Impacts of climate variations on water management and related socio-economic systems

by Arie S. Issar

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A review and reinterpretation of existing information

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Foreword

Although the total amount of water on Earth is generally assumed to have remained virtually constant during recorded history, periods of flood and drought have challenged the intellect of man to have the capacity to control the water resources available to him. Currently, the rapid growth of population, together with the extension of irrigated agriculture and industrial development, are stressing the quantity and quality aspects of the natural system. Because of the increasing problems, man has begun to realize that he can no longer follow a "use and discard" philosophy -- either with water resources or any other natural resource. As a result, the need for a consistent policy of rational management of water resources has become evident.

Rational water management, however, should be founded upon a thorough understanding of water availability and movement. Thus, as a contribution to the solution of the world's water problems, UNESCO, in 1965, began the first worldwise programme of studies of the hydrological cycle -- the International Hydrological Decade (IHD). The research programme was complemented by a major effort in the field of hydrological education and training. The activities undertaken during the Decade proved to be of great interest and value to Member States. By the end of that period a majority of UNESCO's Member States had formed IHD National Committees to carry out the relevant national activities and to participate in regional and international co-operation within the IHD programme. The knowledge of the world's water resources as an independent professional option and facilities for the training of hydrologists had been developed.

Conscious of the need to expand upon the efforts initiated during the International Hydrological Decade, and, following the recommendations of Member States, UNESCO, in 1975, launched a new long-term intergovernmental programme, the International Hydrological Programme (IHP), to follow the Decade.

Although the IHP is basically a scientific and educational programme, UNESCO has been aware from the beginning of a need to direct its activities toward the practical solutions of the world's very real water resources problems. Accordingly, and in line with the recommendations of the 1977 United Nations Water Conference, the objectives of the International Hydrological Programme have been gradually expanded in order to cover not only hydrological processes considered in interrelationship with the environment and human activities, but also the scientific aspects of multi-purpose utilization and conservation of water resources to meet the needs of economic and social development. Thus, while maintaining IHP's scientific concept, the objectives have shifted perceptibly towards a multi-disciplinary approach to the assessment, planning, and rational management of water resources.

As part of UNESCO's contribution to the objectives of the IHP, two publication series are issued: *Studies and Reports in Hydrology* and *Technical Papers in Hydrology*. In addition to these publications, and in order to expedite exchange of information, some works are issued in the form of *Technical Documents.*

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PREFACE

This report concludes the work can-id out in the framework of UNESCO'S IHP Project IIc, "The impact of climate variations on water management systems and related socio-economic systems". In order to be able to predict impacts, once climate change occurs, the principle of "the past is a key to the future" has been adopted. This meant the collection and interpretation of the numerous studies already done on climate changes during the Holocene, followed by an assay to correlate them on a global scale. In order to accomplish such a task of correlation, a base climate change reference section had to be adopted. The sequence suggested by Blytt & Semander (in Lamb, 1977) for Europe, and adopted by most Holocene paleoclimatologists, was found not to be accurate enough, a climate change base-section for the Levant has been compiled. This was used later for correlation with Holocene climate profiles suggested by investigators for the different regions. Agreements and disagreements with the Levant base section were discussed. In other cases when only part of the stratigraphic section was investigated or only general views expressed, reference has been made without discussing the various and differing interpretations.

It goes without saying that this report does not pretend to match the voluminous work which was done by Prof. H.H. Lamb. It is aimed to accumulate the data that has been gathered since the 70's and on that basis to suggest a more detailed paleoclimate section for the Holocene, drawing conclusions regarding the future.

The author is well aware of the fact that the principle of the "past is a key for the future" as well as the method of correlation used, has serious limitations. In the first place the reconstruction of past climates is dependent on the limitations of data as well as on objective and subjective interpretations. Secondly, the correlation between the numerous sections was not based on statistical methods, but on a general comparison. Yet, notwithstanding these limitations, a rather detailed paleoclimatic columnar section for the Holocene seems to come out, and a rather useful correlation between the Levant, the Circum-Mediterranean region, Europe, Africa north of the equator, Northern America and China is available.

On this basis general conclusions regarding the climate change can be deduced. In summary, in case of a general warming of the globe, the regions dominated by the westerlies, will become drier, those in the monsoonal area will become more humid. In the case of general cooling the reverse will be true. This is in agreement with the forecasts derived from the General Circulation Models.

This means that warming from a "greenhouse" effect maybe of benefit to a big part of the monsoonal lands (Including Egypt and Sudan which are dependent on the Nile, which is partly fed by the monsoonal rains). Taking into account that the population of these countries is an important part of the total population of the world, and is mostly dependent on agriculture, the question rises whether all the issues of possible future greenhouse effects have not to be re-discussed. It goes without saying, of course, that it is not recommended that the efforts to reduce the pollution of the atmosphere by greenhouse gases should be stopped, since for the time being at least there is no substitute for fossil fuels as the main source of energy. It is suggested, however, that a search should be pursued for policies which will integrate the reduction of pollution with benefits of this unavoidable process.

It is also recommended that the bibliography and reference list included in this report should be considered a contribution to a data base needed for advancing studies of correlations between the numerous paleoclimatic sections -- especially so with regard to the continuation of this work to the continents south to the equator. It is clear also that the base section itself, which should be regarded as tentative is in need of further verification, if not changes.

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1. INTRODUCTION

In this report the impact of an expected global rise in temperature on water management systems and related socio-economic systems is discussed. This expected rise is related to the "greenhouse effect", namely the continuous increase of CO_2, CH_4 and other "greenhouse" gasses in the atmosphere.

The prediction of the temperature increase on the global and regional climate is generally investigated by the running of Global Circulation Models (GCM). The evaluation of the results obtained by these models is extremely difficult because of the very large number of parameters within the system and the often unknown or poorly known sets of interactions involved. There are extraterrestrial planetary, solar, atmospheric, hydrospheric and lithospheric factors which exert influence on the climate on earth, often in an intricate manner, which is not comprehensively understood (Morner, 1984b; Fairbridge, 1984). Ruddiman & Duplessy (1985) maintain that although the driving force behind a large fraction of Quarternary climatic change lies in orbital variations, the generally dominant response of the earth at the 100,000-year periodicity cannot be a linear response to orbital forcing. The 100,000-year period of orbital eccentricity acts mainly to modulate the amplitude of the 23,000/19,000-year precessional rhythm of the received insolation signal, while internal climatic feed-back effects, in addition to the Milankovitch orbital forcing, are necessary to explain the deglaciation. In addition, there might be other phenomena that act as triggers to climate change, especially when they happen simultaneously with an orbital insolation signal -- for example sun spots (Degens et al. 1984) and volcanic emanations (Lamb, 1968).

Even when these difficulties can be assumed to have been overcome, one is faced with the difficulty of the construction of hydrological models connecting the climate and the hydrological cycle. For example, the validity of the models to predict future precipitation is questioned by Kukla (1990), who points to the fact that . . . . . although the general mean global increase of precipitation is predicted by all models, the differences among the individual predictions of regional precipitation rates in the CO_2 enriched atmosphere are such that none can be trusted. The more serious question is raised by the fact that the models are consistently overestimating current precipitation in the low latitudes. Could this mean that some of the processes operating in the real world climate system such as for instance convection, are incorrectly represented in the models, forcing them to simulate unrealistically high precipitation? If so, isn't it possible that the predicted precipitation increase in the CO_2ich world is exaggerated as well? It is also disquieting that the past insolation input was unable to increase precipitation in the areas where the ice started build at the onset of the last glaciation. Since precipitation is intimately related to the cycle of atmospheric water vapor, which is by far the most important greenhouse gas and since the sensitivity of surface temperature to the atmospheric CO depends closely to the water vapor pressure (Raval and Ramanathan, 1989) it is only natural to ask whether the GCM temperature forecasts for the CO_2 rich world could not be flawed as well. It is necessary to keep in mind that only a small fraction of the solar energy received by the earth system is absorbed in the atmosphere. The bulk of it is received and redistributed in the oceans. The oceanic circulation did change substantially in the past. This is extensively document by abundant paleoclimatic evidence. In our current climate models we presume that the pattern of oceanic circulation did not change in response to a major insolation shift or that it will not react to a major radiative perturbation such as the CO_2 doubling. This is an unrealistic expectation. The reliable prediction of the future CO_2-rich climates, both in terms of precipitation as well as surface temperatures requires significant upgrading of the modelling efforts. Only a new generation of models linking atmospheric and oceanic circulation can lead to reliable climate forecasts."

Moreover, there are other determining factors which are poorly understood: extraterrestrial influences and the effect of magnetism, gravity and heat production from within the lithosphere of the planet earth (Mërner, 1984). perhaps even these factors ought to be included in climatic simulation models in order to arrive at more realistic predictive models. With the present state of knowledge, however, this seems unlikely.

These difficulties are augmented when changes in temperature and precipitation have to be interpreted in terms of runoff, infiltration to ground water, not to speak of their impacts on socio-economic systems, which also involve human responses to changes in the natural environment. These may be caused not only by changes in the quantity of precipitation, but also in the intensity, and seasonal diversity. In order to be able to predict hydrologic responses reliably, hydrological models capable of simulating the effects of climate variations on the hydrological cycle must be employed. The reliability of these depends very much on the their success in simulating observed processes. Because in most regions climatic and hydrological data is restricted to a few decades (and very seldom to more than a century), the simulation runs of the models are restricted. These run times become even shorter when countries of the Third World where such information is generally scarce. A partial answer to this problem is in the construction of synthetic climate and hydrological curves on the basis of a multidisciplinary proxy data. This includes all sorts of data, such as ancient levels of lakes and sea, soil and lake bottom profiles, botanical data such as tree rings and pollen assemblages in sediments of marshes, as well as data derived from historical and archaeological research. Although such are not directly quantitative, it can be incorporated in models in a semi quantitative way, as it shows what changes can result and the extent such variation might reach.

This approach may even promote a greater understanding of the atmospheric and hydrological systems. In other words it may help to open the "black-box" or "boxes" which are often hidden behind the stochasticity of Climatological and hydrological phenomena. For that reason the data obtained while investigating climate changes is important, as it may serve to help calibrate the climatic models when they are run backwards in time. This calibration is necessary due to the fact that these models are very complex and incorporate many assumptions which can be tested by the observations derived from the historical data.

We can summarize this approach in the following way: The knowledge and understanding of the changes of the paleo environment are the key for the Understanding of the impact of climate changes in the future. Hence the analysis of former climates may give us an indication of what to expect in the future, if the temperature were to rise globally or regionally. Moreover, paleoclimatic data are necessary in the development and testing of climatic models (Kutzbach and Street-Perrott, 1985).

In order to apply this approach a general survey was made of the climate dependent records during the Holocene, in the Mediterranean basin. These included environmental isotopic data from core holes and speleothemes, palynological and dendrochronological data, ancient levels of sea shores and lakes, as well as archaeological records regarding phases of settlement and desertion of the desert. From these the most significant co-variations which could be assumed with a high degree of certainly to be a function of climate change were designated. From these were also selected the most pronouced co-variations of which a preliminary survey had shown a global significance. For each of these the paleo-environmental and historical records were investigated in order to determine the impacts of the climate changes on the hydrological and the ancient socio-economic systems. Special emphasis was put on the records from the regions of the ancient civilizations, as their historical (and thus paleo-hydrological) and socio-economic data is abundant. The two main areas on which the present report will concentrate will be the Circum Mediterranean region, and China.

It should be emphasized again that this report does not pretend to match with the voluminous work which was done by Prof. H.H. Lamb (Lamb, 1977), It is intended to study the data which has accumulated since the 70's. This is especially important since the sequence suggested by Blytt (1876) and Semander (1908) has been found not to be sufficiently accurate. Thus, in order to make this study more relevant to actual problems, a special emphasis has been put on co-variations, which identity simultaneous climate changes in the past, and inter-regional comparisons, which is the basis of stratigraphy (Bruins and Mook. 1989).

An important aspect to be emphasized at the beginning of this report is that of the geographical extent of climate change, especially the warm periods in the past. The anticipated CO_2 induced greenhouse effect is usually expected to result in warmer temperatures on a world-wide scale. Hence the popular term: Global Change. Yet, many of the warm periods in the past have had a regional or hemispheric extension, not global. Mörner (1984c) argues that when in the 1960's and 70's global correlations in temperature, glaciation, sea level, etc. were established, only a few voices were heard (e.g. Mercer, 1969; Mörner et al., 1982). Now new and conclusive evidence on the question of a global or regional extensions of climatic changes shorter than the Milankovitch variables forces the fundamental conclusion that none of the rapid climatic changes and shifts of the types evident during the last 20,000 years have had a global extension. These have been limited to a hemispherical or regional extensions, of a type where the change is counter-balanced by compensational changes in the opposite direction in other regions. This applies even to the changes at the Pleistocene/Holocene boundary (Morner, 1984c).

Taking the above into consideration, one of the main targets of the present study was to investigate the extensions of the climate variations and their impacts globally. This was done by establishing the extension of the co-variation observed in each region on an inter-regional scale, and later on a global scale. Thus, although the study was made primarily on data available from the Mediterranean region and China, it covers many other regions in the northern hemisphere or subregions within these areas. Inter-regional comparisons with areas in the Southern Hemisphere, are planned for the more advanced later stages of this project.

It should be emphasized that the time boundaries of the climatic periods designated in this report are preliminary in nature and vary from one region to another. A general problem is the different resolutions of paleoclimatic data in time, which obviously influences the recognition of the beginning or end of a warm or cold period. Toward the present more detailed paleoclimatic data are available. The difference in duration and amplitude of the warm periods designated below implies that they cannot br compared to one another on the same level. This is a methodological problem inherent in paleoclimatic research, which must be taken into consideration.

All these problems are technical in some way. There exist, of course, also the methodical problems, such as the dating scale. For example when tree rings or lake varves are involved the dating can be made on an annual scale, by counting the rings or varves. On the other hand when pollen, isotopes or soil profiles are used, the dating depends on the presence of carbonaceous materials necessary for ^{14}C dating. When sea or lake levels are used, however, one must be sure that younger and more widespread changes have not obliterated older and smaller changes.

Another methodological problem occurs when correlations between curves are based on different type of data. This is due to the fact that each of the changes on which the different curves were based have been constructed on the basis of different materials responding to climate changes in different ways. Thus, for example, when widths of tree rings are interpreted, this may bear evidence to different climate changes. For example, in warm regions it may signify change in precipitation, while in colder regions it could signify more warmth and solar radiation. Beyond this, there is the impact of human intervention which may have nothing to do with climate change. In order to establish correlations (in spite of these differences) a standard reference curve has been constructed to serve as a basis for correlation. But before this was possible the general understanding of what happened during the decisive global change at the end of the Pleistocene and the beginning of the Holocene had to be understood.

2. CLIMATE CHANGES DURING THE END OF THE PLEISTOCENE AND THE BEGINNING OF THE HOLOCENE," AND ANALYSES OF THEIR HYDROLOGICAL IMPACTS.

Following the principle of "the past is a key for the future", it is important to trace the changes, and especially the co-variations which took place at the shift from the Pleistocene to the Holocene, and learn from them on possible variations in the future. This must be done with a great deal of caution, however, for there were great variations in the magnitudes of the changes between periods of extreme cold which caused the glaciers to spread significantly, or periods of extreme warmth which caused the glaciers to melt nearly completely. In addition, one has to take into consideration the fact that changes in climate are not linear, and beyond some threshold point new altogether different situations may develop. Yet the understanding of the climatic and hydrological scenarios during global warming periods at the termination of the Pleistocene is of importance. It may help our understanding regarding the future trends of present climate systems which may be considered a warm period (even of high amplitude).

The last Glacial Maximum occurred between 22,000-16,000 BP. Soon afterwards (15,000-8,000 BP), the massive deglaciation started, characterized by strong fluctuations. A warm period seems to have occurred between 13,500-" 11,000 BP. This is referred to as the "Windermere Interstadial" in Britain and h-eland, and called the Boiling Interstadial in Scandinavia. A brief cold phase (mini-stadial) occurred within this warm period on the British Isles around 12,000 BP, equivalent to the Older Dryas in Denmark and Scandinavia. The 11,800-11,000 BP) followed the Older Dryas (Mangerud, 1987).

Pollen data from Luxembourg, France, Corsica and Spain indicate : A sudden and marked cooling (4-5°C) as well as a fall in precipitation (400-500 mm) occurred at 11,000 BP in southwestern Europe with a temperature and precipitation minimum that still persisted in 10,500 BP. At that time the temperature was similar to that of the Older Dryas, but precipitation was intermediate between that of the Older Dryas and today. But the precipitation fall was more marked in winter than in the other seasons. It is likely that the important advance of the glaciers in the Alps was made possible as a result of a medium accumulation of snow in spring and autumn and, more important, weakened melting in summer (Pens et al. 1987).

In Europe at the beginning of the interstadial, around 13,000 BP, temperatures rose abruptly and summer temperatures lay at (or even above) those of today. These palaeoclimatic deductions are based on data from fossil beetles (coleoptera) which react quicker than plants to temperature changes (Roberts, 1989). Around 13,000 BP temperate beetle assemblages replaced arctic ones and the rate of warming at a site in north Wales is calculated to have been an amazingly fast 1°C per decade (Coope, 1975). Summer temperatures at about 13,000 BP in central Britain rose suddenly by 7°C and winter temperatures may have risen by 20°C at the same time (Coope, 1987). Paleotemperatures in Britain from fossil beetle assemblages have been calibrated by Atkinson, Briffa and Coope (1987) for both surer and winter in the period 14,500-9,500 BP. Climatic reconstructions based on Coleoptera for the Holocene period have so far proved to be more problematic (Roberts, 1989).

The climate in the Levant, during the upper part of the Pleistocene, namely during most of the Last Glacial period, (and even parts of the lower Holocene) was most probably colder and more humid than the climate during the upper part of the Holocene (Issar and Bruins, 1983; Issa and Tsoar, 1987; Issar, Tsoar and Levin, 1989). The Mediterranean Sea was lower, and the paleo-Dead Sea, i.e. Lake Lisan, higher (Picard, 1943, Neev and Emmery, 1967). Stiller and Hutchinson (1980). Investigating stable isotopic composition of carbonates of a 54 m core in Lake Huleh, northern Israel, found that the ^{18}O data suggest that no very drastic climatic changes happened. From about 16,000-10,000 BP the mean temperature seems to have been slightly lower than later on, as would be expected. There is some evidence of a period warmer than today between about 6,600-4,300 BP.

Vegetational and climatic interpretation of palynological data (Van Zeist, 1980) show the main tendencies in the Late Quaternary vegetational development in the Near East:

24,000 to 14,000 BP, colder temperatures, markedly drier than today, although pollen shows fluctuations and regional differences in humidity.

14,000 to 10,000 BP, increase in temperatures. Many sites suggest a distinct rise in humidity around 14,000 BP.

Early Holocene, dryer in northern Israel (Huleh pollen record), but much expansion of tree growth in Greece.

± 4,000 B-P., the present day distribution of forest and steppe had established itself in broad outline.

The climate in the lower Jordan Valley appears to have been moderately humid in the Geometric Kebaran (14,500-12,500 BP) and Early Natufian (12,500-11,000 BP), based on pollen analysis. A similar analysis of three Natufian sites in the lower Jordan Valley, for the period of 12,500-10,300 BP (early, recent and late Natufian period) shows slight humidity in the Early Natufian period and progressive desiccation which reaches its peak during the Late Natufian period. The geomorphological and faunal data confirm the results. At present the area receives 100-200 mm annual rainfall. The slight humid phase is also found in other ancient Natufien sites:- at Hayonim terrace (niveau D) in North Israel, dated at 11,970 BP, and at Wadi Judayid (southern Jordan), dated at 12,140 BP.(Darmon, 1984, 1987, 1988).

