, the Boreal and the Lower Atlantic, derived from the Central Sahara curve are taken into account, the question can be raised whether the Lower and Middle Holocene have in the other regions as well not been subject to 450 - 500 years oscillations. Indeed, since the lake Tchad environment (Central Sahara) is situated in a so-called "marginal" region (transition area of ecosystems), it is quite possible that its investigation material was exceptionally fruitfull (similar regions are the mountaneous vegetation belts) and that the sediment records in the other regions still hide this information. In this respect, taking into account the results of the comparison of all curves, the existence can be supposed of generally even more than fifteen cycles for the Holocene, putting forward a detailed and well defined subdivision of the Holocene in the eastern Mediterranean, based on transdisciplinary work, as was achieved in Greece, whilst Greece is no "marginal" region.

SYNOPTIC TABLE OF POSTGLACIAL CLIMATIC CHANGES



by Elfi VAN OVERLOOP (1984)

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PACDATA AND SIGNAL PROCESSING

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Since the initiation of PACDATA Bank, started two years ago, with the collaboration of N.Shackleton (Cambridge, UK) and J.C.Duplessy (C.N.R.S., France), other Geologists decided to confide their original paleoclimatological data. After having put them on computer usable form, we analyse them with appropriate data analysis methods and discuss the results with the owners in the framework of long term climatic changes. The main objective of this approach is to study the dynamical behaviour of the climate system both in space and time. For such a study, the appropiate paleoclimatic records are few due to their limitations in time and/or accuracy.
Signal Processing of climatic data plays a fundamental role in the understanding of the climatic system, the statistical predictability and comparison of climatic data strongly depending on their information content and therefore on their quality.
Modern data analysis methods provide valuable indications not only about the information content of the data but also on the extend to which they support the results of the climatic models.
In the time domain, high performance digital filters have been designed together with ARMA stochastic models.
In the frequency domain, linear and non linear spectral analyses and frequency transfer functions have been impremented (P.Pestiaux and A.Berger, 1983). These methods have been applied in order to analyse the insolation parameters and different climatological time series as surface temperatures or tree rings.
In the case of deep sea paleoclimatic records, the results permitted to
(1) evaluate the impacts of deep sea process on paleoclimatic spectra,
(2) emphasize the strong non linearities of the climatic response to orbital forcing at these time scales,

(3) confirm the role of the precipitation-temperature non linear feedback in the high sensitivity of monsoon countries to insolation variations, (J.F.Royer, *et al.* 1983).

(4) follow the time evolution of the spectral characteristics of the climate system during the quaternary.

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PALEOCLIMATIC DATA OF SEDIMENTOLOGICAL ANALYSIS OF THE GRANDE PILE PEAT BOG (VOSGES. FRANCE)

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Summary

In the "Grande Pile" peat bog, the following sedimentological studies were applied to the same samples as the ones G. Woillard used to build her famous pollen-diagram :

- heavy and light mineral composition ;
- sedimentary microstructures in thin sections ;
- grain size analysis ;
- organic carbon content.

They showed that :

1. grain size analysis and mineralogy indicate that loess supplies are more sensitive than pollens to detect of climatic transitions ;
2. and also to detect small fluctuations during the very cold periods ;
3. the paleoclimatic transitions are rather long, especially from the St.-Germain I temperate period to the Mélisey II cold one wich lasted more than 20 centuries ;
4. six ash layers were found in Grande Pile peat bog :

   4.1 - One at the top of Eemian (s.s.). This ash is different in composition than the Brørup one in Belgium. It has basaltic hornblend (55 %), titanite (25 %), green augite (15 %) and enhedral zircon (5 %). No enstatite.

   4.2 - four levels with small (100) yellowish - brown hollow sulphur balls. Those ashes were supplied by spray through some crater lakes, as one can nowadays find in some Costa-Rica volcanoes.

   The four sulpur balls levels are :

   - Melisey II (with 25 % in a fine sand compound)
   - Pile event (in traces)
   - Charbon event (6 %)
   - Marcoudan event (20 %) (1)

4.3 - One ash layer between 3.0 m and 4.50 m depth (in GP XVIII core) with bastlatic hornblend (65 %), titanite (30 %) and traces of green clino-pyroxenes and euhedral zircons.
This level is of Tardiglacial age. No further precision is possible. These ashes have been reworked from floating peat rafts. At this Tardiglacial period, pollens were also reworked. This part of the Grande Pile accumulation can not be stratigraphilly distinguished.

5. The rate of sedimentation was permanently changing during the filling of the Grande Pile lake. An important compaction mainly affected the gyttja layers. These postsedimentary microstructures have modified the initial tickness of layers.
6. $_{14}C$ absolute datations do not indicate improvement of climate at 22.000 y (Kesselt soil age in Belgium). The main stadials occured during the "Middle Weisschel" at 29.980 BP, 34.100 BP and 40.000 BP. (2).

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THE PHYSICAL EXPLANATION OF THE TYPICAL SPECTRAL SHAPE OF DEEP SEA CLIMATIC RECORDS.

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Abstract

The three main deep sea processes modifying the original paleoclimatic spectra are considered as a succession of three spectral transformations. They will be modelled to simulate their global effect on the original undisturbed paleoclimatic variability.

The two first transformations correspond to a reddening of the original paleoclimatic spectrum due to the long memory of the climate system and to vertical mixing (bioturbation). The total length of this memory has been estimated by an ARMA (p,q) stochastic model to range from 4 to 14 kyr.

A non-linear time dependent mixing model has then been designed to evaluate theoretical mixing functions, being given the diffusion coefficient and the sedimentation rate. The effect of the mixing has been evaluated using a cubic deconvolution, which allows demixing experiments with both experimental and theoretical mixing functions. This deconvolution method has been applied successfully to the first 170 kyr of the core V28-238, the restored paleoclimatic variability being of the order of that observed in high sedimentation rate deep sea records.

Finally, a Fast Fourier Transform Stretching algorithm applied to the core V28-238 showed that the variations in the sedimentation rate, which can be identified as a dilatation-compression process of the paleoclimatic signal, broaden the width and decrease the amplitude of the spectral peaks.

GEOMORPHOLOGIE PALEOCLIMATIQUE DANS LE BAS-ZAIRE (AFRIQUE CENTRALE)

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Le versant du massif du Bangu est successivement, de haut en bas :
- 1. un abrupt subvertical
- 2. des lobes de solifluxion
-3. un glacis pédimentaire.

Au bas du versant, le réseau hydrographique actuel s'est surimposé à un système de braided river.

-En milieu équatorial, une altération chimique intense a transformé le substratum du Bangu en argile. Les failles et diaclases subverticales, au contact de ces argiles de néoformation ont servi de plans de lubrification le long desquels des monolithes ont glissé. De plus des décrochages par solifluxion sont responsables de l'élaboration du relief en lobes en contre-bas de l'abrupt.

-En milieu aride, le ruissellement aréolaire était particulièrement actif. Ce processus d'érosion aplanissante a établi des glacis pédimentaires qui ont remanié le matériel de solifluxion. En contrebas des glacis fonctionnaient des braided rivers à séquences de graviers et de fines, qui ont édifié de larges plaines alluviales. Dans la région de Kimpese s'observent plusieurs niveaux de glacis pédimentaires, vers 420 m, vers 360 m et vers 320 m.

-En milieu de savane, comme au stade actuel lors de l'humidification du climat, un réseau hydrographique permanent, à large méandres libres, s'est d'abord surimposé aux braided rivers. Plus récemment, l'humidification croissante a progressivement détruit ce tracé en méandres libres, au profit d'un réseau en baionnette dû au réveil karstique des joints sous-jacents, surtout diaclases et failles plus propices à la dissolution des carbonates.

Au début du climat de savane l'activité des termites a été très importante. Lors de l'accroissement récent de l'humidité, les termites du type Macrotermes ont disparu. Cette espèce est encore vivace actuellement au Shaba, savane plus sèche qu'à Kimpese.

Conclusion

Le paysage de la région de Kimpese est le résultat de changements climatiques succesifs. Des formes du même type peuvent s'observer à différents niveaux. Des variations d'extension des glacis pédimentaires étagés indiqueraient des intensités variables pour chaque phase climatique.
En bref, l'évolution cénozoïque indique une succession de cycles comprenant une phase très humide à forêt ombrophile, une phase subaride à pédiments, une phase de savane à réseau hydrographique bien hiérarchisé.

MODELLING THE UPPER OCEAN FOR CLIMATIC STUDIES

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An oceanic mixed-layer model, specially designed for longterm simulations is presented. It uses Garwood's (1977) entrainment hypothesis but includes a stability dependent parameterization of the turbulent dissipation. This model is calibrated using results from laboratory experiments, observations at sea and numerical simulations with higher-order turbulence closure models.

A first simulation of the annual cycle of the sea surface temperature at the Ocean Weather Station (OWS) P (50° N, 145° W) has been obtained. The results are presented and discussed.
The model is now further tested. Simulations over several years at OWS P and N (30° N, 140° W) are in preparation.

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RADIATIVE APPROXIMATIONS IN CLIMATOLOGICAL STUDIES

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The heating terms, especially the radiative terms, are of crucial importance in climate modelling. The radiative processes indeed determine the energy available to the earth-atmosphere system while the other heating terms act only to distribute the available energy between surface and atmosphere and among different latitude zones.

In climate modelling, particularly in studies of the long-term climatic changes, the radiative calculations must achieve two opposite goals : accuracy, because of their importance, and speed, because of the large total amount of computation. Various approximative radiative schemes have been developped and compared as regards the surface and top atmosphere radiative balances (Tricot and Berger, 1981 ; Tricot, 1983).

Finally, in view of modelling man's impacts on climate (e.g., CO_2 or other trace gases concentration variations), a global radiative-convective model has been designed in which the vertical non-radiative heat transport is approximated by a simple numerical procedure called "convective adjustment" (Tricot, 1983). A first sensitivity study for a doubling of CO_2 concentration has given a surface temperature increase of about 2 K (figure 1 and Tricot, 1984).

This kind of climate model can be used as an important tool for the study of various properties of the global climate system, in the future and in the past. They can be considered as an important addition to GCMs to study certain feedback mechanisms, to estimate the climate sensitivity to forcings of anthropogenic and natural origin as well as to investigate the transient response of the climate system to some forcing.