The Dead Sea clearly shows a humid period in the terminal Pleistocene with lisan-type greenish-gray laminated clay sediments (Neev and Emery,1967). The Dead Sea (Lake Lisan) had a much greater areal extent and reached in fact its maximum level during this period from about 13,000-11,000 BP (Begin et al., 1985). In northern Sinai Goldberg (1977) found evidence for a fresh water lake which occupied a large region at Gebel Maghara during this period. Abundant arboreal pollen from this period were found in the Central Negev.

Continuous pollen diagrams for the last 3.5 million years from Israel, from boreholes penetrating the entire Quaternary sequences of the Hula and Dead Sea lakes, show that the

Dead Sea served as the lowest part of the drainage basin through this period, then ensuring a complete, continuous deposition. The glacial phases are manifested in Israel by periods of somewhat lower temperatures and higher rainfall, some in summer time. The interglacials are hot and dry, with Saharan conditions prevailing. The interstadials are of the present day character of a short, rainy winter and a long, hot, dry summer. It is possible that short dry phases would occur in Israel at peaks of the glacials, but in general the periods of recorded low sea levels had a wet climate (Horowitz, 1979).

Leroi-Gourhan (1980) investigating pollen spectra, in the Middle East, found that during Lower and Middle Würm fluctuations of wet and dry phases, as well as temperature existed. The cold-wet maximum seems to be dated around 45,000 BP. On the other hand a drought characterizes the coldest Würmian phase, which probably explains the minimal archaeological evidence of occupation between 23,000 to 19,500 BP. The late glacial shows some improvements in climate, dated to 17,000, 13,500 and 12,000 BP. A richer and more diversified flora marks the beginning of the Holocene. While the desertification of certain regions since 6,000 BP is certainly due to climate, the increasing population densities in the last 10,000 years, pastoral and agricultural practices have had such an influence on vegetation and soils that it is impossible for the Middle East to refer to a Holocene botanical stratigraphy.

Pollen analysis from archaeological sites in inner Syria provide new data about the end of the Pleistocene and the beginning of the Holocene. Flint typology and radiocarbon dates have enabled us to obtain a finer chronology for the botanical fluctuations. Wetness, beginning about 10,000 BP, seems to increase after 9,800 BP. (Leroi-Gourhan, 1974). Pollen diagrams from archaeological sites all over the Levant brought Leroi-Gourhan (1981) to the following conclusions:

± 14,000 to 13,000 BP, dry
± 12,500 t. 12,000 BP, dry
± 10,000 to 9,000 BP, wet
± 11,000 to 10,000 BP, dry
± 10,000 to 9,000 BP, wet
± 9,000 to 8,000 BP, dry

Pollen diagrams were prepared from Epi-palaeolithic and Neolithic sites in the Jordan Valley, including the regions of Fazael/Salibiya and Mallaha (Leroi-Gourhan and Darmon, 1987). The results are the following:

Kebaran (c.19,000-14,500 BP), chronological definition is not clear and it is difficult to place with assurance a slight humid period found in this interval.

Geometric Kebaran (c. 14,500-12,500 BP), another humid period is observed (possibly correlated with the Boiling).

Natufian (c.12,500-10,300 BP), next humid period in Early Natufian, 12,000 Bp in Hayonim terrace, Galilee; climate progressively drier through the end of the Natufian period.

Prepottery Neolithic (PPNA) (c.10,000-9,500 BP), wetter, marked development of trees, around 10,000 BP. Relatively forested conditions continued to be observed at Netiv Hagdud, dated between 10,250-7,900 BP.

Weinstein (1976) investigated the late Quaternary vegetation of the northern Golan, as manifested by the pollen assemblage of samples from a borehole P/8, drilled at the center of the lake of Birket Ram. This is a rather small, elliptical volcanic crater lake, 900 x 600 m, bordered by very steep slopes. Present average annual precipitation is 1,000 mm. The fluctuations seen in this section are significant and it calls for a more intensive dating effort:

From 92 to 69 m more than 60% arboreal pollen -- mainly Quercus (75-80%) and Conifers (<10%), while the Irano-Turanian assemblage is low (12-20%).

From 69 to 63 m the arboreal level is 30-40%," from which, Quercus is 65% and the Irano-Turanian is high (55-60%).

From 63 to 47.5 m the arboreal level is 71%, with Quercus 57-81% and Conifers 8-23%; Irano-Turanian up to 61%.

From 47,5 to 39 m the arboreal is 30%, with Olea 17%, and the Irano-Turanian low at 20-25%.

From 39 to 30 m (36 m ^{14}C age 28,400 $\pm$ 3,000 BP) the arboreal is 80%, with Conifers 88%.

From 30 to 22.5 m the arboreal is 33%, with Quercus 75-80% and Irano-Turanian >40%.

From 22,5 to 11,5 m the arboreal is 60% with Quercus 89% and the Irano-Turonian 20%.

In the southern part of Israel, layers of loess were deposited during most of the Last Glacial period. The uppermost stratigraphic paleosol with calcic horizon (Bruins, 1976; Bruins and Yaalon (1979) found widespread in the northern Negev, was formed around 13,000 BP, according to radiocarbon dating by Goodfriend and Magaritz (1988).

A support to the suggestion that the warming at the end of the Pleistocene was accompanied by the strengthening of the monsoon one finds in North Africa, many lake levels rose after 13,000 BP only to fall again after 11,000 BP (Street-Perrott and Roberts, 1983; Roberts, 1989).

A study by Klein et al. (1990) of fossil and modem *Porites* corals from reef terraces in the south-eastern Sinai, along the Red Sea, indicates that during the Late Quaternary high sea levels and a wetter climate prevailed in Sinai, possibly with an increased summer rainfall regime. Most fossil corals show degrees of fluorescent banding after irradiation with long-wave ultraviolet light, whereas living *Porites* corals do not exhibit distinct fluorescent banding. The source of fluorescence is humic acid of terrestrial origin, as already found in corals from the Great Barrier Reef of Australia (Isdale, 1984). The distinct fluorescent banding in the fossil Sinai corals is understood to be a function of periodic terrestrial floods during the life of the corals, irrespective of later events (Klein et al.,1990). Modem corals show skeletal banding patterns, low-density bands being deposited in summer and narrow high-density bands in winter. Fossil corals have a similar density-banding pattern. An important finding is that the fluorescent bands, related to humic acid from terrestrial floods, are superimposed on the low-density portions of the skeleton bands, which implies summer rainfall (Klein et al.,1990).

The Nile had been a highly seasonal, braided river between 20,000 and 12,500 BP. At the end of this period, from 14,500-12,500 BP, the amount of water flowing through the Nile began to increase and several high substages can be recognized, as well as moderate activity in the wadis (Butzer, 1980). Around 12,500 BP, the Nile experienced a huge increase in discharge, partly on account of overflowing East African lakes expanding the river's catchment area. The overflow from Lake Victoria and other lakes, as well as higher rainfall in Ethiopia sent extraordinary floods down the main Nile. This marked a revolutionary change from ephemeral to continuous flow with superimposed flood peaks. As a result, the main Nile and its tributaries formed more stable channels of higher sinuosity, from which suspended Ethiopian silt and clay was deposited on the floodplains (Adamson et al., 1980).

The "wild" Nile regime, characterized by floodplain sedimentation, lasted until around 11,500 BP and was followed by a period of strong dissection and down cutting in the Nile and wadis of at least 20 meter (Butzer, 1980).

In the Chad basin the climate was humid from about 12,500-11,000 BP (Servant et al., 1976), after a long dry period in the Sahel (20,000-12,500 BP) characterized by major dune building (Talbot, 1980). Rain in the Sahel during this period was very seasonal, perhaps a strict monsoon regime (Rognon, 1987).

Sonntag et al. (1980) investigating the environmental isotopes of the fossil groundwater under the Sahara, found that this water was mainly recharged during the long humid period between 50,000-20,000 BP (Great Pluvial), while from 20,000-14,000 BP there was a long dry period.

Preliminary pollen data from sediments from Lake Bosumtwi in Southern Ghana (West Africa) dated between about 15,000 and 8,500 B. P., show the presence of Olea hochstetteri pollen with percentages between 3-8%. The spread of this typical mountain species is limited to low altitudes. This phenomenon may have been caused by a heavy cloud cover lasting most of the year (low clouds of stratiform type and fog); it coincided with a lowering of the mean temperature of at least 2 to 3°C (Maley and Livingstone, 1983).

A focus was put on a fossil sequence recently obtained in the mountains of the Zaire/Nile divide in Burundi, which emphasizes the region south of the Equator (Bonnefille et al., 1990). It shows the first continuous palaeotemperature record for the tropics. During the past few thousands years the average of the reconstructured temperate curve is close to the modern values. The over-estimation of the top-most sample is due to the recent deforestation around the site. The record indicates an average temperature decrease of 4 ± 2°C for the glacial period between 30,000 and 13,000 BP. Moreover, the curve seems to be consistent with the evolution of past vegetation indicated by the same fossil pollen record. Before 30,000 BP the cooling is consistent with the occurrence of mountain conifer forests in the Burundi highlands (>2,200 m), which are now tropical rain forest. It also agrees with other pollen data from East Africa. The lowest temperate is recorded between 25,000 and 15,000 BP, According to Thunell and Williams (1983, 1989), who investigated the paleotemperature and paleosalinity history of the Eastern Mediterranean during the Late Quaternary, the Mediterranean isotopic signal is a complex record of regional temperature and salinity changes superimposed on the compositional change due to the global ice volume effect. Hydrographic conditions in the Mediterranean at 8,000 BP must have been considerably different than at 18,000 BP and than today. The water balance became positive as precipitation and run-off exceeded evaporation. Salinities were considerably lower at 8,000 BP, and the west-east (increasing) salinity gradient was reversed to an east-west gradient. This is supported by east African climate records which indicate the onset of very humid conditions at approximately 12,500 BP, with wettest conditions occurring between 10,000-8,000 BP. This was also a time of intensified African monsoons and increased Nile discharge.

Deposition of the most recent sapropel (9,000-7,000 BP) on the eastern Mediterranean Sea floor seems related to a circulation reversal of water (flowing out of the Mediterranean to the Atlantic during this period), which would have caused the eastern Mediterranean to become a nutrient trap, thus enhancing productivity. According to Herman (1989) the record of past environments is more complete and the revolution of climatic events is higher in the Mediterranean than in the major ocean basins. However, the overprint of tectonic activity is widespread (ash layers, turbidites and slump deposits). During glacial temperature minima, surface water temperatures were about 3°C lower in summer and about 3-4°C lower in winter. Salinities were highest during glacial maxima when climates were more arid than today - sea level was very low (-130 to -140 m), the discharge of the Nile was greatly reduced, and the connection between the Mediterranean and Black Sea (Bosporus sill at -36 m), which is a major supplier of low salinity water, was reversed.

A well dated core from the Atlantic Ocean, about 200 km off the African coast near Mauritania (core 13289, Koopmann, 1981), revealed a marked and sudden decrease of Trade wind speed around 14,900 BP, as indicated by grain size data from aeolian-marine dust deposits according to the equation of Sarntheim et al. (1981). This sudden reduction in speed of the Trade Winds above North Africa is remarkably synchronous with the onset of global deglaciation and the decrease of plankton primary productivity in core 12392 (Sarnthein et al;, 1987).

Deep sea sediment cores from the Atlantic show that the polar front shifted as far north as Iceland during the interstadial of 13,000-11,000 BP (Ruddiman and McIntyre, 1981; Roberts, 1989).

The impact of a warm peak around 11,600 BP, on the hydrological cycle in North America can be learned from the fact that meltwater flooding down the Mississippi reduced surface seawater salinities in the Gulf of Mexico by 10 per cent (Emiliani, 1980; Roberts, 1989).

As paleoclimatic records adjacent to India and Africa show that monsoon maxima occured during interglacial conditions (and coincide with precession maxima and maxima of northern hemisphere summer radiation), Prell and Kutzbach (1987) have used a community climate model (CCM) in order to identify the processes causing changes in monsoon circulation. The model indicates that "...the regional monsoonal response to glacial age boundary conditions is variable. At 18,000 BP the monsoon is greatly weakened in southern Asia but precipitation increased in the Western Indian Ocean and in equatorial North Africa' '..."Both areas show stronger monsoon with increasing solar radiation during interglacial conditions. "

The climate in eastern Asia, especially in China, during the transition period from the Pleistocene to the Holocene was different from one region to another, according to the climatic belt to which each region belonged. While the most north western region belongs to the westerlies belt, the rest of China is influenced by the monsoonal regime. Thus, *while the* first region was cold and humid during the Last Glacial period and became warmer and drier as the glaciers melted, most other regions, especially the inner ones, were dry and cold during the ice age and warm and moist as deglaciation proceeded (Li, 1990). This was manifestated by the changes in the character of the different deposits. Chaiwopu lake in the Xinjiang region, for example, which is dominated by the westerlies, was 25-28 m higher than its present level, with grey lacustrine deposits of c.15,000 to c.12,000 BP (Jingtai and Kaqin, 1989). On the other hand, the Qinghai and Qaidan lakes in the monsoonal zone nearly or entirely dried Up during the Last Glacial Maximum (Ponyxi et al., 1988; Kezau and Bowler, 1985). Yet when maximum interglacial conditions prevailed, monsoonal influence could have extended also to north eastern China, usually under the westerlies influence (Li, 1990).

The climate changes during the Pleistocene on the whole, as well as that of the Pleistocene-Holocene transition period, is well identified by the interchange between the loess and paleosoils layers. In the loess plateau of China influenced by the monsoon, the dry and cold glacial periods are characterized by the deposition of aeolian loess. During the warm and wet interglacial periods, however, paleosoils were formed. In the most northerly region, influenced by the westerlies, during the glacial periods brown and black soils were formed, while during the interglacial periods, mainly aeolian loess was deposited.

This general pattern changes according to the specific local character of the different regions. Thus in the Yulin area, in the northern part of the monsoonal loess plateau, which *is* on the margin of the desert, during the glacial cold dry period, aeolian sand was deposited instead of loess (Guarong, 1988).

Climate changes in western China, based on palynological analysis, has been derived from investigating the sediments of the mountainous Lake Barkol in eastern Xingjiang (Han and Yuan, 1990). The sediments were found to be continuous from the upper Pleistocene to recent. This permitted an understanding of the role of the uplifting of the Tibetan Plateau in deciding the difference between the climate of western China and that of eastern China. This uplift, which reached its present elevation of 4000 m at the upper Pleistocene, has cut the influence of the east-western monsoon. This resulted in changing the pattern of the cold-dry and warm-wet regimes to cold-wet and warm-dry. The first regime, which was consistent and which characterizes even today most of China, prevailed only during the lower and middle

Pleistocene. The uplift also caused the aridization of this region, especially during the interglacial periods. During glacial periods, effective precipitation increased, resulting in lake level rises. During such cycles low temperatures and extended icebound season inhibited evaporation, and more moisture caused the forest belt to expand. The climate changes during the upper-most Pleistocene shows that most of the period from 21,000 to 12,000 BP was cold and wet, except for three cool and wet periods at 20,000, 16,000 and 13,000 BP. During these periods a retreat of the glaciers occurred. At 11,000 BP there was a warm dry spell, but at 10,000 BP it became warm and dry again.

An et al. (1991a) discuss the history of the variations of the East Asia Monsoon using two time intervals, the last 130,000 years and the last 18,000 years, on the basis of the analysis of the loess paleosol sequence in China. In the context of the present report the last, and shortest, time span will be discussed. The accelerated accumulation of thick loess layers from 18,000 to 14,000 BP, indicates the domination of polar continental air mass (PCAM) or winter monsoon and the decline of the influence of the tropical-subtropical oceanic air mass (TOAM) or summer monsoon. The small amount of pollen, dominated by pollen of arid resistant herbs, records the decrease of rainfall on a large scale. Around 12,000 BP a paleosol developed in the western loess plateau. There was a significant increase of pollen of broad leaf trees, such as spruce and fir, and the rising of lake levels. These are evidence of the strengthening of the summer monsoon. During the transition interval from Pleistocene to Holocene, i.e. around 11,000 BP, there was an accelerated accumulation of dust, an increase of herb pollen, a disappearance of the spruce and fir, and the rapid lowering of lake levels, which indicates a short return of the dominance of the PCAM. This may have lasted for several hundred years. Later on the climate changed quickly. From 9,000 to 5,000 BP a paleosol developed on the loess plateau, containing an increased pollen count of broad leaf deciduous trees.

An et al. (1991b) also summarize the paleoenvironmental changes of China during the last 18,000 years as evidenced by dust accumulation, vegetational evolution, mountain glacier's advance and retreat and sea level changes. They use the magnetic susceptibility of the sequence of loess as proxy data for climate, which is related to the precipitation and vegetation cover ratio (Kukla and An, 1989). From their composite curve, they conclude that the last full glacial age began to decline at about 14,500 BP, and that there was a cool and humid ripple about 12,000 BP. Then it rapidly changed to cold and dry around 11,000 BP. After this the climate changed rapidly towards a Holocene warm climate optimum spanning from 9,000 to 5,000 BP. From 5,000 BP to the present the climate changed towards cool and dry, showing some ripples of neo-glaciation. The same conclusions are reached by Zhou and An (1991) who, in their article on ^{14}C chronology of the loess plateau in China, report on the correlation of ^{14}C chronozones and indication of climatic change history during the Upper Pleistocene and Holocene. They also consider paleosol layers which have high magnetic susceptibility as a record of low aeolian accumulation and strong biological accumulation, which reflects a warm and humid climate. On the other hand, low susceptibility should indicate rapid dust accumulation and weak pedogenetic processes typical of dry-cold or dry cool climates. They use also the magnetic susceptibility age conversion equation developed by Kukla and An

(1989). From 13,000 to 12,000 BP evidence for a mild climate is found in the form of a carbonate nodule horizon. From 11,000-10,000 BP there was a cold and dry climate, while a thick black paleosol was formed between 1(),()00 and 5,000 BP indicating a warm humid climate. Since 5,000 BP the loess plateau has been dominated by the deposition of recent loess, generally reflecting a dry and cold climate. Some intercalations of weakly pedogenetic paleosoils reflect periods of milder climate.

Wang & Li (1991) discuss the temporal and spatial distribution of lake sediments in China during the Late Cenozoic and the climatic environments revealed by lacustrine sediments since the Last Glacial period. They have found an impressive difference in the evolutionary features of the lakes in different areas since the last glaciation. The Qinghai and Daihai lakes lying on the northwestern margin of the east Asia monsoon area had low lake levels during last glaciation and the sediments were coarse. The palynological assemblage from the Daihai Lake indicates that there was an Artemisia steppe scattered with Ephedra shrub thickets and with a few conifer or Abies-Picea-pinus trees. The level of the lake rose from 10,000 to 4,500 BP, building a terrace 40-45 m above present level.

Li (1990), investigated the fluctuations of closed lake levels, and variations of paleo-climatology in the North Tibetan Plateau, China. His conclusions are that since 18,000 BP to present the lakes in North Tibet went through four stages of fluctuations:

Stage 1. From 18,000 to 12,000 BP, levels of the lakes were rising, while mountain glaciers advanced. This is possibly related to the pattern of atmospheric circulation, i.e the belt of westerlies was shifted over the region and precipitation increased.