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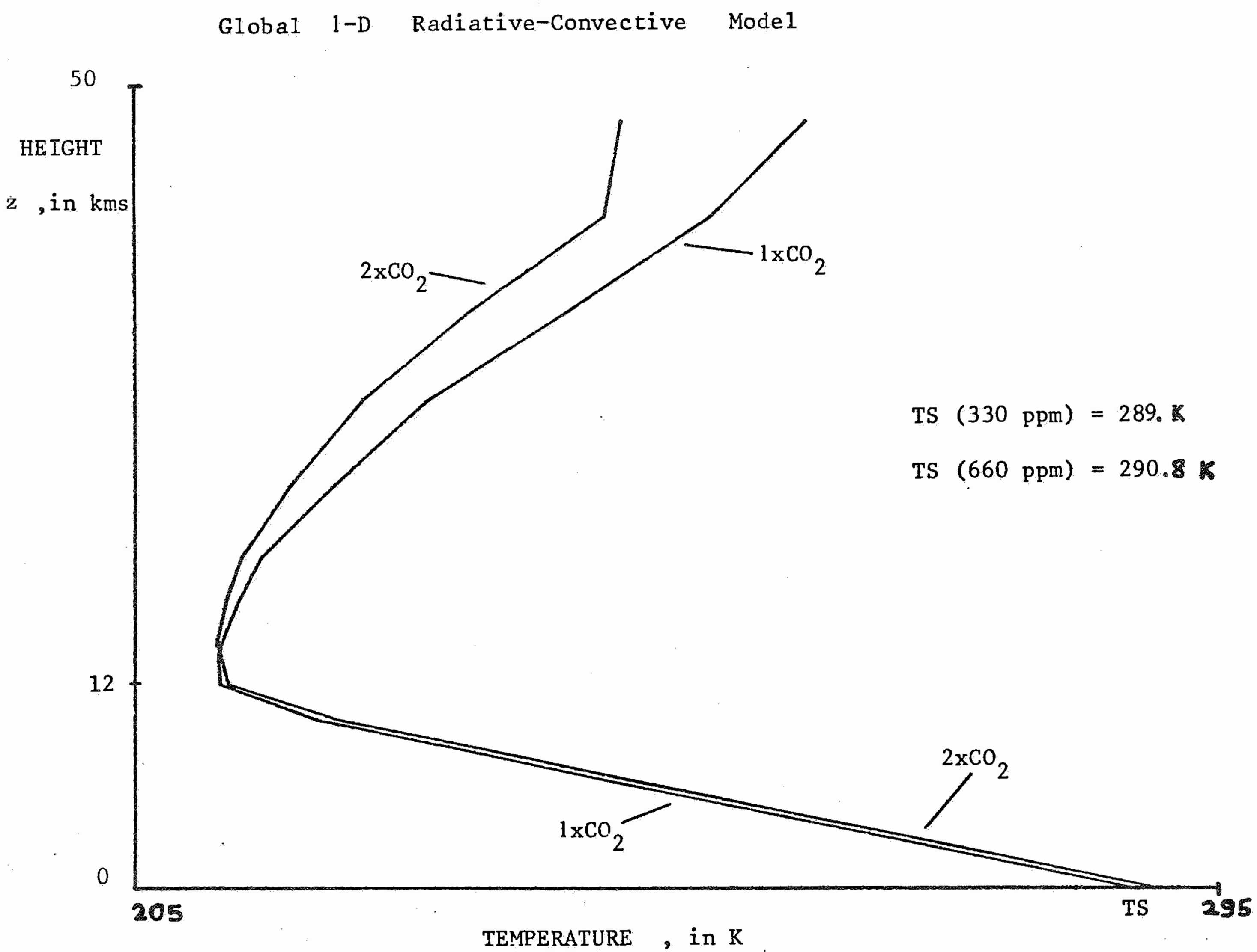

Figure 1 . Equilibrium temperature profiles obtained from the mean one-dimensional radiative-convective model for the normal (330 ppmv) and twice the normal (660 ppmv) amount of atmospheric carbon dioxide concentration. The tropospheric lapse rate is adjusted to 6.5 K/Km. A cloud layer is situated between 5 and 6 Kms, with a cloudiness of 0.5.

MODELLING ANTHROPOGENIC IMPACT ON CLIMATE : A CASE STUDY

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Abstract

Antropogenic perturbations have already affected local urban climates and are now implicated in global climatic change (Hansen et al., 1981). Land clearance in developing countries results in the removal of original forest and intensification and modification of agricultural practices in the reclaimed areas. Both these effects can have significant influences upon the local climate and ecology and may feed back to cause regional and global climatic perturbations. The changes caused to the environment which are believed to be of significance for the climate are : I) increased surface albedo ; II) perturbation of the carbon cycle causing variations in the atmospheric levels of CO_2 ; III) local changes in the water balance ; IV) addition of pariculates to the troposphere, both directly from combustion and by increasing the wind-blown dust and V) perturbation of the hydrological and turbulence characteristics over areas where tall forest stands are replaced by low crops or cleared land.

The consequence of these changes are difficult to predict since they almost certainly involve internal feedbacks within the climate system itself (e.g. modification of cloud cover as a result of surface albedo and water balance changes). Although the agricultural changes may be localized, the consequent climatic impact could manifest itself over larger areas. Climate modelling techniques are not yet sufficiently well developed to permit definitive statements about the magnitude of such perturbations. However, it is possible to consider the nature of the impact of these environmental changes and through ananlyzing model results to establish possible climatic effects. Results from both a simple 1-D radative-convective model and from a 3-D general circulation model are described. It is concluded that in terms of the potential for influencing climate at local scales deforestation is a highly significant land use change. However it is unlikely that surface albedo changes alone can modify the regional or global climate.

1. Introduction

Tropical rain forests are biologically diverse, multi-layered, predominantly evergreen forests, with little to no seasonality, heavy rainfall (200-300mm/month) and relatively constant temperatures (around 25°C). The dark, dense, moist vegetation gives these forests a lower surface albedo than almost any other natural or man-made area.

Deforestation is detrimental in the tropical environment, since most of the nutrients are concentrated in the above-ground biomass and therefore removal of the vegetation leads to rapid decrease in soil fertility (Jordan, 1982; Richards, 1973). The claim is often advanced that the removal of tropical rain forests will substantially alter the climate, either by adding CO_2 to the atmospere, thereby enhancing the greenhouse effect, or by increasing the global surface albedo (Bolin, 1977; Woodwell et al., 1978; Hampicke, 1980; Sagan et al., 1979; Potter et al., 1981; Shukla and Mintz, 1982). Such a claim merits careful investigation. In this paper, the climatic impact of albedo changes associated primarily with tropical deforestation are discussed. While tropical forests undoubtedly have considerable ecological value, the results of this study, which excludes the impact of tropical deforestation on the level of atmospheric CO_2, do not substantiate the assertion that their removal will have anything other than local significance for climate.

2. Simple Climate Models

Sagan et al. (1979) use results from the 1-D RC of Manabe and Wetherald (1967) to estimate a 2K temperature decrease caused by a planetary albedo change of 0.01. They further suggest that anthropogenic changes over the last 25 years have led to a global temperature decrease of around 0.2K. However the rates of vegetation change and the albedo values they use are questionable. Henderson-Sellers and Gornitz (1982) have repeated the calculations of Sagan et al. (1979) based on an improved version of their table (Table 1). Their calculated planetary albedo increase of between 0.00036 and 0.00067 will give rise to a much smaller temperature decrease of the order of 0.07K to 0.13K. This temperature alteration is probably too small to be detected above the interannual and longer period variability (Hansen et al., 1981). The calculation of the global albedo from the likely surface albedo changes also depends upon the appropriateness of the assumptions made about and values assigned to all of the other climatic variables. For example no feedback effects are included in the calculation of Sagan et al. (1979).

Hansen et al. (1981) in their 1-D RC climate model find a temperature decrease of -1.3K for a surface albedo change of +0.05. Here we use a linear interpolation of this result to estimate surface temperature change. The anthropogenic surface albedo changes given in Table 1, (last column) may cause a temperature decrease of between 0.02K and 0.01K. These temperature changes, which except in case (a), are forced almost entirely by the alterations in tropical forest areas, are very small. Henderson-Sellers and Gornitz (1982) suggest that within the error ranges of global 1-D RC climate models, the climatic impact of deforestation over the last 30 years is close to zero.

3. A General Circulation Model Study

The estimate of the possible impact of tropical deforestation, using simple climate models, suggests that the claims of substantial global climatic modification may have been premature. However these calculations considered the alteration caused by forest removal as a globally averaged phenomenom, which is unsatisfactory since deforestation is highly localized and therefore the effects of such local surface modification could be magnified and possibly transmitted by the action of general circulation of the atmosphere (Namias, 1979; Chervin et al., 1980).

The nature of local feedback effects which could amplify the climatic impact of tropical deforestation are complex. Initially the increased albedo is likely to be offset by the reduced ability to lose energy through evapotranspiration and surface temperatures may increase. The stripping of vegetation from grassland areas, however, (Charney, 1975) leads to a net cooling and hence an overall descent of air over the modified region. Initially convection may increase and therefore, if there is sufficient water vapour available (say, transported from an upwind source area), cloud formation and possibly precipitation will increase. Hydrologists have not yet been able to make detailed studies of the local vs. regional movement of water vapour, and therefore estimating the environmental impact of forest removal and agricultural irrigation is difficult. Modelling studies by Lettau et al. (1979) suggest that a considerable proportion of the precipitation over the Amazonian forest results from regional evaporation rather than from advected moisture. Convective activity may be enhanced by providing an effective heat source at the surface. Water consumption for bare soils and young partial vegetation cover is found to be between 400-500 mm per year, whereas mature forests consume from 750-900 mm per year (Baumgartner, 1979). The decrease in evapotranspiration must lead to increased local runoff if precipitation rates remain constant. The interactions between the perturbed energy and water cycles as a result of deforestation are likely to be very complex. Negative and positive feedback effects may exist and predominate at different times and heights in the atmosphere.

The interrelationships between radiation fluxes ,vertical velocities, the hydrosphere and surface can be investigated using a general circulation climate model. It is likely that while the local-scale change will be a function primarily of the nature of the local environment, even at this scale pertubations in precipitation may depend upon the regional climatic regime. Regional and global scale effects will be a result of feedback from the local change and will, once again, be a function of the geographical location of the perturbed area.

Here we report on the results of a simulation of deforestation using the GISS GCM (Hansen et al., 1982). In this study, the possible impact of tropical deforestation was maximized by concentrating all the alterations of the surface vegetation into one locality : a large-scale deforstation of the Brazilian Amazon region. The magnitude of the modification is equivalent to 35-50 years of deforestation at the current global rate concentrated in the Brazilian Amazon. This is therefore the locational antithesis of the global estimates considered in Section 2.

The Amazon river drainage basin covers approximately 7.8×10^6 km^2; of which over 85% is forested. Over 60% of the Amazonian forest lies within Brazil, while the remainder is divided among Bolivia, Colombia, Ecuador and Peru (Figure 1). Almost 58% of the total area of Brazil ($\simeq 8.5 \times 10^6$km^2) lies within the boundaries of Amazonia (see Figure 1). Thus approximately 5×10^6 km^2 of the nation's land is covered by this important biome. Climate and geography cause regional variations in the vegetation. A large portion (approximately 2.9×10 km^2) is lowland tropical moist forest, while vegetation occuring on the older, higher, crystalline soil is tropical forest, vs. equatorial forest in the lower basin sedimentary soils. Vegetation that is too close to the river to retain significant undergrowth is classified as varzea (occasionally inundated) and igapo (permanently flooded)

The climatic regime of northern Brazil is dominated by the seasonal movement of the equatorial trough but is also affected by the changing pressure fields at higher latitudes (Figure 2). The western part of the state of Amazones has an equatorial climate almost without seasons and heavy rainfall of 3000-3500 mm/year. The forest area to the south of the river, and the northern and southern states of Amazonia Legal (Raraima and Mato Grosso) show slight seasonal variations in rainfall and foliage density.

The Atlantic anticyclone is more important than the Pacific high pressure system since the Andes "protect" Brazil from changes occurring to the west. A heat low becomes well developed in the centre of the continent in the summer which results in slightly weaker easterly winds over the Amazon region compared to the winter pattern (Figure 3) although winds are generally weak and variable in all seasons. In response to the formation of the continental heat low in the summer, air flows southward from the Amazonian region. This phenomenon has been compared to a monsoon circulation pattern. The Amazon flows towards the east almost parallel to the equator, i.e. in the opposite direction to the prevailing easterly winds. Thus advected air loses moisture by precipitation as the continental divide is approached. Figure 2 and 3 include both climatological data and the results of simulations of the global circulation pattern by the GISS GCM (Hansen et al., 1982). Although modelled precipitation is lower than observed (Figure 4), the local climatology is fairly well simulated (Figure 5). Despite the generally weak winds and hence reduced turbulence, the Amazonian rain forest provides a very large energy source for the region, mostly through evapotranspiration ($\simeq$1300-1600 mm yr^{-1}). Insolation is high in spite of a consistently high percentage cloud cover; air temperatures remain high (24-26°C) and almost invariant. Fluxes of water vapour from the vegetation are higher than from most tropical oceans ($\simeq$1500 mm yr^{-1}). Rainfall is large and experiences a double peak in phase with the movement of the equatorial trough. Since the belt of heaviest rain coincides roughly with the position of the trough which reaches its extreme latitudinal positions in February and August, the peak rainfall on the equator occurs around May and November (Figure 4 upper). Total annual rainfall can be as high as 2700 to 3200 mm yr^{-1} in the upper basin. The almost continuous heavy rainfall in the upper reaches cannot be due to advected moisture alone and must therefore result from a partially closed hydrological cycle. Much of the water within the Amazon basin is recycled through the atmosphere, possibly more than once (Molion 1975 and Lettau et al., 1979).

Detailed climatological data for the region are extremely difficult to obtain. There were only 515 climatological stations in the whole of Brazil in 1976 (Schwerdtfeger, 1976) of which only 4 were radiosonde stations.

The studies of Molion (1975) and Lettau et al. (1979) were based on hydrological data from as few as 28 stations in the Amazon basin. Despite this dearth of data it can be seen that any climatic modification which perturbs the hydrology and/or radiation regime of this area could have fundamental significance. The climatology of two observation stations, Algrete and Belem, are compared with model-derived rainfall and temperature statistics from the GCM in Figure 5. The regime of these two locations are surprisingly well simulated in view of the coarseness of the model's horizontal resolution (8° x 10°).

Deforestation in the Amazon has recently been observed to be accelerating. In 1979, a joint survey by the Brazilian National Institute for Space Research and the Brazilian Institute for Forestry Development using LANDSAT Imagery revealed that 1.5% of the total Brazilian Amazon forest had been cleared between 1975 and 1978 (Tardin et al., 1979). This represented a 170% increase in deforestation over the 3 year period from 1973-1975. Around 260,000 km^2 of forest has been cleared since the 1950s. If this estimate is correct (Myers, 1980), then over 5% of the original tropical moist forest has already been removed. It is important to note that it has not been established that climax vegetation ever existed over the whole of the area which bioclimatologically could support it (Sommer, 1976).

Prior to the LANDSAT survey (Tardin et al., 1979), an estimated 114×10^3 km^2 of forest had been cleared during the 11 year period 1966-1977, i.e. an annual average rate of 10^4 km^2. This was comparable with the official "expected" rate calculated by addition of the various known clearance plans (e.g. cattle ranches, small-holdings, other agriculture, highways and timber). This total suggests an average rate of approximately 12,800 km^2 per year. The annual removal rate seems to be increasing from 9,500 $km^2 yr^{-1}$ for the 1973-1975 period to 16,000 $km^2 yr^{-1}$ for 1976-1978.

The deforestation imput to the GCM simulation is much greater than any likely near future deforestation of Amazonia. It was intended here to maximize any climatic perturbation resulting from a 30-year total tropical deforestation for comparison with the global calculations which represent a minimum likely response from simpler climate models (Section 2). Therefore the albedo increase could cause an alteration in the global climatic regime. In the latter case a global temperature decrease would be expected. There is no such temperature deviation that can be detected above the natural variability of the standard climatic regime in our simulation. Even in the latitude band in which the surface change occurs (the zone of width 16° and centered at the equator) there is no detectable climatic perturbation. This seems to be because certain of the feedbacks effects resulting from the deforestation act in opposition to one another. For instance, surface temperatures rise in spite of an increased surface albedo because the reduced evapotranspiration dominates the response. Similarly enhanced convection which might be expected to result from occasional temperature rises appears to be opposed by the moist convective cooling. It seems unlikely, at least from the result of this simulation, that even massive deforestation of the Amazonian region can directly cause global scale temperature effects.

There remains, however, the possibility that the local scale changes already noted could become coherent enough to establish a significant regional departure from the standard model's circulation. If, for example the climatic change is large enough to affect ocean temperatures then a perturbation in the Walker circulation may occur. The possibility of such an interaction may, however, be hindered by the intervening topographic feature (the Andes). If such a regional anomaly were to be established, its effects could impinge upon the Hadley circulation. Such systematic regional departures were suggested by Potter et al. (1975 and 1981) from the results of their 2-D climate model. We have examined the hypothesis that the disturbance to the normal circulation pattern in the region of the Amazon bould be large enough to cause a feedback effect which feeds into the general circulation of the atmosphere by analyzing the vertical velocity field in the area of the perturbation. Figure 8, based on the averaged data from the standard run, shows the ascent of air over the continental regions between 8°N and 16°S latitude for January and July. The classical Walker circulation cells shows clearly in the January simulation (see Julian global tropical forest removal was concentrated on the Amazon region.

Figures 1 and 6 show the region of South America for which a simulated vegetation change was made. A total area of 4.94×10^6 km^2 of tropical moist forest was removed and replaced by grass/crop cover. The "deforestation" occured in 7 of the 8° x 10° grid squares of the medium resolution model lying between 7.8°N and 15.65°S and 45°-75° W. All the areas classified as tropical forest were altered to grass/crop vegetation type but other vegetation types already existing in the squares remain unaltered. The area within which the vegetation change occured and the 4 boxes for which averaged Amazonian statistics are derived are indicated in Figures 1 and 6. Both the control and deforestation experiments were conducted using an interactive GCM with fixed ocean transports. Interactive oceanic circulation simulation requires a considerable time period before equilibrium is achieved (e-folding time of 5-10 years), because of the ocean's heat capacity. An interactive simulation is therefore only possible with a computationally efficient model.

The perturbation of vegetation caused a number of immediate effects. The surface albedo increased as a result of the vegetation replacement. This effect is particularly noticeable in the near infrared spectral region where grass/crop cover is known to exhibit high albedoes (values range from 0.1 at 0.5 m to 0.35 at 1.0 m). The roughness length is also significantly affected by the vegetation change due to the small scale of topography in this region. An anticipated climatic response to the changes shown in Figure 1 would be a lowering of surface temperatures together with a reduction in turbulent fluxes from the surface to the atmosphere, similar to the results of Potter et al. (1975, 1981). However the field capacities, which are related to the vegetation in the GCM are also perturbed by the simulated deforestation. The consequences of the subsequent hydrological changes are found to reduce the surface temperature change to close to zero (Figure 8).

The effects of deforestation in the Amazon have been assessed by comparison with a control (standard) climatic simulation. In the case of the control run the data from which the maps in Figures 2-5 were drawn were taken from ten-year averaged data (years 6-15 of a 20 year run). The data shown in Figure 7 are from years 16-20 of the standard control simulation and the last five years of the ten year perturbation run. The control run produces reasonable simulations of the regional (Figures 2-4) and even the local (Figure 6) climatological regime. The results of deforestation were to alter significantly all the climate parameters considered, except the surface temperature. Figure 7 shows the departure of the hydrology from the control run particularly well. The rainfall has decreased by between 0.5-0.7 mm day^{-1} and evaporation and total cloud cover are both reduced. The effects of reduced rainfall and increased run-off are shown in the significantly lowered values of water available in both ground layers. These results contradict the simple assumption that the surface albedo is the most important parameter. Figure 7 shows that despite the increase of surface albedo from 11% to 17%, the temperature does not decrease. This is because the reduction in evaporation caused by a combination of less available water and reduced ability to transpire has offset the radiative cooling by an evaporative (or latent heat) warming; The overall effects is a negligible temperature change. It must be noted that the surface is not the only area in which the hydrosphere seems to oppose the input alterations. The decreased cloud also opposes the increase in the surface albedo, leading to a smaller averall albedo increase than would have been expected form a model which did not incorporate both atmospheric and hydrologic feedback effects, e.g. the model used by Sagan et al. (1979).