Stage 2. From 12,000 to 8,000 BP, the level of the lakes dropped.

Stage 3. From 8,000 to 6,000 BP, rising of lake levels.

Stage 4. From 6,000 BP to present, dropping of lake levels.

Xue et al. (1991), summarizing the history of the Holocene coastal sedimentation in China, reports a transgression with sea depth of 60 m from 11,000 BP, reaching its climax at 6,000 BP. Since then the coasts prograded seaward at various rates.

Ice cores from the Tibetan Plateau of China provide a detailed record of the climatic conditions in the period from the Upper Pleistocene to the Holocene (Thompson et al., 1989). "It reveals that the late glacial stage conditions were apparently colder, wetter and dustier than Holocene conditions. The late glacial stage part of the cores is characterized by more negative d^{18}O ratios, increased dust contents, decreased soluble aerosol concentrations, and reduced ice crystal sizes than the Holocene part. These changes occurred c. 10,000 years ago. The d^{18}O record reveals a short interval of apparent warming about 35,000 years ago that is also associated with less dust deposition. Full glacial conditions were soon reestablished. Between 30,000 and 10,000 years ago, concentrations of Cl and SO_4 increased gradually. One explanation for this increase is that conditions became drier and areal extent of salt and loess deposits increased. At 10,750 years ago, a marked increase occurred, probably reflecting the drying of the fresh water lakes".

From a core in the north-west Pacific ocean, a tie between the loess sequence of China to the $d^{18}O$ chronostratigrahic and paleoclimatic sequence could be established (Hovan et al., 1989). Since the formation of the mid latitude loess occurs during the glacial periods of the Plio-Pleistocene epochs (Pye, 1987), a correlation was observed between the deposition of loess layers in China and greater accumulation of aeolian material in the deep sea. From the correlation diagram presented by Hovan et al. (1989) the melting of the ice period of the end of the Last Glacial, which is characterized by the water of the oceans becoming lighter in their $d^{18}O$ contents and by a reduction in the aeolian flux.

3. CONSTRUCTION OF A STANDARD DIAGRAM OF CLIMATE VARIATIONS FOR THE LEVANT, AND THE ANALYSIS OF CLIMATE CHANGE IMPACTS

The reason for choosing the Levant, i.e. countries bordering the eastern Mediterranean, as the region for which a standard diagram was constructed was due, in the first place, to the fact that a preliminary survey had shown that some major climate changes had affected this region during the Holocene, and are well pronounced in profiles of various paleo-environments. This is most probably connected with the fact that this region serves as an intermediate zone between the humid and arid belts on the global scale, and as a transitional area between the Mediterranean Sea affected lands and the deserts of Egypt, Syria and Arabia, on the local scale.

Another reason was the substantial archaeological and long historical records available for the region, from which could be deduced the impact of past changes of climate on the hydrological cycle as well as on the socio- economic systems.

The climate of the Levant is affected by both the subtropical high pressure belt, mainly during the summer, and by mid-latitudinal depressions during the winter. Summers are hot and rainless and winters are wet and cool. The level of precipitation decreases as one travels in southerly and easterly directions (reaching 400-500 mm on the western coastal area, 1,000-1,500 mm in the mountains of northwest Lebanon, and 50 mm in the desert areas). Average air temperatures increase in a directional pattern, similar to that of regional precipitation. (In northern Syria, the average January temperature is 5°C and in August, it is 24°C; in Beirut, 13°C in January, and 27°C in August).

During winter, low pressure systems which form over the western parts of the Mediterranean and Southern Europe penetrate the region and produce rainfall. These low pressure systems are often followed by development of high pressure systems which cause clear and cold weather conditions.

During summer, the weather is less variable, being affected by the semi-permanent surface heat trough centered over Iran and Iraq. This surface trough is coupled with an upper air high pressure system producing stable hot and dry weather.

During autumn, cool and moist air masses occasionally penetrate the Levant from the north and upon passing over the hot ground produce rainfall. Spring is characterized by frequent occurrences of *khamsins* and dust storms, caused by the penetration of heat lows from North Africa, although some rainfall may occur.

overall, six main air masses affect the weather over the Levant, originating over the following areas:

1) The Arctic Ocean;
2) The Atlantic Ocean, south and west of Iceland;
3) Northern Russia and Siberia;
4) Northern Russia, being modified upon passing over the Volga-Ural basins;
5) The Atlantic Ocean south of the Azores;
6) The North African desert, or over Iran and Iraq.

The first four air masses originate at high latitudes and are characterized by low temperatures and dryness, acquiring moisture upon passing over the Mediterranean. The last two air masses originate at low latitudes, and are characterized by high temperatures.

Being located on the eastern coast of the Mediterranean, the Levant's climate is affected by the sea-land breeze. During the summer, the wind is most developed, being enhanced by the synoptic scale circulation. The quasi-permanent wind has a northwesterly component which is from the same quadrant as the day sea breeze but opposite to the nocturnal easterly land breeze regime. Thus, the nocturnal wind component is weaker. During the winter, penetration of cold fronts and low pressure systems often obliterate or reduce the effects of the local sea-land breeze. On such days, the winds are usually from the westerly sector and are moderate to strong. On winter days in which high pressure prevails, the sea-land breeze may again be more pronounced.

A synoptic analysis of excessive rainfalls in Israel (Amiran and Gilead, 1954) shows that they are the result of an influx of deep, moist and cold polar air into the eastern Mediterranean along meridional trajectories, which establishes contact with the warm surface air in a Cyprus Low. With the build-up of the Siberian anticyclone as winter progresses, such a situation becomes less probable and there is less chance of the formation of a strong jet stream over central Europe and the Mediterranean which would feed sufficient air into such a circulation system. The excessive rains are therefore restricted to the beginning of the season, i.e. November or December. The rains lashing down on soils when vegetation is still scarce and weak cause severe soil erosion. Crown (1972) shows that rainfall in the Middle East, on the whole has an inverse correlation with temperature except where stations have a double rainfall regime (e.g., Tunis) or are under the influence of the summer rainfall regime (e.g., Khartoum). According to Otterman (1974) aridity can arise in the Levant from three general causes:

1) Separation of the region from oceanic moisture sources by distance or topography;
2) The existence of dry stable air masses that resist convective currents; and

3) The absence of influences that cause convergence, create unstable environments and provide the lifting of air necessary for precipitation.

Zangvil (1979) investigated the temporal fluctuations of seasonal precipitation in Jerusalem during the period 1846/47 to 1953/54, by the method of time spectrum analysis and filtering techniques. A prominent peak appears in the spectrum at a period of 3.0-3.3 years. (Rainfall oscillations in California also show a peak around 3 years). In the spectra of most of the East African stations, the most prominent peak occurs at 3.3 years. As more than average rainfall in East Africa during the main rain period of January to April is probably associated with a more intense Hadley circulation, this circulation is causing strong westerlies in the same longitude resulting in reduced rainfall in the Eastern Mediterranean. Thus perhaps a connection exists between the southern oscillation and the rainfall in Jerusalem The southern oscillation is a world wide phenomenon, having a dominant period of 3 to 6 years which corresponds, for example, with the first (3.0-3.3 yr) peak and the secondary one (5 yr) in Jerusalem rainfall oscillations.

The base section of the middle and upper part of the Holocene (Fig. 1) was prepared by Issar after he had found a good correlation between climatic changes and the desertification of the Negev desert (Issar and Tsoar, 1987; Issar et al., 1989; Issar, 1990a; Issar et al., 1992). This was done by re-interpreting the $d^{18}O$ and $d^{13}C$ time-series obtained from lacustrine carbonates in a core taken in the Sea of Galilee and the corresponding palynological results. Initially, these data were interpreted as reflections of anthropogenic factors acting during historical times (Stiller et al., 1984; Baruch, 1986). Additional evidence was derived from $d^{18}O$ and $d^{13}C$ time-series evaluated from results of 41 stalagmites taken in 10 caves in Galilee, northern Israel. By these it was proved that the growing rates of the individual stalagmites were rather constant during the whole formation periods as was the initial ^{14}C content which is the presupposition for the absolute age determination by the ^{14}C method. The temporal calibration of the chronostratigraphic stable-isotope record was done with ^{14}C and U/Th dates only the upper middle part of the Holocene are discussed). A precision of around ±300 years have been obtained taking a reservoir effect of -900 years into account (Geyh et al., manuscript). Thus one has to regard the speleothem ages as calibrated dates with &300 years margin of error. That these dates, within this margin of error, are reliable, can be deduced from the close similarity between the two isotope curves, the speleothem curve, and the one from the Sea of Galilee. on the other hand as one is dealing with isotopes in totally different environments, an anthropogenic common reason is difficult to accept.

Archaeological data were derived from a detailed study of a representative area of 100 km^2 carried out in the framework of the general archaeological survey of the plain of Beer-Sheva and Arad (Govrin, 1992) (see Fig. 1).

These data were complemented by the results from studies on the paleo levels of the Mediterranean sea (Bloch, 1976; Raban, 1987) and Dead Sea lake levels (Klein, 1961, 1981, 1985; Frumkin et al., 1991) (see Figs. 2,3 and 4).

The lower-most part of the curve from 10,000 to 6,000 BP, which spreads mainly on the neolithic period, was constructed on the basis of the interpretation of archaeological findings in the Negev and Sinai (Bar Yosef, 1980), the submarine findings along the Mediterranean shores of Israel (Fig. 4) (Raban, 1987; Galili et al., 1988), and the Archaeology of Jericho (Kenyon, 1957).

Thus the base section is based on the co-variations observed by many investigations carried out independently by different research groups, in different fields and different environments, but responding in a coherent way. These are spread over the transition zone between desert and Mediterranean climate. Certain changes which may be marginal in more humid areas, from the socio-economic point of view, may be crucial in the more arid parts.

The interpretation of the stable isotope data as humidity indicators follows the basic idea that the ^{18}O values of precipitation are interrelated with the temperature (Geyh and Franke, 1970). But it is certainly an over-simplification, and other meteorologic factors such as changes in the storm trajectories, in the seasonal distribution of the precipitation, and in the humidity may also pay a role (Gat, 1981; Leguy et al., 1983). This simplification is, however, thought to be justified owing to the obvious interrelation among the isotope data, and between these, the archaeological and sea level data. One also has to take into consideration small discrepancies which may be due to the time scales used in the different data sources. These apply to the different amplitudes of the $d^{18}O$ records of the lake sediments and of the speleothems previously mentioned. It is known, for example, that the water budget of the Sea of Galilee is also determined by an inflow of groundwater from the flanks of the rift valley. Spring water collected along the shore yielded ^{14}C dates of more than 10,000 years. This damps the corresponding isotope variations. In contrast, changes in $d^{18}O$ values of speleothems reflect the fluctuations of the isotope composition of the meteoric water over decades. The 1 mm thickness analyzed represented age ranges of about 10 years. Yet, the overwhelming observation is that the main trends in the two curves from the caves and Sea of Galilee, fit well. And these correspond well with the settlement distribution in the Negev and levels of the Mediterranean and Dead Sea. The best explanation to the opinion of the present author is that this correspondence reflects the impact of climate change.

The general conclusions which can be derived from the study of the master curve is that the earliest humid climate phase happened during the pre-pottery Neolithic period (9,S00-7,500 BP) signified by settlements in the Sinai and Negev deserts, and Jericho. As there is no conclusive isotope evidence for this period, it is not yet clear whether the higher precipitation were a result of the migration of the monsoon to the north or the cyclonic belt to the south. The correlation with the warm climate periods of Europe may point to the fact that the first alternative was the case.

A marked depletion of the carbon and oxygen isotope compositions is observed during the Chalcolithic, i.e. around 6,000 BP and ended at around 5,200 BP, namely from the mid to the end of the Chalcolithic period. Other investigations, dealing with the climate changes during the Holocene, point t. the fact that during the Chalcolithic period the climate was more humid, and most probably colder, than the present. (Neev and Emmery, 1967; Horowitz, 1974, Bruins, 1976). During this period 12 settlements existed in the region, located in caves along the foothills and in proximity of riverbeds along the plain. The most famous of these is Beer-Sheva (Perrot, 1972), which most probably received its water supply from shallow wells located in the riverbed. It is possible that the river flowed during most of the year. There are many indications that the people harvested fields along the riverbeds. Diversion dams were also used in order to bring water from the river to the fields (Alon, 1988).

Around 5,200 BP, towards the end of the Chalcolithic period and at the beginning of the early Bronze age, a trend of warmer climate may be deduced from the isotopes data. At the same time, a major change in the population throughout the plain area was observed. The Chalcolithic, an agricultural population with high technological skills to produce ceramics and to irrigate was replaced by a most probably semi-nomadic population, which apparently came from the south and had very rudimentary skills for the production of ceramics (Govrin, 1992).

A relatively cold period may be derived from the depletion of the isotopic composition starting around 4,600 BP and extending to around 4,200 BP. During this period the semi-nomadic people settled and even built the fortified city of Arad (Amiran, 1978), as well as 10 other settlements. The existence of a large and flourishing urban settlement in Arad, which served as a focal point for a system of small settlements, is according to Amiran et.al. (1980) the best evidence, that the climate, at that period in the Negev, was more humid than the present climate. Alluvial deposits in river beds north of Beer-Sheva indicate a moist climate during the Chalcolithic period and Early Bronze age, while the deposits of the Middle Bronze age show a sudden increase in alleviation which may represent rapidly fluctuating rainfall patterns intercepted with drought (Rosen, 1986).

A warmer period started around 4,200 BP (beginning of Middle Bronze I age) according to the isotope curve from the Sea of Galilee. The speleothem curve reflects this event somewhat earlier (see Fig. 2). However, this may be due to a bias in the calibrated radiocarbon time scale or insufficient correlation for the reservoir effect. Yet, the signature of a warmer phase is clearly seen on all curves and is also evidenced by a total desertion of all the settlements in this region. The city of Arad was deserted at the end of the Early Bronze II. During the Middle Bronze I age the area is practically void of settlements. The survival of many large settlements during the Middle Bronze I period, more to the south in the highlands of the central Negev, despite desiccation, is explained by the fact, that these were inhabited by local populations. These were established during an earlier period, than the settlements in the northern part of the Negev, i.e. during the relatively better climatic conditions of Early Bronze II age. Ceramic evidence regarding the earlier age of these settlements, is presented by Cohen (1986). The settlement of the Negev continued all through the Late Bronze age. During the Middle Bronze II only two camps, most probably military ones, have been found.

Supporting evidence for this severe climate change was found in the Coastal Plain of Israel. An archaeological investigation at Tel-Aviv (Ritter-Kaplan,1984) located black clays deposited during the Early Bronze period. The black clays contain a pollen assemblage showing characteristics of a humid climate. This climate terminated at the beginning of

Middle Bronze. Sands without pollen were deposited later. A marine archaeological investigation (Raban,1987) along the coast he showed that at the same time sand caused the inlets of the river to be silted up. This probably means that there was an increase in the supply of sand from the Nile. The connection between the warming of climate and sand deposition will be explained later when the formation of the sand dunes on the coastal plain is discussed. A very low level of the Dead Sea (Frumkin et. al., 1991) shows that this drying up was severe. A rise in the level of the Mediterranean Sea (Bloch, 1976; Raban, 1987), show that it was global. Crown (1972) claims that the driest period of the Holocene was from 4,500 to 3,500 BP (see Figs. 3 and 4).

A new cold phase, pronounced mainly by isotope depletion started around 3,200 BP (the beginning of the Iron age). At more or less the same time, the reestablishment of settlements in the Negev began. All together 10 settlements were found. At this time the Negev was under the rule of the Kings of Judea who supported, according to the Bible, settlements in the desert (Chronicles, 8:17, 20:36, 26:10). During this period, Beer-Sheva was resettled and a citadel was built at Arad.

Around 2,700 BP the isotope composition became heavier. This may speak for a new aridization phase, but can not be correlated with the number of settlements. A reduction is only obvious around 2,500 BP. At that time, the first kingdom of Judea terminated followed by a partial abandonment of the Negev. It is not clear whether this was due to the onslaught of conquering armies from Mesopotamia, or to aridization. Shortly afterwards, a trend towards the depletion of heavy isotopes took place, most probably a result of a colder more humid period. At that time the Negev was part of the Persian empire which extended to Egypt. Four settlements have been found in the region dating from the Persian period, similar to the Hellenistic period with six settlements.

During the Roman period there were 16 settlements on the plain of Beer-Sheva. The period of heavy isotope depletion continued (with a short interval sometime around 1,700 BP) throughout the Byzantine period. There was also a rise in the level of the Dead Sea and a regression of the level of the Mediterranean. During this period sixty settlements thrived on the plain. Beer-Sheva itself was a big city extending over an area of 100 hectares.

Around 1,400 BP, toward the end of the Byzantine period, the isotopic composition becomes heavier. Parallel to this, one can observe a retreat in the level of the Dead Sea and a rise of the level of the Mediterranean Sea. Archaeological surveys have shown that a gradual abandonment of the settlements occurred along the peripheries of the desert occurred. During the first half of the 7th century AD, the Arab conquest of this area took place. During the early part of the Arab period about 15 settlements survived. They were abandoned around 1,200 BP and, until quite recently, no sedentary settlements existed. The city of Beer-Sheva was totally deserted. From around 1,200 BP, an invasion of sand dunes to the coastal plain began. This was a result of the higher rate of sediments supplied by the Nile due to increased rainfall on the Ethiopian highlands, as explained by Issar (1989, 1990b). It conforms with the climatic model suggested by Nicholson and Flohn (1980).

Based mainly on the isotope evidence, a cold short period can be observed around 1,000 BP, which may correspond to the crusaders period. The low peak above this may indicate the Little Ice Age.

A warming trend can be observed during the last two centuries.

4. CLIMATE CHANGES DURING THE HOLOCENE AND THE ANALYSIS OF THEIR HYDROLOGICAL IMPACTS

4.1 The Circum-Mediterranean Region

As mentioned, the Mediterranean climate with its cold wet winters and warm dry summers is a result of the seasonal shifting of maritime polar air masses (during the winter) with cyclonic perturbations and precipitation, alternating with maritime tropical air masses (during the summer), and excessive drought. Regions exhibiting this weather pattern include in addition to the Levant also:

- Anatolia
- The Balkan peninsula
- The southern Italian peninsula
- The southern Iberian peninsula
- The Atlas region.

4.1.1 The Levant

In the Levant the Dead Sea sediments clearly indicate a humid period in the Early Holocene, as reflected by lisan-type greenish-gray laminated clay deposits (Neev and Emery, 1967).

Wet conditions for the period 10,000-8,000 BP in the Negev are suggested by the very small size of fossil land snail shells *(Trochoidea seetzeni)*, being about 20-2590 smaller in diameter than modem ones (Goodfriend, 1986). The analysis of ^{13}C of organic matter in Early Holocene land snails from the northern Negev gives paleo-geographic evidence of earlier distributions of xeric C4 semi-shrubs (enriched in ^{13}C relative to C3 plants). This suggests an approximate doubling of rainfall in the northern Negev during the Early Holocene as compared to the present (Magaritz and Goodfriend, 1987). Faunal remains from the site of Gilgal (Pre-pottery Neolithic A) in the lower Jordan valley (Tchernov, 1980) and from Wadi Tbeik (Pre-pottery Neolithic B) in southern Sinai (Tchernov and Bar-Yosef, 1982) suggest the presence of fresh water resources at these places in the Early Holocene periods (Magaritz and Goodfriend, 1987). A high level of the Dead Sea apparently occured around 6,500 BP (Goodfriend et al., 1986).