The local results of the deforestation experiment can be summarized as follows from Figure 7 : no change in the surface temperature; precipitation decreased by around 0.6 mm day $^{-1}$; evapotranspiration decreased by 0.4 - 0.5 mm day^{-1}; planetary albedo increased by between 1% and 1.5% as a combined result of the increased surface albedo and the decrease in cloud cover (the latter is a highly variable parameter - see Figure 7) but we calculate that it has decreased by between 5% and 15%.

There are two ways in which the changes described here could affect the global climate; either through a circulation modification or, as suggested by those using simpler climate models, the local planetary and Chervin 1978 and 1980). One of the major areas of ascent occurs over the Amazonian region. Figure 9 is a contour plot of a climatic derived following Chervin and Schneider (1976), which presents the perturbed vertical velocities minus the standard vertical velocities (i.e. those in Figure 8) divided by the standard deviation of the standard run where all variables are derived from 5 year simulations. Following Chervin and Schneider (1976) and Chervin (1981) it can be shown that for an absolute value of this statistic greater than 3 there is a significant difference between control and perturbation at the 99% confidence level. Figure 9 shows only a very small region of significant decrease in the vertical velocity above the deforested region. This is consistent with the reduction in the surface evapotranspiration and the reduction in the moist convective heating aloft. This latter effect becomes statistically significant at two levels in the atmosphere in January only. The resultant decrease in the vertical velocity over the deforested area is similar to that predicted by Charney (1975) as a result of desertification. In this case, however, there is no strengthening feedback and decreased upward motion never becomes descent. Nor does the effect ever penetrate beyond the area local to the perturbation.

During the course of the simulation, the excursions of the regional climate were carefully monitored (Henderson-Sellers, 1982). There seemed to be a slight tendency towards more variability in the deforested run. For instance, during the second year of simulation, a change in the relative magnitude of the winter and summer hemisphere Hadley Cell circulations was noted. This anomalous circulation pattern was also clearly visible in the regional values of vertical velocities over the deforested region. The upward velocities were decreased in all but the most easterly sector. However the anomaly did not persist and neither the vertical temperature nor the zonal wind celocities for the region shown in Figure 9 suggest any systematic change. Locally the changes are mutually consisted. The winter (July) surface temperature increases affect the second layer of the atmosphere. This penetration is also seen in the occurrence of moist convective cooling in the second layer as well as the lowest layer of the atmosphere. This seems to suggest that, due to the considerably reduced moisture flux from the surface, this layer remains unsaturated and therefore as subsidence occurs evaporation is cooling the air at higher levels. However as can be seen from Figure 9, the 5 year averaged vertical velocities suffered only a slight disturbance in one season, July. The Hadley Circulation, as monitored by the stream function, differed significantly from that of the standard simulation only in one area at a pressure level of 200 mb. Even after years of simulation the effects of a very large deforestation could be detected only in the immediate vicinity of the perturbation. Furthermore, there seems to be no reason to suspect that further changes will take place in the circulation pattern in subsequent years. We therefore conclude within the constrains of a model without variable oceanic circulation, there is no reason to suggest that any regional to global scale effects on the climate can be expected as a result of large-scale tropical deforestation.

4. Conclusions

The possible climate impact of tropical deforestation has been considered using two extreme prediction modes. The calculations presented in Section 2 were derived from simple climate models, which must estimate the effect of a climatic forcing factor as though it were globally averaged. The results presented in Section 3 are from the GISS 3-D global climate model. In this case, the probable effects of a highly specific alteration to the surface and atmospheric features of one locality can be considered. In both cases the outcome for the global climate was similar, i.e. the alteration caused to the global temperature regime was undetectably small or zero. This result does not imply that local changes will be unimportant. On the contrary the GCM results show that the effects upon the deforested area itself are considerable - e.g. a reduction in rainfall of around 200 mm $year^{-1}$. Since vegetation has been removed and therefore infiltration capacity decreased, despite this decrease in rainfall, soil erosion, run-off and river discharge are likely to increase causing further environmental deterioration (Aubreville, 1949).

One of our major conclusions is that the effects of the hydrosphere, which have generally been neglected in simpler climate models, must be include if useful simulations are to be produced (see also Manabe et al.,1981). The climatic effects of a surface albedo change are much smaller in a very moist atmospheric environment than in arid regions (cf. Charney et al., 1977 and Sud and Fennessy, 1982).

The changes, potentially important for the climate, currently being caused by man in the tropical forests can be listed under the headings of I) surface albedo; II) CO_2 levels; III) hydrospheric changes; IV) addition of particulates; V) perturbations to surface atmosphere exchanges and to the atmospheric part of the hydrosphere. Estimates of the perturbations caused can only be made if the rate and mode of vegetation removal are well known.

It is very important that we draw attention to the fact that only first order surface modifications have been considered here. No attempt has been made to consider the impact on the global carbon budget of either the initial removal of the forest vegetation nor the subsequent loss of this large sink of atmospheric carbon. Also it is likely that the forest clearance itself would produce atmospheric particulates as well as additional CO_2 (e.g. Seiler and Crutzen, 1980).

Here we have sought to quantify the current extent of tropical deforestation and to calculate its possible impact upon the climate primarily as a result of the albedo change. We conclude that even when considering extensive tropical deforestation it is unwise to suggest that the surface vegetation changes alone will produce a climatic perturbation except in the immediate environment of the removal

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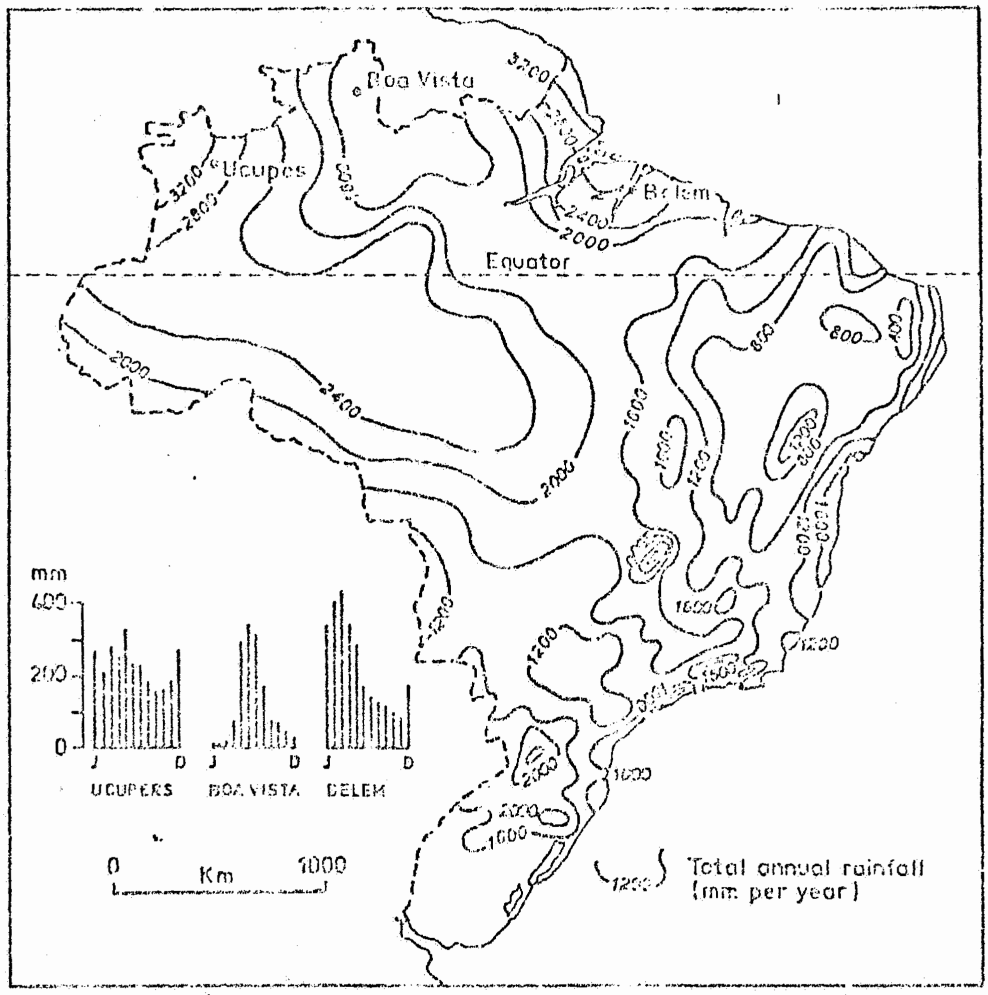
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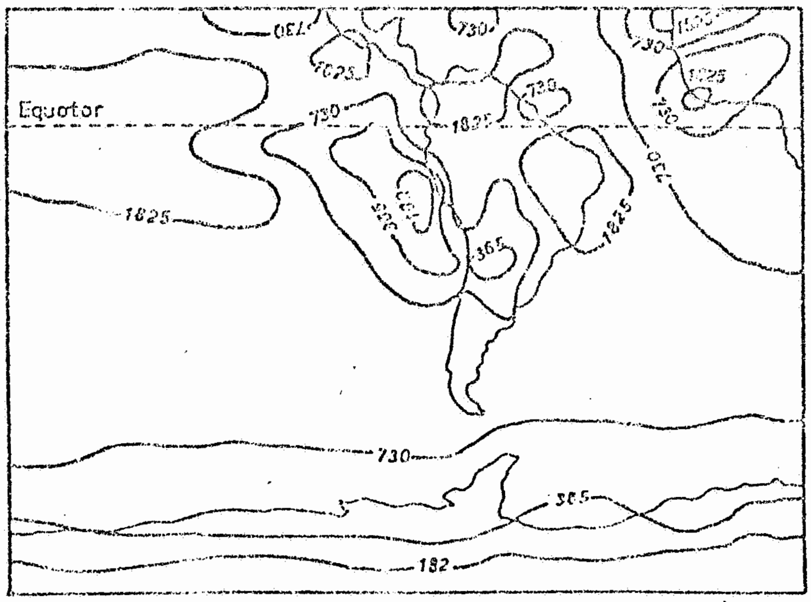
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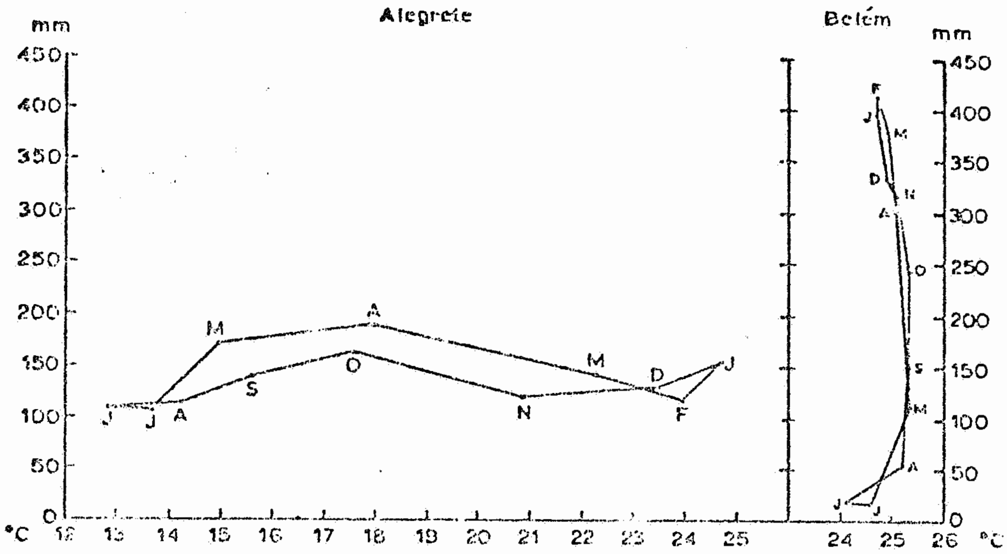
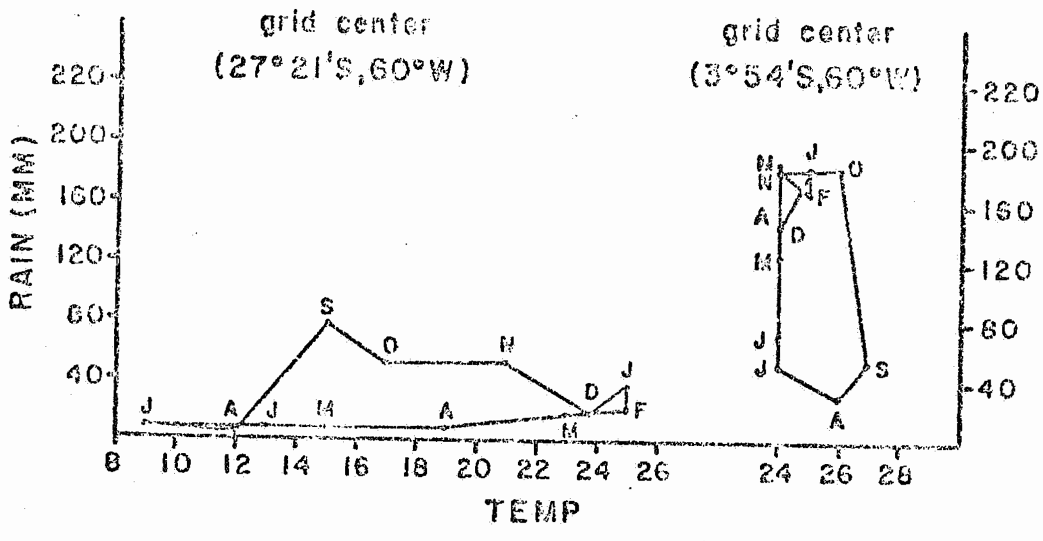
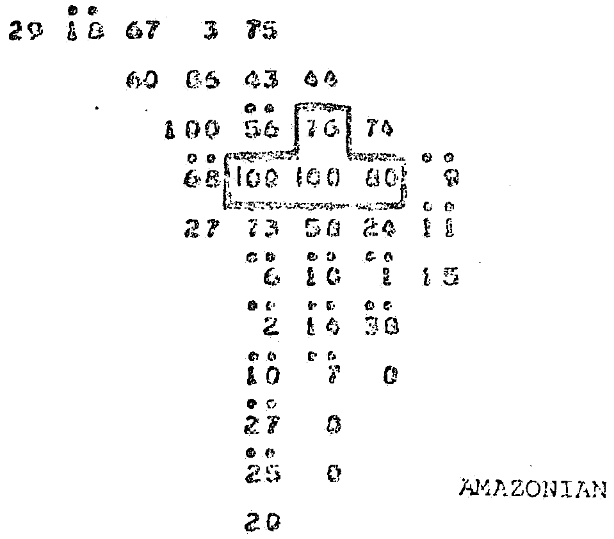


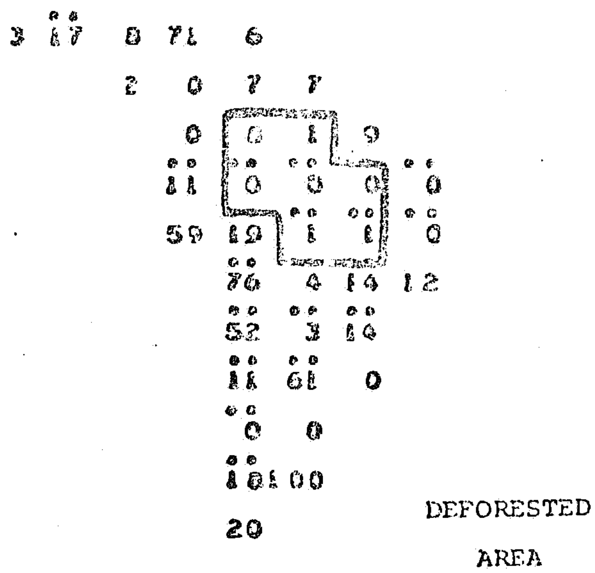
Fig.11. Climate diagrams; mean monthly temperature vs. precipitation, for Alegrete (29°46'S 55°47'W 104 m) and Belém (1°28'S 48°27'W, 24 m).



Ratio of rain forest to land area (percent) before deforestation



Ratio of grassland to land area (percent) before deforestation



12 MONTH RUNNING MEANS

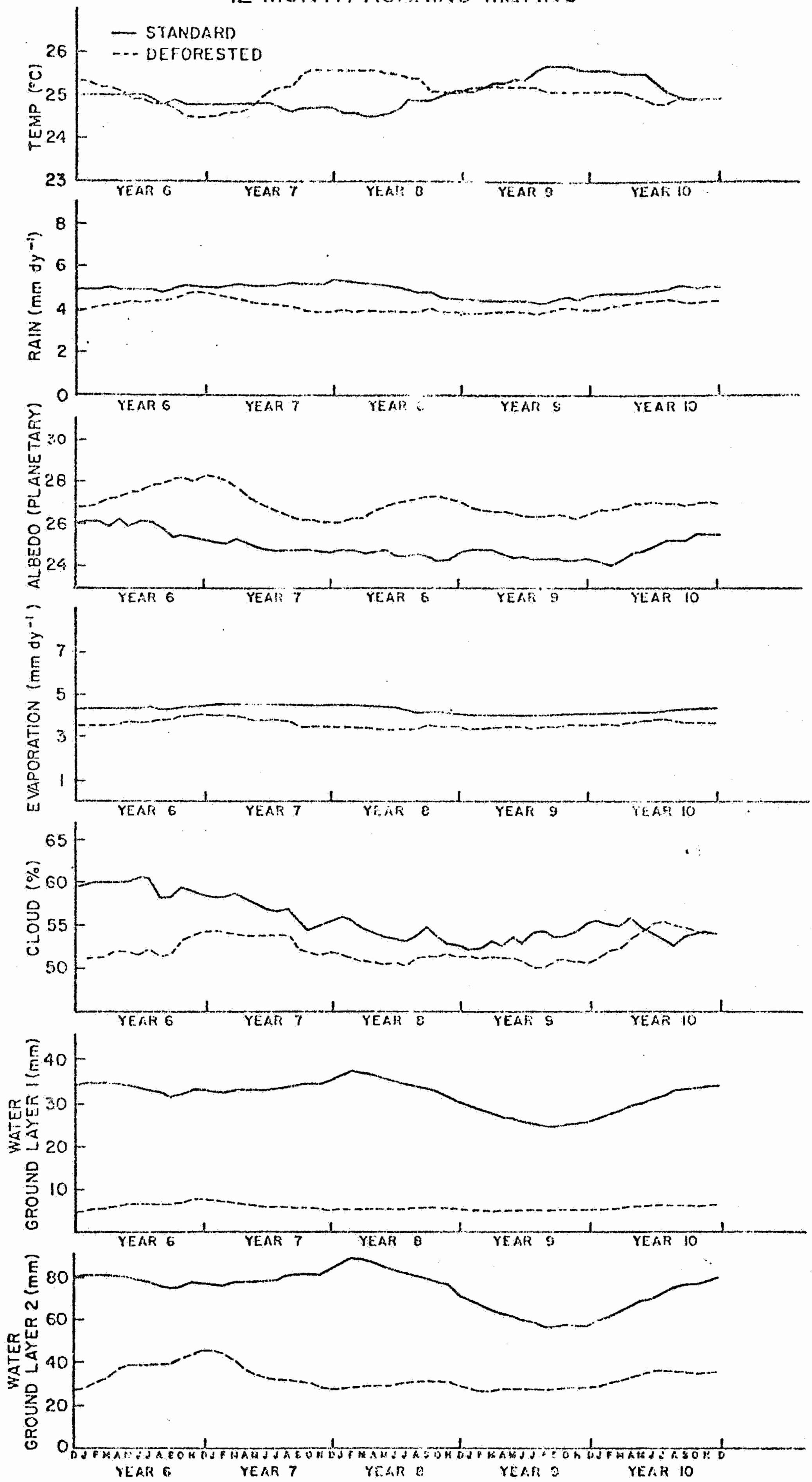


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Figure 1. Amazon region with the GISS GCM grid boxes in which asimulated "deforestation" occurred superposed. The numbers in the grid boxes are the total % of rain forest. This was modified to grass/ Crop in the experiment. Alegrete and Belem and the grid elemants from which statistics for Figure 6 are taken are also shown.