Widespread distribution of the so-called lower salt tongue in the Dead Sea sediments above the clay deposits indicate that the driest period during the past 10,000 years occurred between about 6,500 BP and 5,500 BP (Neev and Emery, 1967). However, dating of the lower salt tongue is not precise, and is in fact inferred from its stratigraphic position in between two radiocarbon dates: 1. Dating of disseminated organic carbon within the laminated clay sediments below the salt tongue (9,850 ±150 BP); 2. Dating of disseminated organic carbon within another laminated clay bed above the salt tongue (4,410 ±320 BP). Taking into consideration the previously noted high level of the Dead Sea at about 6,500 BP (Goodfriend et al., 1986), the beginning of the lower salt tongue phase and inferred dry period probably started later than 6,500 BP.

Dead Sea sediments indicate a humid phase, by lisan-type laminated clay deposits, which has been approximately dated by Neev and Emery (1967) to about 5,500-4,300 BP. A radiocarbon date of 4,410 ±320 BP from disseminated organic carbon found in the top of this clay bed, near the base of the uppermost salt tongue, has been taken to infer the end of this humid period. The dating is not very precise."The great thickness of the uppermost salt tongue indicates that halite has been deposited continuously in the South Basin from about 4,300 BP until a few hundred years ago." (Neev and Emery, 1967).

Goodfriend (1988, 1990) investigated the organic matter of land snail shells in the northern Negev.The study (Goodfriend, 1990) is based on 40 samples of snail shells *(Trochoidea seetzeni),* whose radiocarbon ages fall within the period 6,700-2,800 BP. Goodfriend (1990) concludes that the rainfall isohyets in the northern Negev during this period must have been located about 20 km south of their present positions, which implies roughly a twofold increase in mean annual precipitation in the northern Negev (between the present-day 150 and 300 mm isohyets) relative to the present. Within this period from about 6,700 to 2,800 BP, the last part (3,900-2,800" BP) may have been somewhat drier, since $d^{13}C$ values of organic matter of land snail shells appear to indicate a somewhat lower C3 plant presence in the southern part of the northern Negev (according to Figs. 2, 3 and 4 of Goodfriend, 1990). The range of years indicated by Goodfriend (1990) during which the Negev was wet (6,700-2,800 BP), includes (in the opinion of the author of the present report two wet periods, namely the Chalcolithic and Lower Bronze, the short dry spell in between them and the dry spell of the Middle Bronze from around 4,300 to 1,200 BP. Goodfriend (1988) discusses his results in relation to other paleoclimatic data from the region, which seem to indicate a level of humidity not much above the present regime. Horowitz (1979), however, shows a peak of relatively high arboreal pollen (mostly quercus) from the Hula Basin for this period. Tsukada (cited in:Van Zeist and Bottema, 1982) shows variable but generally low arboreal pollen for this period.

According to the data from Israel, as expressed in the isotope composition, and Dead Sea levels following the dry period of the upper Bronze, a colder (and more humid in the Levant) climate phase started at ca 3400 BP, i.e. the beginning of the Iron Period. This may correspond to a weaker monsoon expressed in the recorded low flood levels of the Nile which occurred from 3,180-3,130 BP (Butzer, 1980).

The same data from Israel show a warm period around 2,500 BP during the end of the period that the Persian Empire dominated the Levant.

From about 2,300 to 1,700 BP a global cold period occurred. In the opinion of Issar (1990a) this was the early stage of "a mini-glacial period which caused an advance of the ice in the polar regions and high mountains, a regression of the sea, a colder climate in northern Europe, and a more humid climate in the deserts of the Fertile Crescent. The regions benefiting from the monsoons, such as the Ethiopian highlands, the Sahel, southern Arabia and maybe even India, most probably suffered ii-em drought, as indicated by the rather low levels of the Nile. The more benign climate in the desert of Judea and Edom promoted the flourishing of the Nabatean kingdom which was established in the first century B.C. in the Negev by nomad tribes, apparently of Arabic origin. These tribes controlled the trade routes, especially those of the frankincense and myrrh, which was brought from southern Arabia to the Mediterranean ports. Their capital city was for some time in Petra, situated in a canyon which cut through the mountains of Edom. In the desert of the Negev, they built six beautiful cities, which flourished until the 7th century A.D. During these years there occurs in the core in the Sea of Galilee, a depletion in the ratio of ^{18}O and ^{13}C isotopes, there also occurs an increase in the ratio of olive pollen and reduction in that of oak. These fluctuations are explained by the cutting of the natural vegetation and planting of olives.

Kedar (1985) in his study on the Arab conquest and agriculture, in the seventh century AD, maintains that the destruction of the settlements of the Negev was by no means the repercussion of the Arab conquest on fertile crescent agriculture. The keen interest of the early caliphs in agricultural exploitation in the area between desert and sown land is convincingly attested. The restoration of the hot baths at Hamat Gader in 662 AD points in the same direction.

Flemming (1989), discussing the tectonics and sea levels of the harbors of Caesarea Maritima, gives estimates for the eustatic sea level in general at:

8,000 BC -- -50 m
5,000 BC -- -10 m
3.000 BC -- ± 1 m

Then from the early bronze age data until the present the data should reveal the fine structure of both tectonic and eustatic changes.

Overall, there is a slight statistical indication that the relative sea level on the Mediterranean coast of Israel has been 0.5 m lower throughout most of the last 4,000 years. The author concludes that there is no archaeological evidence for vertical tectonic movements of more than 2-3 m anywhere on the present Mediterranean shore-line of Israel within the last 4,000 years. However at Caesarea, at 150 m off shore, there has been a 5 m subsidence since the late 1st century B.C.

Galili et al. (1985, 1988), who investigated the Holocene sea-level changes based on submerged archaeological sites off the northern Carmel coast in Israel, report that the northern section of the Carmel coastal plain between Atlit and Haifa has so far revealed the largest number of submerged prehistoric sites along the eastern Mediterranean and indicate a continuous marine transgression between 8,000 and 1,500 yr BP. The sites are embedded in the upper part of a marshy clay that fills the trough between the coastal ridge and a ridge now submerged some 1,000 to 1,500 meters in the sea to the west. The submerged prehistoric sites belong to two main chronological units : prepotterey Neolithic B (8000 P) at depths of 12-8 m, and Late Neolithic (about 6,500 BP) at depths of 5-0 m. Bronze Age and Byzantine anchors were found at depths of 5-3 m and 4-1.8 m respectively. A continuous Holocene rise of sea-level from -14 or -15 mat 8,000 BP to the present level, which was definitely reached by 1500 BP and possibly by 2000 BP. According to their evidence no tectonic movements have occurred in the area during the last 8,000 years. Wreschner (1977) investigated the Mediterranean sea level changes according to settlement location in the coastal plain of Israel during the Holocene, his curve is very tentative, the only conclusive information is that of around 6,000 BP, i.e Chalcolithic cold period, in which the level of the sea fell about 5 meters below present MSL.

A series of warm years occurred between 1,920 to 1,963 AD. After 1,930 AD the level of the Dead Sea dropped by about 2 meter from a level of about -390 m, which had been maintained since 1,893 AD, to a level around -392 m. The Dead Sea remained at the latter level, albeit with fluctuations, from 1,940 until 1,959 AD, after which its level has continued to decline (Ashbel, 1938).

Lev-Yadun, et al. (1987) investigating annual tree rings as an index to climate change intensities in the Levant, found that the most suitable trees in this region for dendrochronology are: Red Juniper, Persian Pistacia and Cedar of Lebanon. The rings of Jerusalem Pine, for example, show that the range of changes in rainfall were about 400 to 900 mm per year. The changes are not linear and under 400 mm of rainfall there is no growth at all. Studies of the rings of the Cedar of Lebanon and Red Juniper trees show that in the last 2,000 years there have been only slight changes of climate, some of which were regional and some only local. From around 420 to 480 AD there were very narrow widths of tree rings, which means drought.

Dendrochronological investigations have been carried out mainly by Liphschitz (1967, 1976). The studies show that no correlation could be found between annual radial growth of Quercus boissieri trees and distribution of rainfall in the respective rainy season, neither with the total amount of rainfall nor its monthly distribution. Thirty-eight Quercus boissieri trees were examined, from Mt. Meron, Israel, growing at an altitude of 890 m. This tree is one of the three species of oak native to Israel. Mount Meron can be considered to be humid, and under such conditions it is expected that precipitation will not be the limiting factor for tree growth. Temperature seems to be the most important single factor on tree ring growth in this case, but the dependence on temperature conditions is rather many sided : depending on the months of the year (Liphschitz and Waisel, 1967).

One effect of human activity on the composition of the natural vegetation during historic periods has been the disappearance of date palms from the region of Jericho and Ein Gedi since the 15th century AD (Liphschitz and Waisel, 1974).

The investigation included trees In St. Catherine's Monastery in southern Sinai, which was established in the time of the Byzantine Emperor Justinian (525-565 AD). A tower of a mosque was built there during the days of Sah el-Aphdal, in about 1,106 AD. The widespread use of Populus, Phoenix dactylifera and Cupressus sempervirens logs as building material in St. Catherine's Monastery, from the 6th to 19th century, suggests that these trees were common in the area in the past while today they are scarce (Liphschitz and Waisel, 1976).

Changes in the distribution pattern of various woody species in Israel have occurred during the last 5,000 years. Such changes in plant composition indicate that the climate of Israel is gradually becoming drier and warmer. Overuse of plants by man and the consequent denuding of large areas may have intensfied such a process of desertification. Short term climatic fluctuations left signals in the annual rings of the local trees, but were too small to exceed the physiological tolerances of most plants except in transition zones. (Waisel and Liphschitz, 1968). Waisel (1986), investigating floral remains in archaeological sites concluded that at Tel Sera (NW Negev), which today is semi-desert and where one finds a few trees of Acacia Raddiana and Tamarix aphylla, from 2,000 to 1,200 BC the flora was that of a Mediterranean climate, dominated by olive trees, while acacias were almost absent. In Beer Sheba and Arad, (N. Negev) at 2,800 BP were found Quercus ithalurensis and pinus halepensis. Later on and even today, the above trees had completely disappeared and were replaced by Acacia sp. and by Tamarix aphylla (Liphschitz and Waisel, 1974).

In the Central Negev, Israel, paleo-environments can be deduced from tree assemblages of prehistorical times (Liphschitz and Waisel, 1977): 42,980 ±2,420 BC - Pistacia Olea; 25,000-22,000 BC - Olea, Pistacia, Tamorix aphylla Pistacia atlantica; ±12,500 BC (Geom. Kebaran "A") - Pistacia; 11,580 ±144 BC (Moshabian) - Conifer; 8,280 ±150 BC (Harifian) - Pistacia; ±6,700 BC Paleo-environments (PPN''B") - Pistacia - olea; 4,010 ±100 BC (Late Neolithic) - Conifer. Tonnonex aphylla occurs today in the Beer Sheva region. Regarding olea trees, since the findings are from eras prior to 5,000 BP, a period which is considered more humid, it is possible that olive trees as well as the conifer (cupressus) could have grown under the wetter conditions.

Liphschitz et al. (1981) investigated the assemblage of trees from the Roman siege ramp which was constructed for the siege of Masada in the Judean Desert in 70 AD. The scarcity of Acacia wood and the abundance of Tamarix, relative to the present, indicate a change in climate to wetter conditions. Had there been more running water the Acacia would have been less common than Tamarix. Nevertheless, as only a few species have disappeared from this area during the last 2,000 years and as the species composition appears to be fairly constant, one may assume that apart from possible short fluctuations, the climate as a whole in the area has remained more or less unchanged.

Nir and Eldar (1987) investigated ancient wells in the Israel Mediterranean coast line region, for detecting young tectonical movements. The wells represent the period from the Late Bronze Age to the Crusader period. The curve they present coincides well with the master diagram of the Levant, thus it shows the influence of change in precipitation, which during the Roman Byzantine time reached a maximum receding during the moslem time.

Archaeological evidence indicates that Haifa Bay has dramatically filled in with sand, resulting in the shoreline advancing several kilometers since the Middle Bronze Age, around 4,000 BP (Sivan, 1982).

Striem (1985) investigated the quantitative and qualitative aspects of recent climatic fluctuations in Israel. This was based on the systematic meteorological observations which have been made in Jerusalem since about 1,850 AD, constituting the longest record of climatic parameters in the Middle East. He found that winter (December to March) temperatures rose by more than l°C, while the summer (June to September) temperature remained steady. The mean monthly barometric pressure rose by about 1 mb. The annual rainfall decreased from about 600 mm in the second half of the 19th century to about 500 mm in the first half of the 20th century. There is at present (1,985 AD) a trend of increasing rainfall. The pressure maximum used to occur before November (44910 from 1,861 to 1,885 AD), but only 17% from 1,946-1,970 AD. Its occurrence in November became more frequent (an increase from 24% to 46%). The retardation in the occurrence of the barometric pressure maximum is a characteristic of drought years.

Van Zeist and Bakker-Heeres (1979) investigated the economic and ecological aspects of the plant husbandry of Tell Aswad, situated some 30 km ESE of Damascus. Mean annual rainfall is below 200 mm at present. The subsoil in the Aswad area consists of lacustrine sediments deposited in a lake which Up to Late Pleistocene times must have covered a long part of the Damascus basin. The natural vegetation is that of a steppe. It is likely that in early Neolithic times the lake and the marshes extended up to the site of Aswad, as it did in 1,852 AD, when the lake expanded to the edge of Tell Aswad. In 1,880 AD the south Lake had already receded to some extent and today only pat-t of the north Lake of Aateib' remains.

Egypt, although geographically belonging to the Levant, has the source of its main water resource (the Nile) in the tropical and sub-tropical zones. In this subchapter only the Mediterranean influenced regime will be discussed. A mean sea level rise of 8.4 cm was observed at Alexandria between 1,944 and 1,973 AD, i.e 2.9 mm/yr. At Port Said during the period 1,926-1,970 AD, the mean sea level rose 10.1 cm, i,e 2.2 mm/yr. A part of the rise that has taken place at the Nile Delta coast can be explained by the thermal expansion of the upper layer of the oceans, resulting from the observed warming of 0.4°C in the past 100 years. The other part of the rise may be related to subsidence.The prediction of sea level rise by 2,100 AD of the Nile delta coast is 37.8 cm and 28.6 cm over the 1,970 AD level at Alexandria and Port Said, respectively (El-Fishawi and Fanos, 1989).

4.1.2 Anatolia

A geological study has been carried out in Lake Van, Eastern Turkey (Fig. 2, Degens et.al., 1984). It is a lake with a volume of 607 km^3 and a maximum depth of 451 m in a tectonically active zone in eastern Anatolia. The fine laminations of the sediment of Lake Van are interpreted as varves: a white carbonate layer is deposited in winter and a dark layer during the summer. The lake level was at its highest at 72 m above the present at the height of the last ice age about 18,000 BP. A dramatic drop to over 300 m below present occurred about 10,000 BP with an equally dramatic rise aroud 7,000 BP. Paleoclimate (based on pollen) indicates the following: 10,000-6,500 BP, steppe vegetation; 6,500-3,400 BP, forest vegetation; 3,400 BP to the present, relatively moist, The conflation between the paleo-levels of Lake Van and the base-section gives the results shown in Fig. 5. While at the lowest Base of the Neolithic period the level of the lake is still high, it fell around 10,000 BP to a very low level which caused a "salinity crisis". It rose back dramatically during the lower Chalcolithic and then after a few undulations, with a small recession during the Middle Bronze around 4,000 BP, it rose again to above its present water level during the Iron Age period (around 3,000 BP). After a recession during the Persian and Hellenistic periods the level rose again to more than its present level at the Byzantine period (around 500 AD). A recession characterizes the period around 1,000 BP (Moslem period). A rise of the level at around 500 BP maybe connected with the Little Ice Age.

Correlating this with the sedimentation rates curve, one can say that (excluding the rise is sedimentation rates during the last 10 centuries, which are most probably anthropogenic), in general the rate of sedimentation (especially during the Neolithic) are higher during the drier periods. This may be connected to the reduction in the vegetation cover.

In general it can be concluded that the lake level fluctuations correspond rather well with that of the Dead Sea (Frumkin et al., 1991) except for the Late Roman Cold Spell, where the rise of the Dead Sea level is earlier than that of Lake Van. This might be due to differences in the precision of the methods of dating involved (varve counting in cores in Lake Van and carbon 14 methods in the Dead Sea).

Pollen diagrams from Lake Zeribar Kurdistan, Zagros Mountains, (El-Moslimany, 1986) show the absence of trees during the last glacial period and the migration of forest into the region between 10,000 and 5,500 BP. This has been interpreted as indicating aridity during the Pleistocene with gradually increasing precipitation in the late glacial and Holocene. However, the sensitivity of these species (Quercus aegitops sp. brantic and the associated pistacia atlantica var. mutica and pistacia khinjuk) to snow and their tolerance of low overall precipitation implies higher snowfall rather than low precipitation as the cause of their absence during the Pleistocene. The inability of seedlings to survive the present summer-day conditions suggests that summer rainfall, and not higher total precipitation was the factor that finally allowed migration of these trees to this region. There is ample evidence from lower latitudes of a northerly extension of the summer rainfall zone, centering on about 9,000 BP.

Climatic modelling suggests that this was due to higher summer insolation in the Northern Hemisphere during the same period.

In the seventh century A.D. many of the monasteries, strewn all over Anatolia and Greece were deserted. (Carpenter, 1966). Liphschitz, Waisel and Lev-Yadun (1979), investigating the dendrochronology of Pinus nigra in south Anatolia (Turkey) found, in the Mediterranean region at altitudes between 300 and l,800 m where the annual amount of precipitation is above 1,000 mm. The tree produces distinct annual growth rings and attains ages of several hundred years. Samples from the Taurus mountains, 50 km north to Karsanti at an elevation of 1,775 m., where annual rainfall varies between 735-1,400 mm, shows: wide rings, group 1 -1,670-1,710 AD and 1,800-1,820 AD; group 2- 1,740- 1,760 AD and 1,920-1,950 AD -- narrow rings; group 1- 1,720-1, 740 AD and 1,830-1,850 AD; group 2-1,780-1,800 AD and 1,880-1,910 AD.

Results show that radial growth in Pinus nigra depends on temperature conditions in certain months more than on any other climatic factor (Liphschitz et al., 1979). Dendrochronological investigations in Iran, on Juniperus polycarpos growing in west and central Iran reveals that the radial growth in this species depends mainly on the amount of precipitation in the more arid regions. when the amount of rain is sufficient, i.e. above 450 mm, the prevailing summer temperature seems to become the limiting factor. Wide rings (favorable conditions) were found in 1,685-1,695 AD and 1,790-1,800 AD. Narrow rings (less favorable conditions) were found in 1,725-1,735 AD and 1,855-1,865 AD.

Serre-Bachet and Guiot (1987) investigated summer temperature changes from tree rings in the Mediterranean area during the last 800 years. Their conclusions are the following: 1,150-1,250 AD, intermediate; 1,250-1,340” AD, warm; 1,340-1,400 AD, cold; 1,400-1,455, intermediate; 1,455-1,550 AD, warm; 1,550-1,605 AD, cold; 1,605-1,625 AD, warm; 1,625-1,650 AD, cold; 1,650-1,685 AD, warm; 1,685-1,705 AD, cold; 1,705-1,740 AD, warm; 1,740-1,760 AD, cold; 1,760-1,810 AD, warm; to intermediate; 1,810-1,840 AD, cold; 1,840 AD to the present, many fluctuations.

4.1.3 The Balkan peninsula

The Mediterranean climate in the Balkan peninsula is limited to the coastal area and the island region. On the Adriatic coast, the average winter temperature is above freezing, except for the northern region, where continental cold air is brought by the Bora wind. In July, the average temperature is between 23°-25°C, due to maritime influences. In the coastal area, precipitation is heavy. Some places in (the former) Yugoslavia receive more than 4,000 mm of rain annually. On the western mountain slopes and in low-lying areas, annual precipitation is much less (500-600 mm). In the southern areas, the Mediterranean climate is much more pronounced, with lower precipitations and higher temperatures (for example, Athens never receives more than 400 mm of precipitation annually; the average July temperature is 27°-28°C, while in January it is 7°-8°C).

The climate on the southern Aegean coast tends toward the continental, with comparatively lower precipitation. In the Aegean islands, the climate is relatively dry, while the climate in the Ionian sea is more hot and humid.

Landscape changes in Greece as a result of changing climate during the Quaternary have been investigated by Papae and his colaborators (Papae, 1984). The following are the main results of his work: “R. Paepe and M. E. Hatziotis worked out in the area of Attica (Greece), more specifically in archaeological excavation sites of Academia Platonos in Athens, in the Marathon Plain in coastal sites the Temple of Artemis in Brauron (E. Attica) a lithostratigraphy dated on basis of archaeological elements (6). C. Baeteman at the same time studied the marine sequences where D. Tsouclidou studied the relationship between marine and continental deposits in Brauron.

“Putting together all evidence after comparative study of all sites combined, the lithostratigraphic record (Fig 9 revealed in the Haradros Complex of Marathon six Holocene Soils of which respectively the earliest one (Marathon Soil, HS1) and the last one (Kallileios Soil, HS6) are the most developed. With regard to the Neolithic finds, the Marathon Soil most probably developed about 7.000 BC (9.000 BP); the Kallikleios Soil instead was very accurately dated (725 BC ± 5 y.) thanks to the presence of Geometrical tombs in many sites of the Academia Platonos.

“Strikingly HS 3, HS 4 and HS 5 together with relevant fluvial gravel deposits perfectly encompass the three phases of the Helladic period. together with the Kallikleios Soil (H.S. 6) the subdivide the Subboreal Substage into four cycles of approximately 500 y. Soil formations in the fluviatile valley system perfectly tally with peat development in the marine sequence of Marathon. Furthermore, in between soil development phases, fluviatile sedimentation rates score the highest values.

“In Marathon, however, no soils are found within the timespan of the geological Atlantic substage coinciding with the Neolithic. Nevertheless H.S. 2 and H.S. 1 close the fluvial cycles of respectively Boreal and Pre-Boreal Substage inferring a 1.000 Y. periodicity.

“This sequence was recently completed with a more detailed profile from Academia Platonos (Kratilou section). It produced at least 6 other soils in between H.S. 2 and H.S. 3 namely: H.S. 2 a, b, c, d, e, f. Some of these soils were more weakly developed: gley and steppe soils. It points to the fact that weaker climatic oscillations interfered. However, the presence of these soils testify once more of the 500 years periodicity.

“By the time of the development of the Kallikleios Soil about 725 y. B.C. all valleys and coastal plains are completely filled up, to the level very near of today’s surface.

“AS to then sedimentation in general slowed down except for the peaks coinciding with the fluviatile phases which point to high sedimentation rates.

"In Marathon as well as in Academia Platonos usually a series of five Holocene soils is recorded: H.S. 7, H.S. 8, H.S. 9, H.S. 10 and H.S. 11. They induce a periodicity of 500 years. Clay mineralogy as well as textual evidence furthermore give evidence of four dry cycles respectively towards the end of the Geometric period (8 Cent. BC), in the Middle and Late Roman Period (2-4 Cent. AD), in the second half of the 12 Cent. AD and today. They reveal a periodicity of 1.000 years."

There exist many discrepancies between interpretation suggested by Papae and the base-section. The tentative interpretation by the author of this paper would be the following: Cold wet periods will be characterized by rather small sedimentation due to the fact that the forest expands and there is less soils erosion, while during dry periods higher rates of soil erosion causes higher sedimentation rates.

Correlating the curve of sedimentation rates with the base-section (Fig. 6) one can give the following interpretation: During the Neolithic the Pleistocene forest still dominated the area, protecting the soil cores. Only towards the upper Neolithic do the rates of sedimentation become higher due to the warming. The rates fall again towards the Chalcolithic and Early Bronze, but rise toward the Middle Bronze. The Iron age is again characterized by low sedimentation, i.e. a cold climate.

During the Roman periods the rates are in general high, presumably due to the cutting of the forests, but still one can see fluctuations which correspond to cooling and warming. Around 700 AD, namely the Moslem warm spell, rates of sedimentation became higher, corresponding to the global warming phase, The grain size median (Fig 6) more or less follows this curve. As mentioned earlier, this is a tentative interpretation which has to be further investigated.

4.1.4. The Italian peninsula

The Italian peninsula is characterized by mild, relatively humid winters, and hot and dry summers. Sicily and the southern Italian peninsula are areas which are typical of this climate. Due to the presence of the Appenian mountains, the western half of the peninsula receives more precipitation than does the eastern half. The western coastal area is hotter (8°C in January), wetter (2,000-3,000 mm annually) and has fewer instances of freezing weather (in Florence and Rome, the average January temperature is 5-6°C, with 1,000 mm annual precipitation) than other areas of this type of climate. In the eastern part of the region, precipitation is much lower (500-600" mm annual) with hot summers (average July temperature is 24-25°C in Calabria and 28°C in Sicily). The dominant summer wind pattern is the Sirocco, originating from North Africa. This wind is searing, dry and violent.

A study of the lake levels and climate for the last 30,000 years in the Fucino area, Central Italy, has been carried out by Giraudi (1989). The lake is situated in the Appenines, east of Rome. Before being drained Lake Fucino measured 150 km^2 with a drainage area of 71O km^2. Now approximately oval in shape and 19 km long from NW to SE, it has a maximum width of 10 km and a maximum depth of roughly 22 m. The surrounding area, essentially limestone rock, ranges from the 900 m of Monte Salviano up to 2,349 m of Monte Sirente.

Geological and geomorphological mapping identified delta and lacustrine sediments, shoreline terraces, wave cut terraces and soft-rock pediments. Each provides evidence of ancient levels of Lake Fucino during the late Pleistocene and Holocene.

The following changes can be recognized:

a slight rise in lake level dating back to an indefinite period before 33,000 BP;
strong rise in level during the period between about 30,000 and 20,000-18,000 BP;
a fall between 20,000-18,000 BP and 7,500-6,500 BP;
arise between 6,000 and 5,000 BP:
a drop between 5,000 and 2,800 BP;
arise between 2,800 and 2,300 BP;
a drop between 2,300 and 1,800 BP;
draining by the Romans (2nd to 5th or 6th century AD);
a low level during the 16th century AD;
an increase in levels after 1,750 AD.

Regarding the variations of lake levels, data are available on different scales:
for the last period of existence of Lake Fucino, from 1,750 to 1,861 AD, there are measurements of levels first on a decade scale, then intermittently, then annually and later, in the years before the drainage, monthly;
for the period between the Iron Age and the Roman Age there are indications of variations over centuries;
for the remaining part of the Holocene, variations are known for intervals in thousand of years;
for the late Pleistocene, resolution is on a scale of ten thousand years.

From the curve of the upper Pleistocene (Fig. 7), it can be said that the levels correlate with the advance of the last glacial period. This means that with advance of the glaciers the lake received a higher rate of inflow this can be explained by the higher ratio of precipitation on its drainage basin. This can be seen also by correlating the recent change of levels with the Alpine glaciers.

During the Holocene the picture (Fig. 7) seems to be more or less the same while the level in the Neolithic (from around 10,000 to around 7,000 BP) is still low. It reaches a peak level during the Chalcolithic cold and humid period and recedes toward the Middle and Late Bronze in which it reaches its minimum namely the warm and dry period which affected the whole Mediterranean basin. At the Iron period, around 3,000 BP, it starts to revive. It comes up to such a level (the short warm dry spell of around 2,500 BP was not observed) that the Remans have to drain it.

A warm period of global extent occurred between the years 1705-1880 AD. Four tree ring series, derived from trees which grew at higher altitude (1,750-2,300 m elevation) in southern France and southern Italy (Calabria), were used to study summer temperature changes during the period from 1,150 AD until today (Serre-Bachet and Guiot, 1985). A sharp warming occurred around 1,705 AD which lasted until about 1,810 AD, although with fluctuations and a cold period between 1,740-1,760 AD.

4.1.5. The southern Iberian peninsula

The climate in the southern Iberian peninsula, especially in the Andalusian plain is similar to that found in Sicily. The average January temperature is 12-13°C, while in July the average temperature is 27-28°C; annual Precipitation is 500 to 700 mm Along the length of the Mediterranean coast (the “Spanish Levant”) precipitation is less (300-350 mm per year). These characteristics show that this region is a transitional area towards the hot and dry regions of northern Africa.

Some works on Quaternary climate changes in the Peninsula can be found in the Proceedings of the Symposium of Climatic Fluctuations during the Quaternary in the Western Mediterranean Regions (Lopez-Vera, 1986). A palynological investigation was carried out on the sediments in the marshes close to Guadiana river (Garcia et. al., in Lopez-Vera, 1986). In the general section, (Fig, 8) one can see that there exist changes in the ratio of Pinus to Quercus. These changes are climatic indicators because, as shown by Pens and Reille, 1986) for the Granada region, a high ratio of Pinus is characteristic of the glacial periods, and vice versa. In the Guadiana region the scarcity of ^{14}C dating does not permit correlation with the Levant base curve, but from the dates available one can see that around 6,200 BP there is a high ratio of Pinus which might correspond to the Chalcolithic cold period of the Levant. This is followed by a short period of decrease in the ratio of Pinus to Quercus, which may correspond to the upper-most part of the Chalcolithic. The period of increase in the ratios which follows may represent the cold period of the Early Bronze, followed by a period of decrease which may correspond to the warm Middle Bronze period. Towards the layer dated to be around 1,730 BP, there is again an increase, cut at about 2,000 BP (not dated) by a layer of about 10 cm devoid of pollen. It is suggested by the author of this document that this is equivalent to the period of high precipitation during the Early Roman period, in which the marshes turned into a flowing river, causing oxidation of the sediments. The warm Moslem period seems to be reflected by low Pinus ratio later.

4.2 The Sahara and the Sahel Belt bordering tropical Africa

The climate of the Sahara desert is a product of the geographic location of this region. The Sahara is located in an area of active trade winds, and the anticyclonic subtropical areas. These two factors have a negative influence on precipitation. In winter, the high-pressure areas come from the Azores anticyclone, which makes contact with the high pressure area in the Central Sahara. During the summer these two anti-cyclonic areas move northward. Due to the high temperature of the Sahara, the air masses do not produce much moisture. In most of the Sahara region, the average annual precipitation is less than 200 mm and appears only at irregular intervals. Rainfall may occur in any season, but it is quite possible for no rainfall to occur for several yearsOn the Saharan Mediterranean coast, in the area of Tripoli and Cyrenaica, average annual precipitation is 200-300 mm.

The Khamsin winds are active in the Saharan coastal area. These winds bring large quantities of dust from the desert and deposit them as far away as southern Europe.

The Saharan climate is characterized by high thermic amplitude, and very reduced humidity. The average summer temperature is 36-37°C; the diurnal amplitude is more than 30°C in the shade and 50°C in the sun.

In North Africa, which still belongs to the westerlies belt, a humid period can be clearly recognized from about 5,000-3,000 BP, characterized by a period of geomorphological stability and pedogenesis producing dark humic soils, being different from the underlying reddish soils (Rognon, 1987). Alluviation of silts and fine sands occured during this wet period in many valleys of North Africa. Swamps and small lakes developed in closed basins along the northern Sahara margins (Rognon, 1987).

Pachur and Braun (1980), investigating the paleoclimate of the Libyan desert of Central Sahara, found a remarkable difference between west and east from 10,000 and 3,000 B. C.: In Libya biologically highl$_{y}$ active freshwater lakes; in Egypt swampy environments with high salt content and active sand transport. Autochthonous rainfall in the interim period caused decreasing humidity from west to east. The period after 3,000 BC shows a trend of decreasing precipitation. Around 1,000 BC it was more moist, although insufficient to cause lake formation.

Petit-Maire, et al. (1980) report on Pleistocene lakes in the Shati area, Fezzan (27°30’ N). In the Shati valley of Central Libya, about half-slope up between the sebkhas and the tertiary limestones, a 125 km long line of oases marks the position, between Ashkeda and Adrei of ±800 wells and springs from the “Wadi” Shati artesian rising waters, which correlates with the Early Bronze wetter period.

4.2.1. Eastern Sahara and Sahel

The climate conditions in the monsoonal belt of Eastern Africa can be deduced from the interpretation of the correlation between ’^{8}0 and pollen records from an off-shore marine core in the northern Arabian Sea (Van Campo et al., 1982).

The record from the core has been correlated with the distribution of modem pollen. It was found that low sea-level glacial intervals contain well preserved pollen which indicate saline littoral, arid and steppe inland conditions. High sea-level interglacial intervals are characterized by savanna type vegetation. “The isotopic records of the planktonic and benthic foraminifera show, as usual, several oscillations. These d^{18}O% variations closely match those of other cores from the Atlantic, Pacific and Indian oceans, in agreement with the hypothesis that the major signal in these records reflects isotopic changes in the ocean water mass resulting from the waxing and waning of continental ice sheets. " The general conclusions of the authors are that, "Glacial periods were arid in southwestern Asia and at low latitudes, and the north-east trade winds were intensified against a decreased monsoonal flow. In the tropics, as evidenced by the most detailed Holocene record, incipient interglacial stages were the most humid periods: the south-west rnonsoonal flow was intensified, over the Atlantic and Africa, and over the Arabian Sea and India. The insolation peak developed low pressures over the tropical continents and heavy rains resulted. Meanwhile, the middle and high latitudes, more influenced by the northern ice-sheets, also had warm summers although dry, and reached their humidity peak later." (Van Campo et al., 1982).

Enlarging the upper section of these curves and correlating it with the base-section (Fig. 9) one can see that while during the Holocene, around 10,000 BP, the base isotopic composition is rather positive, there was a decrease of 18°C and increase in the total humid tropical taxa around 8,500 BP which marks the lower Neolithic warm (but wet due to rnonsoonal circulation) of the Neolithic (Pre Ceramic B) period. The Chalcolithic cold period is characterized by a rise in the ^{18}O, a decrease in the tropical pollen and increase in the sub arid Mediterranean, indicating a weaker monsoon and a strengthening of the NE winds. The early Bronze cold period is marked by a decrease in the isotopic composition which indicates melting of ice, on the other hand the pollen still indicate a low monsoonal influence. The lowest peak of the SW monsoonal winds (which coincides with the heaviest composition of isotopes, i.e. minimum effect of waning of continental ice sheet) is during the Iron and Roman Cold periods. Later, toward the Moslem period, there is a strengthening of the monsoonal circulation and lighter composition of isotopes, i.e. melting of the glaciers (with a change sometime around the little Ice Age?)

It goes without saying that this correlation is very general due to the low resolution and thus little accuracy due to the enlargement of the scale. Nevertheless, the general pattern of the distribution of the pollen and changes in the pollen composition correlate well with the base-section. This means that in general there should be a negative correlation between the southern Sahara which is related from the climatic point of view to the sub-tropical Sahel belt (mainly influenced by the monsoonal regime) and the north Sahara which is related to the Circum-Mediterranean area, under the influence of the Westerlies.

The Red Sea is also influenced by the climate regime in the Eastern Sahara, and by the sediments transported to it from its rivers. Thus the impact of its climate regime on the hydrology can be derived from the investigation of the sediments of this sea. Almogi-Labin et al. (1989) investigated the palaeoenvironmental events in a Holocene sequence from the central Red Sea as recorded by pteropada assemblages. Eight events could be recognized by the alternating biological and sedimentological features:

10,000-7,800 BP, namely during the First 2,200 years of Early Holocene, water column conditions were unstable; the invasion of normal marine Indian ocean water caused a sharp decrease in salinity throughout the whole water column, followed by a strong stratification of shallow water depths; increase in humidity in the surrounding landmass probably caused this stratification;

7,800 BP, similar to present conditions first recorded in the Red Sea;

5,000-2,000 BP, the Red Sea was more oligotreophic than today and the oxygen minimum zone was less developed.

The response of floods in the Sinai peninsula to meteorological events is discussed by Ashbel (1938).

The Eastern Sahel region includes the transitional zone between the Sahara and the Sudanese region. The average annual precipitation does not exceed 400-500 mm (190 mm in Timbuktu; 540 mm in Niamey). The rainy period is 3-4 months, during which the effect of the equatorial monsoons are felt. In the hottest month, temperatures vary between 33-35°C; during the coldest month, the temperature is 20-22°C.

The Ethiopian highlands are influenced by this climate and, in turn, they influence the hydrological regime of the Nile, the longest river in the world (measuring about 6,600 km from its headwaters in Rwanda to its outlet in the Mediterranean Sea). The regulating storage of the Nile is supplied by the lakes and marshes of subequatorial Eastern Africa, which are fed by the tropical rains falling on east central Africa and the subtropical monsoonal system stretching north to the equator. The latter system is seasonally, as well as periodically, in an off-phase regime to that of the Mediterranean.

Lezine and Bonnefile (1982), investigating the palynology of the sediments of lake Abiyata (Ethiopia), found that the upper 30 m of a 162 m core taken near the shore of the lake in the Ethiopian Rift Valley covers the time span from around 40,000 to 6,000 BP, with the period prior to around 30,000 BP, 30-19.70" m depth. Generally, the semi-arid climate was close to the present climate. From 30,000 to around 10,000 BP (19-8 m depth) the core is discontinuous, but suggests humid conditions; from 8-5 m - more humid. Around 9,950 ±170 BP, however, a stabilization of the lake level (5,5-5 m) occurred. From 5 to 1,70 m - there was a colder and humid period. From 1,70 to 1 m - around 7,000 BP it was a very dry period, later turning humid.

A review of palynological and paleobotanical data from the Sahel (Lézine, 1989) shows fluctuations in the vegetation zones in tropical north Africa for the past 20,000 years. The migration to the south of the Sahelian pseudo-steppe occurs as far as 10°N during the last arid episode (18,000 BP). During the Holocene, two abrupt changes in vegetational landscape occur at 9,000 and 2,000 BP. The first (9,000 BP), corresponds to a rapid expansion of humid vegetation zones by 400-500 km to the north of their modem position. The 14-16°N latitudes are covered with Guinean and Guineo-Sudanian forest. The second (2,000 BP), corresponds to the disappearance without any transition of the arboreal formation and to the setting of the modem semi-arid environment in the Sahelian zone.