Figure 2. Sea level pressure for January and July: the control run (left) and observations (right).

Figure 3. Zonal wind (0.1 ms^{-1}) for January and July: control run (left) and observations (right).

Figure 4. Total annual precipitation: upper figure observations showing seasonal variation, lower figure control run.

Figure 5. Monthly precipitation vs. temperature plots for the two stations, Alegrete and Belem, cf. two grid elements (see Figure 2 for location).

Figure 6. Grid elements of the GISS GCM (medium resolution) showing % rain forest and grass cover in the control case and indicating the regions of "deforestation" and of averaging for statistics used in Figure 8.

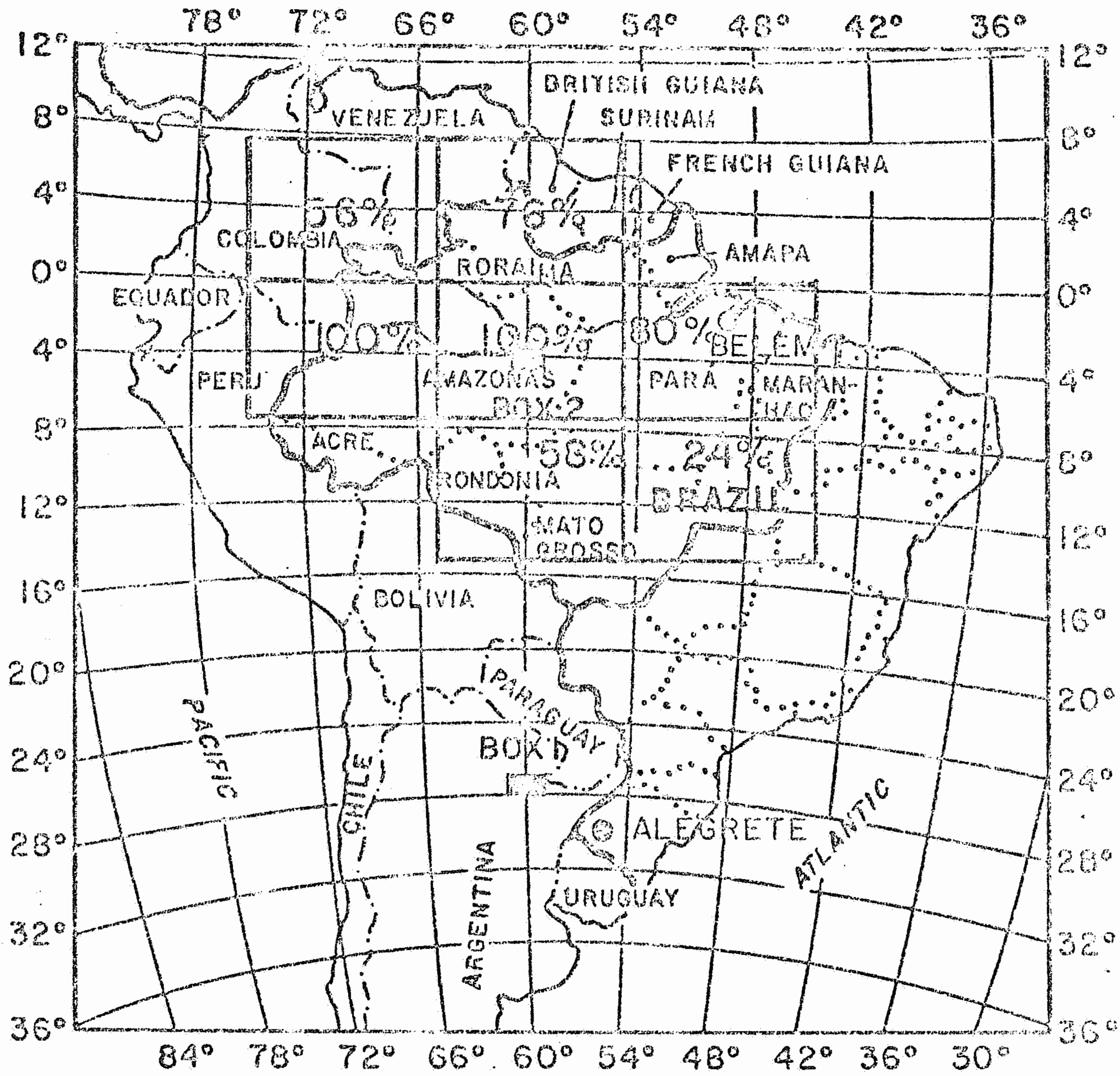
Figure 7. 12 month running means for the last 5 years of the control and deforestation runs.

Figure 8. Vertical velocity in the latitude band from 8°N to 15°S. (10^{-4} ms^{-1}) for January (upper) and July (lower). 5 years average from the control run. The cells of the Walker Circulation are clearly seen in the January case.

Figure 9. Climatic change test statistic - (deforestation-control)/ control standard deviation - 5 year averages (January - upper; July - lower). Values greater than $\pm$ 3 occur in regions where the perturbation vertical velocity differ significantly from the control field (99% confidence level).

Table 1. Global albedo changes due to anthropogenic modifications

Process	Land type change	Δrs, surface albedo change (integrated)	A, change in area, over 30 years, rel. to earth's surface area	F, cloud cover fraction	I, insolation factor	ΔR, fractional change in global albedo over 30 years	$A x \Delta rs$, relative surface albedo change in 30 years
Deforestation of tropical forests	Forest → savanna, grassland, field or pasture	10.8 → 18.5 = +7.7	0.00647	0.5	1.2	4.73×10^{-4}	4.98×10^{-4}
Deforestation of temperate forests	Forest → field pasture	13.2 → 19.22 = +5.3	small	0.5	0.8	small	small
Dam-building	Field → water	18.5 → 4.5 = −14	0.00059	0.25	1	-0.73×10^{-4}	-0.83×10^{-4}
Salinization	Field → saline field	18.5 → 24 = +5.5	0.000074	0.25	1	0.037×10^{-4}	0.04×10^{-4}
Irrigation of arid land	Desert soil → field	35 → 18.5 = −16.5	0.00072	0	1	-1.20×10^{-4}	-1.2×10^{-4}
Urbanization	(Field, pasture → urban	18.5 → 16 = −2.5	0.00059	0.5	1	-0.12×10^{-4}	-0.15×10^{-4}
	(Forest → urban	13.2 → 16 = +2.8	0.00059	0.5	1	$+0.13 \times 10^{-4}$	$+0.16 \times 10^{-4}$
Desertification	i) Scrub/shrub → desert soil	23.2 → 35 = +12	a) 0.00318	0	1	a) 3.81×10^{-4}	a) 3.82×10^{-4}
	or		b) 0.00159	0	1	b) 1.91×10^{-4}	b) 1.91×10^{-4}
	ii) Protected exclosure → desert soil	39 → 44 = +5	c) 0.00318	0	1	c) 1.59×10^{-4}	c) 1.59×10^{-4}
			d) 0.00159	0	1	d) 0.794×10^{-4}	d) 0.795×10^{-4}
Total surface albedo change						a) 6.66×10^{-4}	a) 6.82×10^{-4}
						b) 4.76×10^{-4}	b) 4.91×10^{-4}
						c) 4.44×10^{-4}	c) 4.59×10^{-4}
						d) 3.64×10^{-4}	d) 3.80×10^{-4}



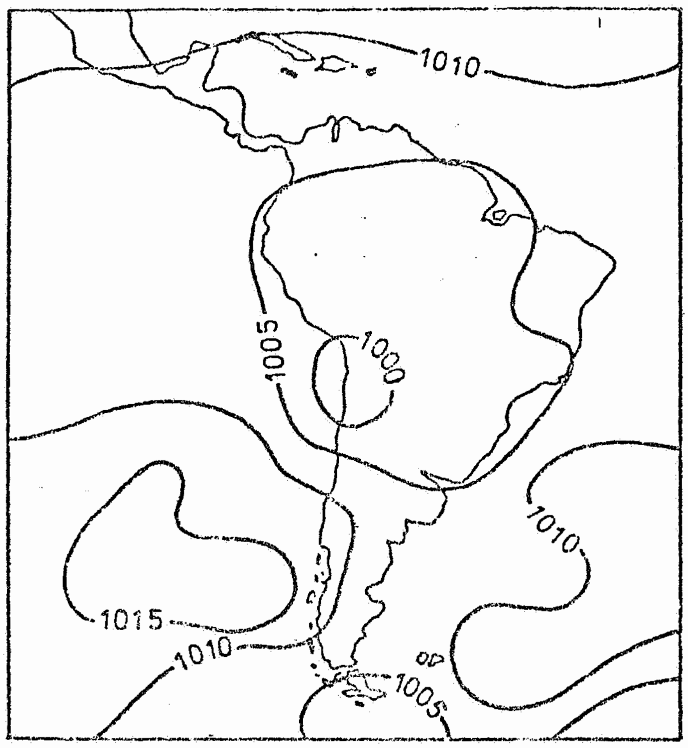
78° 72° 66° 60° 54° 48° 42° 36°
12° 8° 4° 0° 4° 8° 12° 16° 20° 24° 28° 32° 36°
84° 78° 72° 66° 60° 54° 48° 42° 36° 30°
BRITISH GUIANA
VENEZUELA
SURINAM
FRENCH GUIANA
56%
76%
COLOMBIA
RORAIMA
AMAPA
EQUADOR
100%
100%
80%
BELEM
PERU
AMAZONAS
BOX 2
PARA
MARAN-HAO
ACRE
58%
24%
RONDONIA
BRAZIL
MATO GROSSO
BOLIVIA
PARAGUAY
PACIFIC
BOX 1
CHILE
ALEGRETE
ATLANTIC
ARGENTINA
URUGUAY