The late Quaternary history of the Nile (Adamson, et al., 1980) shows that from 20,000-12,500 BP it was in the Intertropical cold dry phase, in which the aggrading Nile was a braided, highly seasonal river. From 12,500-5,000 BP an overflow from Lake Victoria and higher rainfall in Ethiopia sent extraordinary floods down the main Nile, marking a revolutionary change to continuous flow with a superimposed flood peak. The main Nile, and its tributaries, established more stable channels of higher sinuosity, from which suspended Ethiopian silt and clay was deposited on the floodplains. The Blue Nile began to cut down, thereby depriving the Gezika fan of flood waters. From 5,000 BP to the present (by the mid-Holocene) the modem era of the Nile had begun.

Flohn and Nicholson (1980) observed two separate wet periods in the around 7,500 BC which coincides with the warmest Holocene period in southern latitudes, and around 4,000 BC which coincides with thermal maximum in Europe (Atlanticum). The very dry periods in the Sahara were between 5,000-4,000 BC and 1,000 AD to the present.

Lauer and Frankenberg (1980) found a relationship between plant cover, number of plant species, and the annual mean water balance. Their conclusion is that about 15,000 BC the climate was cold and dry while at 3,500 BC it was warm and relatively wet.

Lake levels in Africa and the Arabian peninsula were at their highest levels from about 9,500- 5,000 BP, except for an interval around 7,500 BP when lake levels were lower (Street-Perrot et al., 1985). A climatic model incorporating changing orbital parameters and surface boundary conditions simulates a strengthened monsoon circulation and increased precipitation in the Northern Hemisphere tropics, culminating at 9,000-6,000 BP (Kutzbach and Street-Perrott, 1985).

Two periods with high Nile levels and aggravation can be recognized in Egypt: 11,200-7,700 BP and 7,300-6,000 BP. These two periods are separated by a regression (Butzer, 1980), apparently caused by a more arid climate in the entire region of northern Africa, as recognized from a drop in lake levels around 7,500 BP (Street-Perrot et al., 1985).

The Holocene history of the Nile Valley in Egypt and Lower Nubia can be summarized as follows:

The wet Lower Bronze period as indicated by the Nile was that of low monsoonal rains. There are historical records of declining flood levels of the Nile in the period dated to 5,000-4,800 BP. Catastrophic low floods occurred in the period 4,250-3,950 BP. Repeated high floods were recorded from 3,840 to 3,770 BP, while average flood levels remained high in the period 3.770-3,180 BP. A strong decline in flood levels occurred from 3,180-3,130 BP. “Normal” levels, comparable with the present generally characterized the Nile regime in the period 3,130 BP-1400 BP (Butzer, 1980). The level of the Nile during the Roman period was low (Nicholson 1980), and high flood levels in the period 600-1000 AD (Butzer, 1980).

The present level of the Nile was reached about 5,000 BP. Archaeologically dated high water marks on Egyptian temples and associated Christian construction near Wadi Halfa suggest that the last important high Nile periods were about 500 and 800 AD (Fairbridge, 1962).

Recent discoveries of fossil-bearing Holocene lake sediments from the eastern Sahara have brought further confirmation and detail to earlier indirect evidence that a major pluvial episode occurred between 9,500 and 4,500 BP. The botanical records, mainly palynological show that savanna and desert grassland occupied regions that today are plantless hyperarid deserts. The indication is that the well defined latitudinal zonation that characterizes the modem vegetation also existed in the early Holocene, displaced northward by at least 2° (or 450 km). These and other analyses from the Sahara-Sahelian belt imply that the steep gradient of summer precipitation (100-440 mm/yr) which occurs from 12-17°N was displaced 4 to 5° northward from 10,000-5,000 BP (Ritchie and Haynes, 1987).

Gasse (1980) in her study on late Quaternary changes in lake-levels and diatom assemblages on the southeastern margin of the Sahara reports on a wet period of expanded lakes between 5,000-2,000 BC, a dry period of widespread regression from 2,500-1,500 BC, and that most of the lakes recorded low levels and several of them dropped below their present levels by 1,500 BC. From that time to the present a rather dry period was punctuated by short humid phases (700 BC- 1,000 AD, wetter in Ethiopia).

4.2.2 Western Sahara and Sahel

Servant & Servant-Vildary (1980) found paleoclimate in the Chad basin that seems to be related to variations in the frequency of incursions of advected cold polar air to lower latitudes. Changes in the strength of the sub-tropical anticyclotic cells is a major factor controlling late Quaternary climatic changes. In order of decreasing importance, the wettest phases during the Holocene were:

-9,000-8,000 BP, (P > E);
-6,000 BP, (P < E);
-3,500-3,000 BP, (P < E);

The driest phases were:

-10,000 BP, 7,500 BP, 4,500-4,000 BP and historical times.

The coldest water temperature:

-12,000-7,500 BP.

Tropical diatom associations, indicating warmest water:

-7,000 BP to present.

From 8,000-4,000 BP Sudan-Guinean and Sudan-type pollen abounded in what is now the Sahel zone. Indeed, from terrestrial evidence in the Sahel (deflation and incision), it is clear that an arid phase occurred between 8,000-7,000 BP, followed by another major humid period lasting from 7,000-4,000 BP (Talbot, 1980; Rognon, 1987). The latter climate seems to have been less humid, however, than in the period 10,000-8,000 BP. Moreover, unlike the previous wet period, rains were again concentrated over a few months of the year and the dry season expanded progressively (Street, 1979; Gasse et al., 1980; Servant and Servant-Vildary, 1980; Rognon, 1987). Streams were characterized by a braided regime, but perennial flow probably did occur in the major rivers (Talbot, 1980). In the Sahel following an arid phase recognized in the Chad basin between 11,0000-10,000 BP (Servant et al., 1976), as well as in other places (Rognon, 1987), the climate became most humid in the period 10,000-8,000 BP. Fluvial sediments from this period are characterized by silt and clay accumulation at many sites. The larger rivers of the Sahel were meandering streams with high concentrations of free, suspended sediments. It is understood that precipitation during this period was not only more abundant but also more evenly distributed throughout the year (Sombroek and Zonneveld, 1971; Servant, 1974; Rognon, 1976, 1987; Maley, 1977b; Talbot, 1980).

Lezine & Casanova (1989), interpreting pollen and hydrological data of past climates in tropical west Africa report (in addition to the two abrupt major vegetation changes already mentioned (i.e. at 9,000 BP, wetter vegetation; and 2,000 BP drier vegetation), that the front of moist air progressed from south to north during the early Holocene. The South-Sudanian ecoclimatic zone was almost permanently under the influence of humid air masses. The southern margin of the Sahara recorded major hydrologic phases at 9,500-7,000 BP and 4,000-2,500 BP. Moisture conditions at 6,300-4,500" BP were well defined in the Sahelian zone, but only correspond in the southern Sahara to a mosaic-like pattern, reflecting local morphological settings. In the three ecoclimatic zones, the first evidence of run-off was recorded as early as 12,500 BP, whereas modem arid conditions appear around 2,000 BP. Regular drizzles on a year-long basis occurred from 12,500 to 7,500 BP. The beginning of strong seasonality in precipitation and the appearance of a marked dry season occurred at 7,500 BP.

Lezine et al. (1990) report on an early Holocene humid phase in western Sahara according to data from a section from the Chemchane Sabkha (Mauritania, 21°N, 12°W + 256 m). The age of the lowest part is around 13,500 BP. Maximum lake expansion was between around 8,300 to 6,500 BP, which is recorded in a girdle of stromatolite carbonates. These document a lower-salinity lake concomitant with a general establishment of Sahel-Sudan vegetation. The Chemchane area is the northernmost (21°N) occurrence of humid-phase elements that are related to maximum intensity of monsoon activity during the early Holocene.

Jakel (1987) summarizes the investigations he and Geyh (Geyh and Jakel, 1974) have been carrying out on the climatic fluctuations in the Central Sahara of the late Pleistocene and Holocene. They have constructed a histogram of humid and arid climatic phases in the Tibesti and Central Sahara, reconstructed according to changes in the depositional character of the formations dated by ^{14}C. Layers of a lake which according to ^{14}C dated from 16,000 to 13,500 BP, means that a humid climate existed during this period (although Geyh and Jakel (1974a,b) maintains that these dates are unreliable, as no pluvial deposits are known from the Sahara; it is the opinion of the present author that it was a northern shift of the western African monsoon). Between 12,500 and 11,500 BP, layers deposited in a shallow lake were found. The sequence of lake layers from 9,800 to 7,300 BP a continuous humid period is evident. From 7,300 to 6,000 BP the area becomes arid. A humid period follows from 6,200 to 500 BP when a lake reappears. The period between 5,000 to 4,200 BP was very dry . An oscillation towards increased humidity occurred again between 4,200 and 3,700 BP. From then onwards the climate became more and more arid, with an apparently rhythmic pattern of fluctuations. An exception to this pattern occurred between 2,000 to 1,200 BP.

This and Lezine's (Lezine and Casanova, 1989; Lezine et al., 1990) data show quite evidently the negative correlation between the Circum-Mediterranean region (influenced by the westerlies) and this area (mainly influenced by the trades), although an extreme shift may cause anomalies, the influence of the Atlas mountains seems to play a similar role as that of the Himalaya. regarding western China.

In Southern Mauritania in the SW Sahara margin Michel (1980) found by correlating sediments and climatic changes that at ±900 BC it was relatively humid, whereas since ±500 BC dry climatic conditions have prevailed. However, the latter period has been punctuated by minor variations, e.g. the Middle Ages were relatively humid.

Petit-Maire (1980) investigated Holocene biogeographical variations along the NW African coast (28-19°N) where present rainfall is 30-50 mm/year. She found:

4,000-2,000 BC, Nouakchottian transgression.

Around 1,500 BC, the association of steppe vegetation and animal life with important cultural neolithic traits as far north as 21°N implies climatic conditions very different from the hyperarid present ones, with a minimum annual rainfall of*150 mm (28 mm today).

By O BC/AD, the steppic belt had shifted back south, may be quite rapidly.

At 5,000 BP the West African Sahel was still under a humid climatic regime which had begun around 7,000 BP and was to last until 4,000 BP. In south-central Mauritania the annual rainfall in the period 5,000-4,000 BP appears to have been twice as much as today, based on floral and faunal evidence (Munson, 1974; Talbot, 1980). An arid phase occurred from 4,000-3,500 BP, during which monsoonal precipitation had declinedsufficiently to allow for local aeolian dune formation (Servant, 1973, 1974; Maley 1977a; Talbot, 1980). This arid interval was followed by the last major humid period in the West African Sahel, which lasted from 3,500-2,000 BP (Talbot, 1980). A more arid climatic regime, albeit with fluctuations, has been predominant in the region ever since.

Weisrock (1980), investigating the littoral deposits of the Saharan Atlantic Coast since 150,000 BP found that from 28,000 to 13,000 BP a very dry period (expansion of great Ogolain ergs, while from 9,000 BP in the south (Mauritania) and from 5,300 BP in the north (Morocco) the increase in moisture was general. This moist period ended somewhat earlier in the south (about 1,000 BC) than in the north (about O BC/AD). The process in historical times of becoming arid occurred earlier in the south, is more severe there and continues today, while in the north there are still oscillations, with man certainly accelerating the process of aridity.

Pollen analyses were carried out on Holocene lacustrine deposits sampled every 10-20 or 30 cm in a section of about 7.80 m at Tjeri (13°44'N- 16°30'E) near the center of the great Palaeochad (Maley, 1977 b). The comparison of the Sudano-Guinean element and Sahelian element curves showed that the periods of climatic optima (relatively wetter phases) are in general out of phase with each other (see Fig. 10). It seems also that during the Holocene the Sahelian climatic optima have always been synchronous with the w arming periods, and the deteriorations with the cooling. Indeed the trends of the Sahelian curve at Tjeri - the amplitudes of variations being different - correlated well with the trends which appear on some curves portraying the evolution of the temperature on the Northern Hemisphere, such as that of Camp Century in Greenland. At present nearly all the rains falling in the Sahel zone are monsoonal in origin. In summer (July-August) these rains reach the Hoggar and Tibesti mountains. But, over the Central Sahara the heaviest rains fail chiefly during the intermediate seasons of spring (March-June) and autumn (September-December).

The study of cloud formations over the Sahara also confirms the importance of the intermediate seasons. The rain of interseasons are linked with the tropical depression, also called Sudano-Saharan depressions or Khamsin' depressions in Eastern Sahara. Rains of this type are often fine and continuous, whereas the monsoon rains are stormy. At present this depression occurs rarely in winter except in Western Sahara. On the other hand, these depressions are absent in the height of summer when the ITCZ reaches the Sahara. Schematically the synoptic situations are as follows:

(1) influx of polar air in the middle or upper troposphere above the Sahara along shallow troughs in the upper westerlies, and

(2) frequently, ahead of these cold troughs, undulations occur in the ITCZ with brief invasions of humid equatorial air. The undulations of the ITCZ could be due to the action of boreal cold troughs or to surges of monsoon caused by perturbations travelling in the Southern Hemisphere. The depression Created by the cold air aloft favor the advection of the equatorial humid air. During this part of the year the movements of these depressions are north-eastwards or eastwards. The advection of humid equatorial air is essential for the formation of rain from these depressions.

At present in winter, the scarcity of these depressions over the Sahara can be explained by the fact that at this time of the year the ITCZ is situated at very low latitudes. But when there is interaction between cold troughs and the ITCZ, the trajectory of depressions remains chiefly over the Sudan and Sahel zones.

From the spatial correlation of annual rainfalls since the beginning of the century for Africa north of the equator (Maley, 1977b) it was found that in some years there is a contrast between the Sahel and Central Sal-mean zones in respect to the amount of annual rainfall. (This is less clear for Eastern Sahara, probably due to the lack of sufficient data.) The Central Sahara is influenced during periods of relatively low temperatures by tropical depressions (this situation was more frequent in the Quaternary period), while during periods of higher temperatures it is dominated by monsoonal rains: During the early Holocene the climatic models resembled, most probably, the annual variations of the present. Whether the region should be influenced by the tropical depressions or by the monsoonal depends on the variations in the temperature over the Northern Hemisphere as well as by the variations in the general atmospheric circulation. Thus, during periods of relatively low temperatures there occurs a penetration of polar troughs at high altitudes, which causes the formation of tropical depressions. These are related to the occurrence of large amplitude waves in the upper westerlies. On the other hand, during more warm years or periods, these polar troughs are rather scarce, and the circulation of the upper westerlies is more restricted, which leads to a diminution of tropical depressions, and the extension of rnonsoonal rains. The interplay between these two systems remains Only partial since both are also linked to the activity of the ITCZ (Maley, 1977b).

Studying environmental responses to climatic changes in the West African Sahel over the past 20,000 years, Talbot (1980) deduced the following changes:

20,000-12,000 BP, arid;
12,000-11,000 BP, humid;
11,000-10,000 BP, arid;
10,000- 8,000 BP, most humid;
8,000-7,000 BP, arid;
7,000- 4,000 BP, humid;
4,000- 3,500 BP, arid;
3,500- 2,000 BP, humid;
2,000- 0 BP, arid.

Some remobilization of older dunes also seems to be occurring under present day conditions.

Apart from dunes of anthropogenic origin, which appear almost anywhere in the Sahel where human interference has greatly reduced the natural vegetation cover, there are active dunes at places along the transition zone into the Sahara.

Petit-Maire (1987) summarizes her findings in the Tauodenni Basin, northern Mali, Western Sahara as follows:

40,000-20,000 BP, humid phase, with Aterian prehistoric artifacts. Upper Pleistocene arid phase.

10,000 BP, marshes and lakes form in the area.

9,000 to 7,000 BP, an important lacustrine phase; prehistoric man and big mammals were abundant. The minimum precipitation amount is 300 mm. During this period the Niger river flowed into a large interior delta north of Timbuktu. During its major floods its effluent may have reached an area 300 km north to its present curve.

6,500 BP (approximately), a drier episode is testified to by an evaporitic phase.

5,500 BP (approximately), a new Lacustrine episode is recorded; the water level is, however lower than during the Early Holocene; there are no mega fauna or human settlements north to the 23rd parallel; at 4,500 BP there are sheet flood sediments, evidence of an arid zone “wadi” regime.

3,800 BP, the lakes dry up and a sabkha system dominates; Man groups around water holes and wells in the largest depressions.

3,000 BP, aeolian processes start again to reach the most severe arid conditions on earth.

4.3 Europe

4.3.1 Northern and Western Europe

Warm periods during the lower Holocene coincide more or less with the Boreal and Atlantic stages of the classification by Blytt-Sernander for the European Holocene (Roberts, 1989):

- Boreal Period, warm and dry -9,500-7,000 BP
- Atlantic Period, warm and wet -7,000-5,000 BP

At ca. 9,000 BP the climate becomes warmer and is characterized by forest vegetation dominated by hazel (Corylus) and pine. This is the Boreal period of Blytt (1876) Sernander (1908) and Hazel-Pine period of Iversen (1958). This type of vegetation continues up to ca. 8,000 BP. During this period the forest vegetation becomes more variable. Among the trees that appear are the lime (Tilia), oak (Quercus) and alder (Alnus). This is the Atlantic period of Blytt-Sernander and the Older Lime period of Iverson (op. cit.) This period lasted until 5,000 BP. After this, the pollen assemblage starts to show the introduction of domesticated plants, as farming communities settled in the region. According to Iversen, the latter half of the Older Lime period is believed to be the warmest period of the post glacial periods, the temperature was at least 2°C higher than today.

Summer temperatures in the above period were 2 to 4°C higher than today in the continental interiors of North America and Eurasia according to a climatic model (CohmaP, 1988). In Britain temperatures rose sharply around 10,000 BP following the cold Younger Dryas stadial, based on evidence from fossil beetles (coleoptera) studied by Coope (1985).

The Chalcolithic period from around 6,000 to 5,200 BP was a world-wide colder episode most probably a mini-glacial. It was followed by a warm period which began around 5,000 BP. This period is known as the Sub-Boreal in Europe. Although phases as warm as any since the last ice age occurred during this period, the variability of the climate was greater than before (Lamb, 1982). This was followed by another cold period extending from around 5,000 to 4,200 BP. This humid and possibly also cold period, corresponds to the Early (Lower) Bronze Period in the Levant as can be seen on the base-section.

A good correlation can be seen between the Levant base curve and the curve of glacier fluctuations in Scandinavia during the last 9,000 years presented by Karlen (in Starkel et al., 1991), see Fig. 11. The curve is based on studies of sediments from lakes downstream of small glaciers and radiocarbon dated from organic material found in the sediments. The Lower Neolithic from 8,500 to 8,000 BP, which was suspected to be warm and wet in the

Levant due to monsoonal influence, is marked by a period of retreat of the glaciers (i.e. a warm period), while the Middle Neolithic cold period is a period of the advance of glaciers. The Upper and Upper Middle Neolithic warm period is also characterized by a retreat of the glaciers. The Chalcolithic cold period in the Levant, is also a period of advances of the glaciers (with retreat around 6,000 BP). The Upper Chalcolithic warm period, in the Levant, is again characterized by a short period of retreat, while the Early Bronze cold period is also a period of general advance of the glaciers with two retreats of short periods. The Middle Bronze warm period, in the Levant, is characterized by a retreat of glaciers for a rather long period which ended at the Iron Age cold period. There is again a short period of retreat, corresponding to the late Iron Warm period. The early Roman maximum cold period, in the Levant, is characterized by a general advance of glaciers. Around 300 AD there is a retreat of glaciers which, as will be discussed in the case of England, is characterized by the rise of the sea level. The Late Roman cold spell, i.e Byzantian in the Levant, is marked by an advance of glaciers. The Moslem warm spell which according to the base-section started at 600 AD and reached its maximum at ca. 900 AD is marked in Karlen's (Karlen, 1991) curve by a retreat of the glaciers over a limited area and during a short period of time around 800 AD and another short period retreat, but on a wider area around 1,000 AD. It is suggested that there is a correlation between the Scandinavian glaciers' advance at ca. 1,200 AD and the Crusaders short cold spell, observed in the Levant, at around 1,000 AD, and the three later advances (Fig. 11) with the little Ice Age.

According to Karlen (in Starkel et al., 1991), "A somewhat higher summer temperature during the early Holocene caused some glaciers to melt away entirely and reform first about 2000 years BP". This conclusion to the opinion of the present author is very important for the interpretation of the isotope curves of the cores of the glaciers, as this means that water of melted glacier ice could have changed the composition of underlying ice layers, perhaps causing a hiatus in sequential sections.

According to Roberts (1989), "The British Isles remained largely treeless until the start of the Holocene. By this point in time, however, deciduous trees had begun to expand out of their southern refugia and ground conditions had been prepared by pioneering plant colonizers. Consequently, once climatic conditions permitted, there were few checks to a rapid spread of tree species across Europe, other than their own rates of dispersal. Between 10,000 and 8,000 yr BP tree species moved in to occupy suitable vacant land... After two turbulent millennia of vegetation change, European phytogeography looked fundamentally different. The boreal forest was pushed northwards to Scandinavia and northern Russia, tundra and steppe were all but removed from the scene, and the dominant vegetation type was now mixed deciduous forest. But if the distribution of plant formations had become essentially modem by 8,000 BP, their composition was not. This is immediately obvious from individual pollen diagrams, in which the characteristic feature is the continued arrival and rise to dominance of new woodland taxa. After the pioneer woods of birch and pine, the first deciduous trees to arrive in northwest Europe were hazel *(Corylus avellana)* and elm *(Ulmus)*, both of which expanded without environmental constraints in less than 500 years" (p. 71-72P. Taking the abundance of the hazel as an indicator of warmer conditions and correlating it with the base section (Fig. 12) one can see that during the lower Holocene (namely the Neolithic) the climate was rather warm, enabling the expansion of the Hazel's population. During the Chalcolithic and Early Bronze there is a reduction in the population most probably due to a colder phase. During the Middle and Late Bronze there is an increase in the numbers of this tree, while towards the Roman period there is a strong decrease, indicating colder conditions.

In England many Roman buildings built prior to 300 AD were covered by the transgressions of the sea which occurred just after this date (Thompson, 1980). This is an indication that the sea had retreated when these buildings were built.

The rise of the sea level immediately afterwards indicates a short warming phase around 300 AD. This was followed by a cold period which came to an end around 500 AD when a warming climate change affected the world, continuing until 1,500 AD.

The period 950-1,530 AD was exceptionally warm, except for a cold period (1,000-1,200 AD) named the Crusader Period, reaching a maximum in the period 1,313-1,344 AD. However, according to a graph by Dansgard (1985) showing oxygen isotope variations in Greenland glaciers, (from AD 300 Until the present; arranged to be read as equivalent to a temperature curve (Lamb, 1982), the whole period from 300-1,150 AD was warm except for a downturn from 400-500 AD. Around 1,150 AD a sharp drop in temperature occurred, marking the end of the long warm period.

The archaeology and the coastal change in the Netherlands is given in Thompson (1980). These changes were governed by the rise and retreat of sea-level, sediment being laid upon sediment and a vast stratigraphy originated, in which all environmental changes are documented. A sequence of changing landscape-patterns, a major framework of this sequence, is embodied in cyclic processes known as transgression/regression cycles. (Louwe-Kooijmans, 1980). During transgression phases, estuarine creek systems were gradually extended and tidal flat areas were enlarged. This was followed by periodic sedimentation, during which creek systems were silted Up and tidal flats changed into salt marshes. Mostly, this sedimentation phase is included in the transgression part of the cycle, but it represents, in effect, the first part of a regression phase which culminated in widespread peat formation.

Correlation between the transgression and regression cycles and the base-curve (Fig. 13) from bottom to top, shows at around 5,800 BP a transgression (CII) which does not correspond with the Chalcolithic cold climate shown on the base curve. This, however, could be the end of the Upper Neolithic warm phase. On the other hand, there is a regression at around 5,400 BP which is still in the Chalcolithic cold phase. There is, however, a pronounced transgression (IC III) during the Upper Chalcolithic warm period and two regressions during the Early Bronze, interrupted by a transgression (C IV). The Middle and Late Bronze warm are represented most probably by two transgressions (C IV b and Do) interrupted by a regression. The Iron Age cold period, by a transgression (D la). The Hellenistic period is marked by a regression while the Early Roman Cold Period there is a transgressional phase (D Ib) which may point to the fact that there was a hiatus in the cold spell starting about 300 BC. The Late Roman-Byzantine cold period is marked by two regressions with a transgression in between, marking, perhaps, a warm period around 300 AD.

The Moslem period warm spell is marked by a transgression (D IIIa) with a short regression marking the Crusader's cold spell, followed by a transgression.

Thus it can be said that in general this diagram corresponds with the base-curve, and shows that the major changes in the Mediterranean area correspond with sea-level variations in the Atlantic.

4.3.3 Central Europe

Black Sea surface-level changes compiled by Ghenea and Mihailescu (1991) show in general a low level at the base of the Neolithic period, a gradual rise to the end of the Early Bronze and a decline which reaches its maximum in the Helenistic-Early Roman Period (see Fig. 14). A rise in the Sea level characterizes the end of the Byzantine and the Lower Moslem Period. A small descent marks the Crusaders period. In general it can be said that this curve compares only in a very general sense with the Levant base-section.

Chernavaskaya (1990) investigated the climate of Eastern Europe in the historical past on the basis of analyses palynological assemblages of high bogs in the forest zone. From these he was able to reconstruct temperature conditions of the summer-time period in the south-western part of eastern Europe over the last 3,500 years. He noted that 3,400 to 1,500 years ago was cooler, the lowest temperatures being observed in the 1st to 5th centuries AD. Northwards in the Ukrainian and Polessia regions, during the first centuries AD there was a drop in temperature in the 1st century AD. A relatively warmer period followed during the last 1500 years, with a colder period observed from 900 to 150 years ago. The latter coincides with that of the Little Ice age. During this time there were century-long periods when the amount of precipitation was 10-1570 less than the present. Such were observed in the northern part of the region with a time shift from the north-east (14-15 centuries), to the south-east (16-17 centuries) and the south-west (17-18 centuries). During the Medieval Optimum (13th century) dry periods were more typical in the southern regions.

The observations of low temperatures during the 1st to 5th centuries AD, are in accordance with that concluded for the Levant.

4.4 West Asia

Various estimations and calculations for Central Europe (more precisely for southern Poland) from the time of the maximum extent of the last ice sheet to the present time have made it possible to construct curves showing variations of precipitation, evaporation and runoff. Annual precipitation rates fluctuated from around 250 mm or less at 18,000-14,000 BP to about 550-650 mm from 8,500 BP onwards. Annual variations were probably much higher during the unstable cooler phases, as stressed by Lamb (1977). The warming of the climate, formation of a dense vegetation cover and summer rains caused a general shift from winter and snowmelt runoff in the cold climate to the summer floods caused by continuous rains. The frequencies of extreme events under conditions of a dense vegetation cover and later under increasing deforestation caused a complicated pattern of phases with overloaded and underloaded rivers (Starkel et al., 1991).

4.4 West Asia The climate of west Asia is dominated in winter by the polar continental air mass (PCAM), causing a northerly flow in the lower troposphere which comes from middle-high latitude cold dry air. In summer the region is dominated by the tropical-subtropical oceanic air mass (TOAM) and a tropical continental air mass causing a southerly monsoon in the lower troposphere layer, bringing oceanic warm and moist air from the low latitudes. There are two types of summer monsoon, the southwestern and south-eastern, influencing different areas. Today the south-eastern dominates most of China, and from purely theoretical considerations it should have been the case for the last 130,000 years (An et al., 1991).

4.4.1 China

Zao et al. (1990) report on the Holocene stratigraphy and its reflection on the climatic and environmental changes as deduced from a section in a peat deposition 8 m deep in Qingfeng, Jinahu County, Jingsu province, China, By analyzing and carbon dating the sedimentary sequence, the fauna and the flora they arrived at the following stratigraphy from bottom to top:

"Phase 1. Peat swamp of the Preboreal period. In that time the vegetation was coniferous and steppe. The climate was arid and cold.

"Phase 2. Littoral swamp of Boreal period.The vegetation was salt meadow and needle and broad leaf mixed forest, the climate was dry and warm temperate, and the sea water affected this area for a time.

"Phase 3. Lagoon of the early Atlantic period. The vegetation was saltwater marsh, and pine bearing deciduous broad leaf forest. The climate was semi-humid and warm and the sea water invaded this area, forming a lagoonal environment in the middle to lower part of the intertidal zone.

"Phase 4. Bay of the late Atlantic period.The vegetation was saltwater marsh and evergreen and deciduous broad leafed mixed forest, the climate was humid and warm, and the Holocene transgression reached its maximum. In that time, the area studied became an open bay environment in the middle to lower part of the inter-tidal zone to the upper part of the subtidal zone. The sea water was warmer.

"Phase 5. Littoral lowland of the early Subboreal period. The vegetation was salt meadow and deciduous broad leaf forest. The climate was semi-arid and warm temperate, and the sea water retreated from this area for a time.

"Phase 6. Desalted lagoon of the late subboreal period. The vegetation was of needle and broad leaves mixed forest, the climate was humid and cool temperate, and the sea water affected this area again. "

During the Atlantic Period the sea level was about 2 m higher than that of the present.

Yang and Xie (1984) report on sea level changes in East China over the past 20,000 years. They correlate it with climatic changes. During the Holocene they see the minimum levels of the sea as correlated with the maximum cold climatic phases observed in China, which were from 8,700-7,700 BP (with a peak at 8,200 BP), 6,700-5,700 BP (with a peak at 5,800 BP), 3,100-2,700 BP (with a peak at 3,000 BP), 400-100 BP (with a peak at 200 BP). During the cold periods the sea level fell 2-4 m. During the warm periods there was a rise. Three major rises are discerned: from 10,000 to 8,300, from 8,000 to 7,000, and from 6,000 to 5,500 BP. The sea level and climatic changes during the last 2,000 years were also investigated.

Ice cores from the Tibetan plateau of China (Thompson et al., 1989) show pronounced climate changes during the Holocene. However, their discrete 1000-year time averages of the components of the core restrict the possibilities of correlation with the Levant base section. In general it can be seen that from around 10,000 to 9000 BP (Base Neolithic in the Levant) the d^{18}O is still low relative to the Holocene record, indicating still colder climates than later. During the lower Neolithic between around 9,000 to 8,500 BP the d^{18}O composition increases, due to warmer conditions. This period is also marked by an increase in the salt contents of the ice, which means a drying Up of lakes and swamps in the adjacent desertic areas. There is a maximum increase of d^{18}O around 6,000 BP, which correlates most probably with the warm upper Neolithic period in the Levant. From 6,500 to 5,000 BP there is again a reduction in the d^{18}O composition, which correlates well with the Chalcolithic and Early Bronze cold periods of the Levant. The period after 4,000 BP until 700 BP is marked by an increase in the d^{18}O composition. This correlates in its lower part with the warming and desiccation of the Levant during the Middle Bronze, but there is no sign of the Roman wet period. On the other hand the Little Ice age is well marked. According to Thompson et al. (1989) “A striking feature of the d^{18}O record is the extreme less negative values (which suggest warmer temperatures) of the last 60 years; decades with highest values are the 1940s, 1950s, and 1980s.” This is connected with the greenhouse effect.

Shi et al. (unpublished report) have used the data from the Dunde Ice Core to investigate the Paleoclimatic and Environmental Features in the Mid Holocene of China. They put emphasis on finding the temperature differences between the various periods and the present. Their main conclusions are also based on the interpretation of data from inland lakes, paleosols in loess and the sandy desert area, sea level variations, and a review of archaeological and palynological data.

The trends and features of climatic change in the past 5000 years in east and west China as revealed by comparing historical data from eastern China with data from the Ice core in Dunde glacier in Tibet are discussed by Yao (1990). His main conclusions are that there is a correlation between the historical records and the data from the ice core. Temperature, both in East China and the West China, increased from about 5,000 to 3,000 BP and decreased after 3,000 BP, reaching a minimum around 1,000 BP. After 1,000 BP temperatures began to increase. Curves of climate change for both west and east china show that the little Ice Age was not the coldest period in the past 5000 years, but was the coldest period in the past 500 years. There is a time phase difference between the two regions.

Climatic changes in the Tibetan plateau were also investigated by Fu-Bau and Fan (1987). They analyzed the peat and sand gravel layers distibuted in the fluvial bog and marsh plain along the mountain foothills by studying cross-sections near interior lakes, glacial activity and landform changes. Correlating these results, especially the curve of temperature variations, (Fig. 15) with the base curve of the Levant one can see (from bottom to top) the following:

The Wumandung interval, from 10,000 to 7,500 BP (which according Fu-Bau and Fan (1987) was slightly cold and dry) corresponds to the base Neolithic.

The warm and moist period from 7,500 to 5,500 BP may correspond to the Upper Neolithic and Lower Chalcolithic warm periods in the Levant.

The warm but dry period from 5,500 to 4,700 BP may correspond to the Chalcolithic and Early Bronze periods in the Levant, which were cold and wet. As discussed earlier this can be seen also from the $d^{18}O$ record from the same region (Thompson et al., 1989). (Indeed the curve presented by Fu-Bau and Fan (1987) shows a decrease in temperature at that period and warm and moist to 3,000 BP).

The Qilongduo interval (yoli period) from 4,700 to 3,000 BP represents the warm dry Middle and upper Bronze periods of the Levant. The cold Ice age I, from 3,000 to 2,500 BP may correspond to the Iron age cold and wet period of the Levant. The slightly warm period from 2,500 to 2,000 BP to the Persian warm period, while the Mid-Neoglacial interval (Ice age II) corresponds well with the Roman cold wet period of the Levant. The Dawelong interval from 1,500 to 3000 BP considered mild, corresponds with the Moslem warm dry period of the Levant. A closer study of the corresponding curve (Fig. 15) shows indeed that there was a steep rise in temperature around 1,500 BP. A decrease around 1,000 BP corresponds with the crusaders cold period. The little Ice Age, which is dry and cold, is observed around 300 BP.

Lin and Wu (1987) summarize some features of climatic fluctuations over the Qinghai-Xizang Plateau of Tibet. According to their investigations the climate of this plateau had attained its present general features by the end of the Middle Pleistocene when the elevation of the area was already 3,000 m. The average temperatures were 4-8°C higher than the present. Today the average elevation of the plateau is 4,000 m. According to peat layers dated by ^{14}C, the period from around 7,000 to 3,000 BP, especially during the period from 5,000 to 6,000 BP, the annual mean air temperature was 3-5°C higher than the present. From around 3,000 to 1,500 BP, dated by ^{14}C, the glaciers advanced; however, it was not continuously cold. It is obvious that there was a short warm period toward the end of the second century AD, through the first half of the third century AD (Roman short warm period and transgression in the west, Issar et al., 1990. From the 6th to the late 12th century AD there was a marked warm period; there were many forests around the city of Lhasa. By the 17th century AD, however, the climate was the coldest of the past 1,000 years, the mean temperature being 1°C colder than today. During the 17th to 19th centuries AD there were glacial advances in many sites. Since the mid-19th century AD it has become warmer over the plateau. It was warm and wet at the beginning of the 20th century AD, and is becoming drier. After analyzing historical and meteorological data, Lin and Wu (1987) arrived at a 2.6-3.6 year periodicity of flood and drought events over the plateau, their meteorological and proxy data pointing to a "quasi bi-ennial pulse, and two principle cycles, corresponding to sunspots, at intervals of about 11 or 22 years." In addition, they found some indications of an approximately 30-year cyclicity (Bruckner Cycle).

Lin (1990), investigating the climate change in the Xizhang plateau in Tibet during the last 200 years, found that since the 19th century major snowstorms occurred 15 times, from which those in 1,828-1,829, 1,887-1,888 and 1,927-1,928 AD were the most severe. In the past 50 years (except for a severe storm in 1,967-1,968 AD) snowstorms have become lighter. The droughts and floods appeared alternately, the drought period gradually getting longer. The rainfall in 1,983 AD was the lowest during the last 100 years. In general the climate in Tibet is becoming warmer and drier.

Zheng et al. (1982) studied the Quaternary Geology in Kunlun Mountains, north to the Qinghai-Xizang Plateau of Tibet and South to the Pamir basin. According to ^{14}C dating, he found evidence of neo-glaciation in the periods 3,983 to 3,522 BP and 2720 ±85 BP. During the little Ice age there were 2-3 glacial advances. presently most glaciers are retreating.

Cui and Song (1991) investigated the Quaternary periglacial environment in China. During the transitional period to the Holocene, 13,000 to 11,000 BP, this period was drier and colder period than today. The period from 10,000 to 9,000 BP had a cool to temperate/semiarid climate compared to the former. From 9,000 to 8,000 BP it was cold and dry. From 8,000 to 6,000 BP it was warm and semi-humid. From 6,000 to 5,000, dry and cold conditions prevailed. From 5,000 to 3,000 BP, it was warm and semi-dry. From 3,000 to 1,500 BP it was relatively cool and wet. And from 1,500 BP to the present, semi arid and cold conditions have prevailed.

Zhou et al. (1991) report on the changes of permafrost in China during the Quaternary. During the Holocene from 10,000 to 3,000 BP the air temperature in eastern China was 3-5°C warmer than the present. At 7,300 BP the air temperature was 1°C warmer in the Tianshan and 2-4°C warmer in the Qinghai-Xizang plateau.
About 3,000 BP it became cooler again, and after a series of fluctuations the present pattern of permafrost distribution was formed.

The levels of the lakes in north-western China during the Holocene, followed the pattern which was described earlier for the Upper Pleistocene. Thus several lakes which lie in eastern Kunlun and Xijang deserts and are under the influence of the westerlies maintained a high water level up to 12,000 BP. Then the water levels fell until 8,000 BP. Since then they rose again, attained the highest level during 6,500-6,000 BP and then descended sharply (Jijun et al., manuscript).

Wang and Li (1991) report on a general retreat of the surface levels of Qinghai and Daihai lakes, lying on the north-western margin of the east Asia monsoon area, from 4,000 to 3,000 BP. On the other hand, the lakes in the Yunnan plateau, situated in the southwest monsoon area, have opposite trends of evolution. During the last glaciation they formed large water bodies but during Holocene the lakes retreated. In Xinjiang Province, which is influenced by the westerlies, the lakes were high during the Last Glacial (as well as Holocene optimum) but regressed during the past 2,000-3,000" years. In the middle and lower reaches of the Changjiang River, lakes disappeared and were replaced by a fluvial plain. They were filled with loess-like deposits during the Last Glaciation. Lake Taihu was formed around 2,500 BP, while Lake Poyang was formed during the Han Dynasty, around 2,000 BP. This situation is contrary to the universal regression of lakes during the past 2,000-3,000 years.