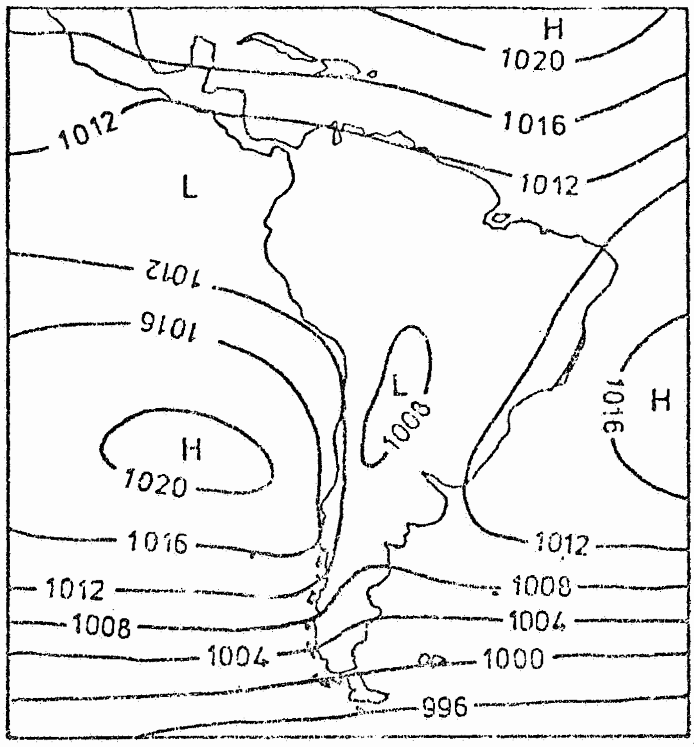
SEA LEVEL PRESSURE (Millibars)

MODEL
(Average over 10 years)

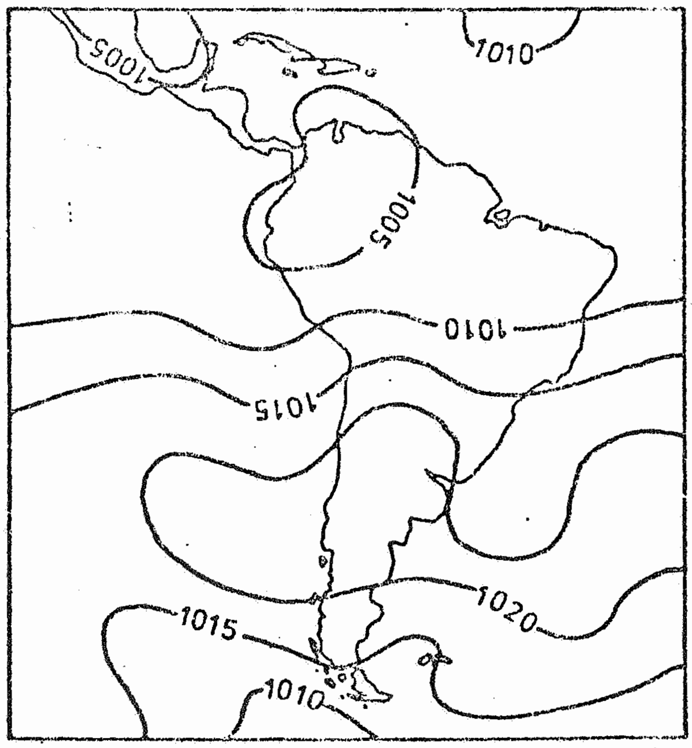
OBSERVATIONS

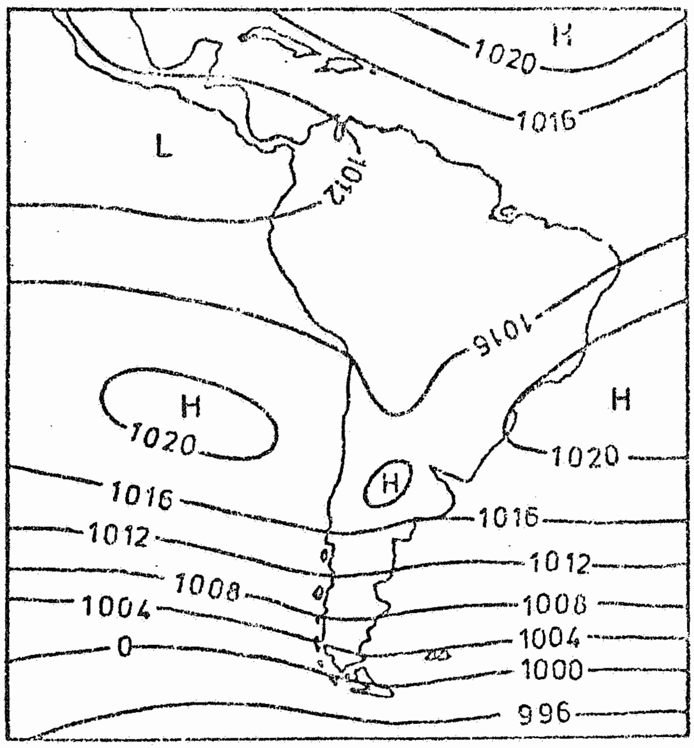
JANUARY





JULY





ON THE USE OF LARGE VALUES FOR THE DETERMINATION OF THE DISTRIBUTION OF MAXIMUM VALUES

by R. SNEYERS and M. VANDIEPENBEECK

Royal Meteorological Institute of Belgium

Abstract

When large values are independently distributed and when their distribution function is known, the distriburion function of the largest (i.e. maximum) value may be derived from the former distribution function. As a consequence of this principle, it is possible to give for the annual maximum of a meteorological variate good estimations for fractiles with large return period such as the secular maximum with using relatively short series of observations. Two examples are given with the use of five years of observations : a)the annual maximum windspeed at Uccle; b)the annual maximum of the daily rainfall at Uccle.

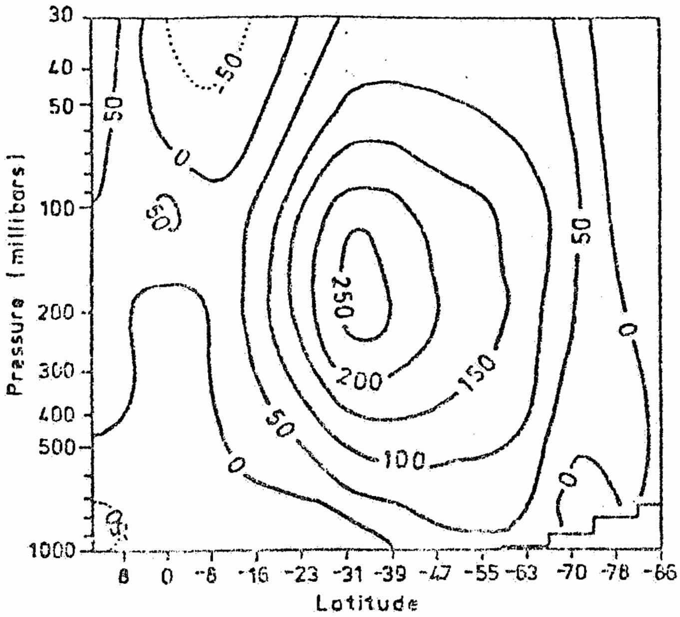
Résumé

Lorsque les grandes valeurs sont réparties de façon indépendante et lorsque leur fonction de répartition est connue, la fonction de répartition de la plus grande de ces valeurs, c'est-à-dire le maximum peut être déduite de la première. Il en résulte que pour le maximum annuel d'une variable météorologique il est possible de donner de bonnes estimations des fractiles à grande période de retour tels que le maximum séculaire à partir de séries d'observations relativement courtes. On donne deux exemples basés sur une série de cinq années d'observations : a)le maximum annuel de la vitesse instantanée du vent à Uccle; b)le maximum annuel de la cote pluviométrique journalière à Uccle.

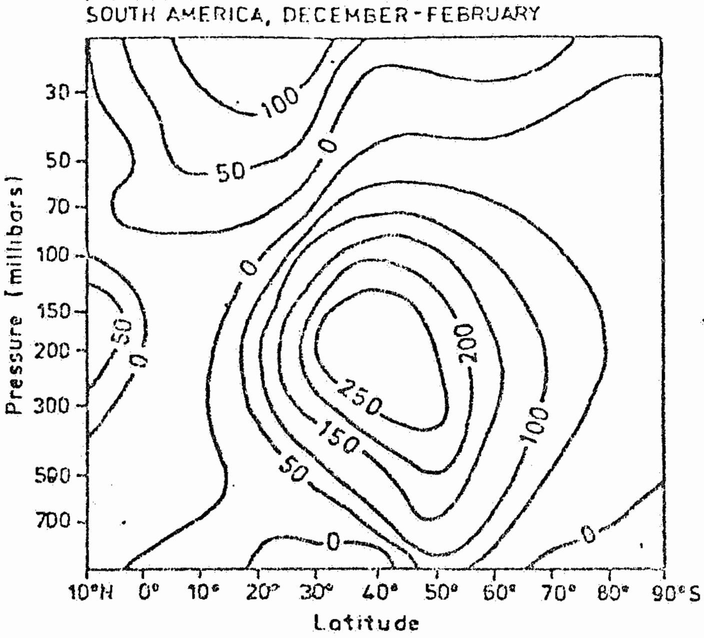
ZONAL WIND (U COMPONENT) TENTHS OF METRES / SECOND

JANUARY

MODEL (AVERAGE OVER 10 YEARS)

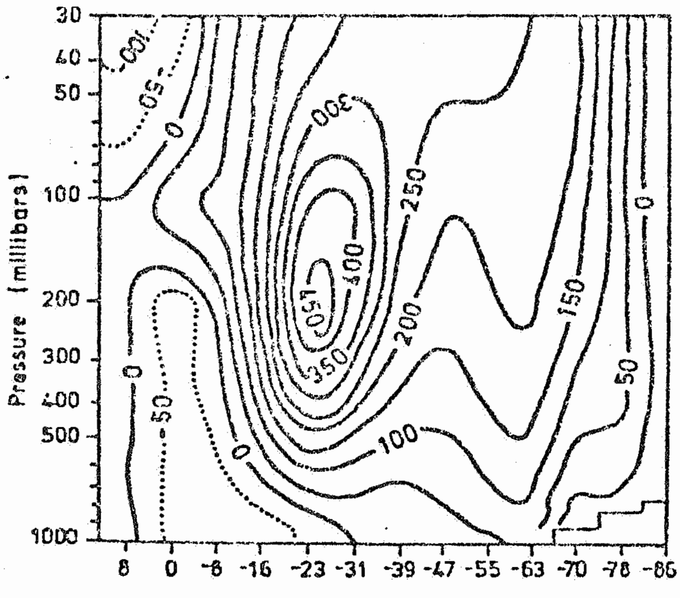


AVERAGE WEST TO EAST COMPONENT OF THE WIND (m/sec) BETWEEN 10°N AND 80°S LATITUDE OVER SOUTH AMERICA, DECEMBER-FEBRUARY

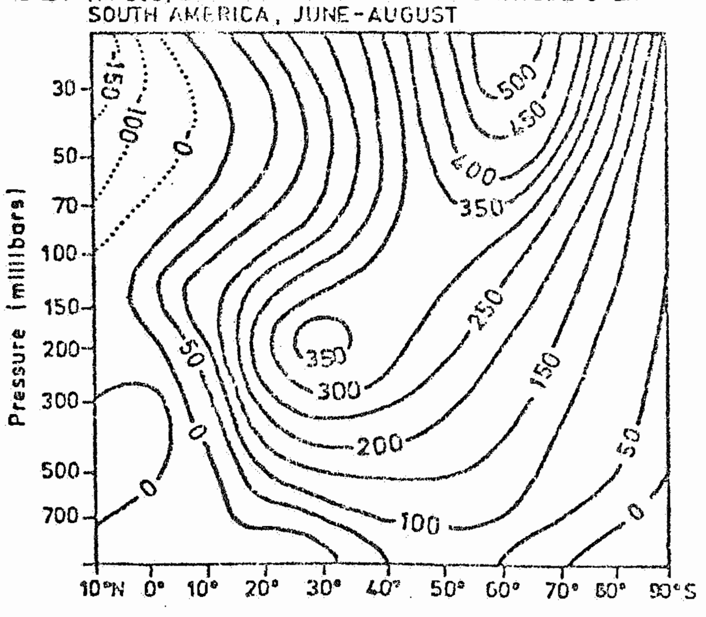


JULY

MODEL 824[8] (AVERAGE OVER 10 YEARS)



AVERAGE WEST TO EAST COMPONENT OF THE WIND (m sec) BETWEEN 10°N AND 80°S LATITUDE OVER SOUTH AMERICA, JUNE-AUGUST



ETUDE DENDROCHRONOLOGIQUE AU MAROC

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1348 Louvain-la-Neuve

Cette étude s'inscrit dans un vaste programme de recherches entrepris par le Professeur A. MUNAUT dans le bassin méditerranéen sur différentes espèces d'arbres convenant dendrochronologiquement.

Dans le cadre de notre doctorat elles sont essentiellement focalisées sur le cèdre (Cedrus atlantica) du Maroc afin de répondre à des problèmes concrets d'écologie forestière et pour effectuer ultérieurement des reconstructions climatiques.

Nos analyses dendrochronologiques ont fournis des séries chronologiques de cernes continues, définies à l'année près et s'étendant sur plus d'un millénaire, de 960 à 1979.

Différents paramètres dendrochronologiques et des relations cernes - climat révèlent dans les données dendrochronologiques la présence de gradients montrant les difficultés de croissance du cèdre en parallèle avec les gradients climatiques nord-sud et ouest-est existant au Maroc.

PRINCIPAL COMPONENT ANALYSIS OF THE MEDITERRANEAN RAINFALL

by Chr. GOOSSENS

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Principal component analysis has been used to group the European Mediterranean stations into homogeneous regions. The annual rainfall records used were from 90 stations distributed all over the European Mediterranean countries.

Major rainfall patterns were selected by a dominant-variance selection rule called rule N (a Monte-Carlo methode technique).

The most important component accounting for over 26% of the variance, comprises a fairly uniform field over much of the area covered, and hence represents rainfall events common to all stations. It is referred as a uniform pattern.

The second eigenvector accounts for 8 per cent of the variance and represents a gradient from northwest to southeast, with dry year in the northwest of Spain, the North of Italy and the South of Greece corresponding to wet years elsewhere and vice-versa.

A gradient in the same direction is likely to emerge as a third eigenvector (7 percent) with a considerable orographic influences superimposed. The orographic effects are slight over Pyrennees and Appenins but much more important over the Iberian mountains and the Alps.

The fourth eigenvector (7 percent) marks the influence of the cyclonic disturbances which originate in the Mediterranean Basin. The main region lies in the Western Mediterranean, producing Gulf of Genoa depressions. In the Mediterranean North-East region, the central Basin depressions tend to move to the northeast or eastwards where further intensification may occur in the eastern Basin area.

Eigenvectors beyond the fourth contribute individually little to the total rainfall. Generally, as the eigenvector number increases, the patterns they represent become smaller scale and represent shorter time fluctuations.

To summarize we may conclude that these eigenvectors classify the rainfall stations into some homogeneous rainfall groups. The regional grouping, based on the spatial patterns formed by the four major eigenvectors, shows the existence of six climatic regions :

- group 1 : north-west Spain and north Portugal which have a west-European rather than a Mediterranean coastal climate with rainfall exceeding 700 mm in much of the region, this rainfall is heaviest under the westerlies in winter, but considerable in summer also.
- group 2 : north-east and south of Spain, south of Portugal and south of France characterized by rainfall in the winter half-year, but very little rain in summer.
- group 3 : north-Italy which is characterized by a rainfall distribution all over the year, but in contrast to the regime of the Mediterranean climate, the summer half-year has more rain than the winter. The rainiest months are in autumn and spring, the driest are January and February.
- group 4 : the rest of Italy without Sicily, where most rain falls in the winter half-year, but summer is not rainless. However, the amount of rain becomes smaller and the rainy season shorter when going to the south.
- group 5 : Yougoslavia, Albania and north of Greece characterized by a single rain maximum in winter, summer being almost but not quite rainless.
- group 6 : Sicily and South of Greece characterized by considerably less precipitations, with dry and dusty lands through the long summer months.

DESERTIFICATION RESEARCH IN SOUTHERN TUNISIA

by Jean-Luc MELICE

Associate-Expert

World Meteorological Organization
Geneva - Switzerland

Research project of the WMO and the Geophysical Institute of the university of Alaska (Fairbanks, Alaska, U.S.A.).

This study is supported by the National Science Fundation (Washington D C) (Grant ATM 7915798) and by the World Meteorological Organization.

This study proposes to investigate "climatic" feed-back mechanisms of desertification in the Tunisian region. Emphasis is placed on the effects of surface radiative properties as a consequence of modifying surface features by overgrazing and on the effects of associated increased atmospheric turbidity.

Interpretation of the impact of overgrazing on desertification not only is of interest as fundamental knowledge in atmospheric sciences but has profound societal relevance associated with the world food supply.

We will summarize briefly the work already accomplished with this project and the work wich is underway :

1. Ground-based measurements were carried out in the semi-desert of southern Tunisia. These measurements provided results on the ground surface temperature and albedo differences between denuded, semi-desert and vegetated areas. For example, albedo differences between grazed and protected land were found to be in the order of 10%, and bare soils, even with a higher reflectivity, was always warmer during times of bright sunshine than vegetaded soil. Previous controversy in the litterature on this topic might have been clarified through these results.

   Turbidity and aerosol measurements were also carried out in southern Tunisia. The winter months wich comprise the rainy season in that area showed lower values of turbidity than the summer months, a result to be excepted since most of the aerosols are dust particles.

Spring showed the largest number of sand storms and turbidity was high from this time on, until the approach of the next rainy season. Values found approaches those of heavily industrialized areas of the United States. Model calculations on radiative transfer using the measured data (turbidity and surface albedo) as an input were carried out. Perturbation experiments showed that an increased surface albedo (desertification) and an accompanied turbidity increase would have somewhat compensating effects on the planetary albedo.
Over the last 80 years, there has been no decrease in precipitation in southern Tunisia wich is statistically significant. For the semi-desert, a quasi-biennial oscillation was detected and further, there is an indication that the sunspot cycle was detected in the precipitation data.

2. Quantitative analyses of satellite data for the evaluation of the desertification process in the semi-desert of southern Tunisia over the last decade will be carried out. Good results of these analyses will be expected after taking into account atmospheric attenuation wich are obtained from our turbidity and "ground truth" measurements.
Measurements of the surface heat balance and the atmospheric boundary layer over semi-desert and oasis will also be carried out. These data will be the basic input for a radiative-convective model wich will be applied in order to check the climatic stability of the desertification process.

PRELIMINARY ANNOUNCEMENT

CLIMATE AND DESERTIFICATION, ERICE, SICILY, 10-22 OCTOBER 1983

by J.P. VAN YPERSELE
Vice-Director of the Course

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Université Catholique de Louvain
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The Third Course of the International School of Climatology (Director : Professor A. Longhetto) will be held at the Ettore Majorana Centre for Scientific Culture, Erice-Trapani, Sicily, from 10 to 22 October, 1983. It will be devoted to "The Climatological Aspects of Desertification : Facts, Theories and Methods".

This Course intends to go beyond the controversy about the main cause -man or climate ?- of desertification, understood here as the various processes leading to a reduction of the biological potential of the land ; it will try to address the questions : what is the present state of knowledge on the climatological aspects of desertification ? How to make the best use of the existing atmospheric data for a better management and planning in regions at risk ? What are the gaps to fill ? How can modern remote sensing techniques be applied to this problem ? What do the climatologists need to know about desertification ? In which way can the atmospheric scientists, in collaboration with scientists from other disciplines, contribute to the understanding, the monitoring and finally the control of desertification ? How to present the climate-related information to the potential users ?

The goal of this interdisciplinary Course will not only be achieved through the lectures, but also through the informal discussions between all the participants. Lectures will be delivered by the 25 specialists who will review the following subjects : Ecology and climate of arid, semi-arid and subhumid regions, Desertification processes, Wind and water erosion, Salinization, Drought and bio-geophysical feedbacks, Climatological and bio-geophysical monitoring of desertification processes. A panel discussion will also be devoted to the interaction of climate variability and human activities leading to desertification.

A set of recommendations for further research/action also emerge from this meeting.

This Course is intended for people having a background in meteorology, geophysics and/or ecologie, but also to indivuals from other disciplines interested in desertification. The programme has been designed as much to supply up-to-date information to researchers already working in the field of desertification as to provide a high-level background knowledge to all geophysicists (including graduate and post-graduate students) planning to work on related topics.

Some fellowships available for travel and/or living expenses will be awarded on a competitive basis. The number of participants will be limited. For further information and applications, please contact the Director of the Course, Dr. R. Fantechi, Commission of the European Communities, Environment Research Programmes (DG XII), Rue de la Loi 200B, B-1049 Bruxelles, Belgium, or the author of this sheet.