Based on data from historical documents cited by Fang (1990), several sharp temperature fluctuations have occurred in historical times in China. The first cold stage was around 3,000 BP, with temperatures l-2°C cooler than today. This must have lasted more than two centuries. The second cold period occurred between 100 and 600 AD; the third, 1,050-1,350 AD; the fourth, 1,600-1,850° AD, with coldest temperatures around 1,70(1 AD. The warm spells were about 700-0 BC, 650-1,000 AD (during this period the mean temperature was 2-3°C above that of the present) and around 1,400 AD. In colder climatic periods, the formerly scarce winter thunderstorms appeared frequently. These are associated with rare frontal movements of very cold air masses from central Asia. Such occurred in the periods 100-150, 300-500, 1,100-1,300 and 1,600-1,900 AD. The longest break of no winter electrical storms was 600-800 AD. Dust storms emerged frequently during the cold periods.

Based on the analysis of a drought-flood index since 0 BC/AD, five drier climatic intervals lasting longer than 100 years have been identified in northern China (Zheng, 1985, as cited by Fang, 1990). They are 0-100, 300-630, 1,050-1,270, 1,430-1,550, and 1,580-1,720 AD. The driest climatic periods in northern China, especially in the region 30-45°N, coincided with colder climate, i.e. the strengthening of the Siberian anticyclone and southward shift of the polar front. It was wetter in the intervening periods. The longest wetter climatic period since 0 BC/AD occurred in the period 810-1,050 AD. The records of the floods of the Yellow river show four periods of low frequency which coincide, more or less, with the colder and drier periods. They are 150-650, 1, 120-1,280, and 1,680-1,880 AD. The longest break of no floods was 153-637 AD according to Fang, 1990), desiccating 29 lakes, of which 27 totally disappeared during the low flood period.

A close correlation between climate changes and the Chinese migration in historical times is suggested by Fang (1990). His study is based on a critical survey of the historical data on Chinese migrations, a literature survey on information about Chinese lake evolution, the history of agro-cities in the Chinese deserts and flood reports of the Yellow River. The main migratory events coincided with social and political unrest as well as with invasions from the north by the mongols. All these events, the author of this document believes, were connected with severe climatic changes. According to Chinese studies (Hanyong and Shanyu, 1984; Wenlin, 1988), as a result of these events the percentage of Chinese population settled in southern China changed from about 17% of the total Chinese population two thousand years ago, to about 65% at the beginning of the 19th century AD.

The Chinese migrations according to Fang (1990) were in phase with the cold climate phases. These occurred in 1,000 BC, then 0-50, 300-600, 1,100-1,300 and 1,600-1,750 AD (Fig, 16). The first recorded migrations at 1,000 BC (not shown on the figure, was that of the Zhou (Chou). They were western nomads who settled in present day Northwest China, moved eastward, overthrew the rule of the Shang dynasty and established the Western Zhou Dynasty. During this period there also occurred an advance of the glaciers, and a 100-300 m fall of the snow lines in the mountains of Tibet. Some sand dunes were also reactive and, some lakes in the west and north of China became more saline.

During cold and dry periods there occurred also a shift southward of the agro-. husbandry boundary. Thus in 300-500 AD the boundary was 200-400 km south of that in the former (and following) warmer and more humid periods.

According to Fang (1990), the analysis of the records of productive outputs and rainfalls for the last 30 years in the Inner Mongolia Autonomous Region suggests that there is a linear relationship between summer precipitation and herbage outputs. Low summer precipitation reduces the growing period of herbage, and thus the carrying capacity of the plain drops, causing famine to livestock and to the human population dependent on them.

The general conclusion of the author of this document is that a greenhouse effect will be positive for China's agriculture as it is would strengthen the monsoon, and thus the agricultural regions of china would become more warm, and more humid.

Correlating these events with the Levant base-section, it can easily be seen that the cold dry periods discussed above correspond with the cold wet periods in the Levant, while the warm wet periods of China corresponds with warm dry periods in the Levant. The 1,000 BC period is that of the Iron Age, followed by the Persian period of warmth and dryness which started to change in the Middle East by 300 BC to a colder and more humid period, but had no impact until 0 BC/AD. This continued during the entire Roman cold period with a short warm interval around 300 AD. The Moslem dry and warm period in the Levant was warm and humid in China. The Crusaders cold and wet period around 1,000 AD in the Levant was again a cold and dry period in China. The same happened again during the Little Ice Age.

Zhang & Wu (1990) have investigated climate fluctuations in China according to the changes in crop patterns. They found that from 8,000 to 3,000 BP mainly millet is mentioned. But later, in the Shang Dynasty, broomcorn millet was often depicted on oracle bones. Little is mentioned about other crops, which implies that by the time of the Yin people, broomcorn millet was the main cereal cultivated throughout the Yellow River Valley. Since broomcorn millet needs a shorter growing season than millet, there may have been a sharp drop in temperature and humidity during 3,400-3,000 BP. After this cold period, the climate warmed up again and millet was mentioned more frequently. During the warring states period (402-401 BC) people lived on millet and beans. Broomcorn millet was shifted or migrated northward.

Another mild climate indicator in historical times, according to Zhang and Wu (1990), is the plum tree (Mei in Chinese). This tree was quite widespread in the Yellow River valley during the 10th to 6th century BC, but practically disappeared from the Yellow River after the Song Dynasty (960-1,279 AD). The cold climate during the 12th century AD also influenced the cultivation of litchi, a typical Chinese tropical fruit, widely grown at least since the Tang Dynasty (618-907 AD) in south-east China. There the litchi were killed twice, in 1,110 and 1,178 AD by cold weather.

The influence of the Little Ice Age, according to Zhang and Wu (1990), can be seen from the records by the number of severe winters per decade. This can be deduced from records of frozen large rivers, lakes and wells. They show that from 1,500 to 1.978 AD there were 136 cases of severe winters. These were especially evident from 1,500-1,550, 1,601-1,720 and 1,831-1,900 AD. Periods of warm winters occurred in 1,551-1.600, 1.721-1,830 and 1.901-1.950 AD.

The variety of data from ancient documents, especially those of government officers who prepared reports for the emperor, provides a more detailed picture of climate of the 18th century (Zhang and Wu, 1990). These were correlated with measured data for the last 30 years. and thus temperature for the 18th century AD has been synthesized. These data show that the ratio of snow and precipitation days in the twenties and seventies in the 18th century AD was 10-15% higher than the present, the derived temperature being 1°- 1.5°C lower than that of the present. Thus, although the 18th century AD is regarded as a warm period within the Little Ice Age, these results show that it was colder than the present. Based upon records of blooming dates, it was deduced that, on average, they were six days later than that of the present, and accordingly it could be estimated that the climatic belt has moved southward in latitude by 1.5°. Comparing observations on the directions of winds made from 1.723-1,769 AD, it can be inferred that wind directions were also different from the present. Analyses of the records also show that the length of’ the rainy season at Beijing has become longer and longer during the last 274 years, the span from 1,947-1,978 AD being the longest. The shortest occurred in 1,746 AD. Dry periods occurred in 1,724-1,763, 1,814-1,863 and 1,924-1,943 AD.

During the last 500 years, many years of floods and droughts were recorded by local historians Zhang and WU (1990,) and Zhang and Ge (1990). They have analyzed these records and found that the main abrupt spots of drought/flood disasters are distributed in five time periods: 1,500-1,550, 1,600-1,650, 1,720-1.740 and 1.900-1,940 AD. Since most abrupt cases in this study correlated with warming trends, it may be deduced that in case of warming due to a greenhouse effect, the instability of drought/flood disasters in sensitive spots could increase.

The impact of climate changes on farming systems was also analyzed by the same authors. Rice-wheat farming was established in the Zhejing and Jiangsu provinces (lower valley of Chang Jiang) after the 12th century .AD. when the climate deteriorated by the 17th century AD, both wheat and rice had difficulty ripening, and the dates of sowing wheat was shifted to an earlier date to protect wheat from winter damage. By the end of the 18th and the beginning of the 19th century AD the yields of wheat and rice were again affected by a cold climate. Wheat-cotton farming, introduced to the middle and lower valley of Chang Jiang during the Ming Dynasty (14th-17th centuries AD), also suffered from the cold. In 1,628 AD the farmers were advised to sow the wheat before the cotton was harvested in order to avoid winter damage. Double rice crops were also affected negatively when the climate became colder. As citrus, which has been grown in China for the last 2000 years, is highly susceptible to cold climate, it can be used as a monitor of climate change. Since source areas of citrus as tribute to the emperor were recorded, it can be seen that during the 7- 10th centuries AD the places from where citrus was supplied are presently badly affected by frost. It must be concluded, then, that during those years the climate was wanner. The period corresponds well with the Moslem warm and dry period of the Levant.

The distribution of ramee and hemp was also used to detect climate change. From this it was concluded that the 13th century AD was the most warm period during the last thousand years (Zhang and Wu, 1990).

A study of the levels of Lake Daihai using the frequent shifts h-em agriculture to pasture shows that the level of the lake rose from 1,214,5 m above MSL in the year 1.881 AD, to 1,218 m in 1,910 AD. then fell to 1,212.5 m in 1.929 AD. In 1,970 AD it reached 1.224 m in 1,980 AD, 1.222.5 m and in 1,990 AD, 1.223 m above MSL.

Zhang and Wu (1990) have also investigated the paleo-climates of the Tibetan Plateau. Their main temporal divisions are the following:

7,000-3,000 BP. climatic optimum; warm and wet, annual mean air temperatures was 2-3°C higher than that of today.

2,900-1,600 BP. neo-glacial period: cold. frequent large glacial advances.

1,500-900 BP, warm period: warmer and wetter than at present.

890-150 BP, Little Ice Age; colder especially in the mid 17th century AD, about 1°C lower than the present.

140-O BP, last warm period: warm and wet at beginning of the 20th century AD. becoming progressively drier.

Based on official historical files. they found that floods and droughts appeared alternately during last 100 years: three rainy periods -- (1.883-1,906, 1,916-1.934, and 1.947-1,962 AD), and three dry periods -- (1,907-1.915, 1.935 -l.946 and 1,963-1,980 AD). The Tibetan Plateau according to these authors is a more sensitive area to climatic change than other areas of the same latitude of that region.

The main conclusions of these authors, in addition to those already stated, are that during a warming trend of climate more catastrophes may occur, and that the history of China shows that cold climates have harmed agriculture seriously.

Wang et al. (1990) have developed an Original method of “factor interpretation” which is based in the frost stage on factor analysis and correspondence analysis, enabling them to find the principal factors which control the pollen changes in a fossil pollen sequence. They then calculate relationships between the principal factors and the pollen taxa. In the second step, relationships are determined and the climatic factor is selected. Accordingly, the climatic factor’s weight sequence is constructed, relating the climatic factor to the samples. In the third stage, the climatic factor weights are transformed into parameters (temperature, precipitation etc.) in relation to ecological features. This method has been applied to deposits of peats in three Holocene lakes, two of which are from northeast China (Sanjiang plain, and Mount Changbaishan) and one from a small lake in southwest China (Mount Loujishan). The results suggest that the period from 12,000 to 10,000-9,000 BP had rapid warming. Yet, summer half-year temperature (SHT), was about 8°C lower at 12.0()(1 BP than at present , But at about 10,000 to 9,000 BP it reached present day temperatures. The average of SHT from 9.000 to 3,000 BP was 2-3°C higher than that of the present. The thermal maximum SHT, 3-4°C higher than at present, occurred between 7,500-6,000 BP in northeast China. There was a cooling period after this thermal maximum and an obvious deterioration of climate since 3.000 BP. Seven periods of cooling have been observed to occur during the Holocene. These were, (1) around 9,100-8,800 BP, (2) around 8,300-8,000 BP, (3) around 5,000-4,700 BP, (4) around 4.400-4.100 BP. (5) around 3,0()()-2.70() BP, (6) around 1,800-1.400 BP, and (7) around 500-300 BP.

Comparing these results with the Levant base curve (Fig. 17) one can see the following: There is no correspondence with the first cooling phase in China. Yet, from a closer study of the curves of Sanjiang plain and Mount Shangbaishan (Fig. 17), one can see that around 10.000 BP it was cooler relative to the younger layers, which may account for the base Neolithic cooler phase in the Levant. Period 2 in China may correspond to the Middle Neolithic cool phase in the Levant, with Period 3 to the Chalcolithic cooler phase in the Levant. Period 4. corresponds well with the cold humid phase of the Early Bronze in the Levant. The significant warming phase one can see in ail the curves around 4.000 BP is most probably the warm dry spell which caused the desertification of the Levant. The cooling at 3000 BP (Period 5) corresponds to the Iron age cold period. with Period 6 to the late Roman cold period. At the end of this period a warming period (the Moslem period. which in the Levant caused severe desertification) can be well observed on the curves. The same holds for the Little Ice Age.

A rather good correlation exists also between the Levant base-curve magnetic susceptibility (MS) curve for Baxia loess Plateau. where high MS means warm and humid and vice versa (Kukla and An. 1989). It can be seen that the low MS between 9.000 to 10,000 BP correlates with the cold period of the Base Neolithic, as does the low MS for 7.500 BP which correlates with the Middle Neolithic cold period. Except for those intervals, the whole

Neolithic to 5,000 BP was warm and wet in China and warm and dry in the Levant. From that time a cold and dry period covers China until around 3,500 BP, corresponding to the cold and wet periods of the Chalcolithic and Early Bronze in the Levant. Around 3,500 BP the climate became warmer, corresponding to the warm trend of the Middle Bronze in the Levant. A low MS (a colder spell) around 3,000 BP corresponds to the Iron Age cooling of the Levant, which after a short period of warming (the Persian period in the Levant) again shows low MS values, corresponding to the early Roman cold period. A high spell of MS, which may correspond to the warm spell of around 300 AD, is replaced by a relative cold spell, which may correlate with the Late Roman cold and wet spell in the Levant. This is followed by a warm spell which may reflect the Moslem warm and dry period of the Levant. The low MS of 1,000 BP (the Crusader cold period in the Levant) and the Little Ice Age also appear to be correlated.

The palynological study of the core of Lake Barkol in Xinjiang previously mentioned (Han and Yuan. 1990) gives a rather detailed picture of climate change in the western-most part of China (the region shielded Ii-em the monsoon and thus under the influence of the westerlies belt). It can be seen from the interpretation of pollen data that at the ^{14}C date of 12,530 BP (layer C6), according to Han and Yuan (1990) the climate was cold-wet, lake levels were high and the glaciers extended. At a ^{14}C date of 10,870 BP (layer W6) the situation was opposite: warm and dry, low lake levels and retreat of glaciers. Above this (layer C5) there was a short cold-wet spell. At a ^{14}C date of 8,970-9,370 BP (layer W5-2) and 8,190 BP (layer W5- 1) the climate was warm-dry. At 7,000 BP (C4) the climate was cool-wet. At ^{14}C date of 4,180-6,618 BP layer W4 was interpreted by these authors as warm-wet, but they mark a cold-wet climate and high lake levels at 5,000 BP (C3). At a ^{14}C date of 3,640-4,000 BP (layer W3) the climate was warm-dry, the lake dried Up completely and the glaciers retreated to more than 3.530 m. At a ^{14}C date of 2,640-3,270 BP (layer C2) the climate is again cold-wet, the lake level high and the glaciers descend to 2,830 m At a ^{14}C date of 1,200 to 2,310 BP (layer W2) warm-dry and low lake levels were to be found. In layers related to the 17th to 19th centuries AD, which consist of the upper 0.36 m of the section, there were three alternating periods of cold-wet to warm-wet.

The correlation with the Levant base-curve is rather good. Layer C5 corresponds most probably to the base cold Neolithic. while W5 (1 and 2) corresponds to the lower warm part of the Neolithic. Layer C4 may correspond to the Middle cool Neolithic, while W4 (which spans too long a period) may mask the warm and cold periods of the Chalcolithic. Layer C3 corresponds to the cold period of the Early Bronze, while the extreme warm-period W3 in which the lake dried up corresponds to the Warm Dry period of the Middle Bronze in the Levant. Layer C2 can be correlated with the Levant's Iron Age cold period. Layer W2 covers the span of the Roman cold and Moslem warm, and thus correlation is dubious.

Han & Yuan (1990) have also observed that all historical disastrous weather conditions have happened during the transition from cold to warm or warm to cold, periods. Almost all heavy snow events in the Barkol region during the last 300 years (1,679. 1.867. 1,882, 1.934, 1,951, and 1.973 AD) happened during such changes.

Based on tree ring and precipitation records these same authors found a marked trend of Warming and drying in the Xingjiang region. From the 18th to the 19th century AD mean precipitation declined by 12.5%, and the frequency of the occurrence of drought years increased by nearly 100%. Wet years have declined to one seventh of that of the 18th century AD. Data from 13 weather stations in the Xingjiang region show that the temperature is rising. Thus, Han and Yuan (1990) forecast that the warming Up effect will have a very negative effect on this region.

Wen & Qiao (1990) investigated the climatic sequence of the last 13,000 years in the Xingjiang region. They based their study on "dated sediments of Lake Aibi, the loess layers in the northern foothills of Tianshan Mountains and dune sections of Mosuowan and Damagou, as well as the river lake sediment section of yuetegan in Hetian country. " The sequence of climate change in this region was found to be the following:

13,000-10,000 BP - cold-moist to cool-dry;
10,000-8,000 BP - warm-dry:
8,000-7,000 BP - warmer-moist:
7,000-6,000 BP - temperate-dry:
6,000-4,000 BP - warmer-moist:
4,000-3,000 BP - Warm-dry;
3,000-2,000 BP - cold-moist
2,000 BP to present - warm- dry (in which around 1,500 BP it was cold-moist).

Yang (1991) investigated the relationship between the paleomonsoon and the Mid-Holocene climatic and sea level fluctuations in China. According to their evidence,Yang and Chen (1987) have shown that at about 8,000 to 6,000 BP the sea encroached and inundated the North China Plain and the Yangtze delta plain. The air temperatures attained a climax around 7,000 BP, but the climax of the sea transgression occurs about 1,000 years later. (Around 6,000, 5,000, and 4,()(X) BP great floods also occurred of the Yangtze River because of greatly increasing discharge.)

From 5,600 to 5,300 BP low temperatures and low sea level existed for some time.

There are also indications that from about 4,600-3,800 BP a period of high sea level occurred simultaneously with a time of heavy rainfall and high temperature. (This may be equivalent to the warm-dry period of the Middle and Upper Bronze in the Levan).

According to Chinese historical documents, in the dynasties of Yao, Shun and Xia (2,400-2,100 BC), prolonged heavy rainfall and great floods occurred in North China. This is the most wide-spread flood legend among the Chinese people. According to the Chinese ancient writings, many rivers overflowed their banks and spread unchecked. Some tributaries of the Yellow River reversed their courses as a result of the elevating of the lowest part of the drainage basin, and silting up of their lower channels. Due to the floods, people moved to higher grounds and began to learn to dig wells.

Li and Fang (1991) have investigated the Quaternary deposits and the environmental evolution of the Guangzhou Plain and Zhuziang Delta in south-eastern China. According to the analysis of sporo-pollen assemblages from deposits in this region the climate in the Early Holocene was generally warm and wet with some colder periods. It turned to warm-dry in the later period of the Holocene. For the upper Holocene they found warm and dry from about 5,000 to 4,500 BP, hot and wet ii-em about 4,500 to 3,500 BP, warm and a little dry from about 3,500 to 3,000 BP, and hot and wet around 3,000 to 2,500 BP.

Zhang et al. (1990) report on severe droughts since the beginning of the 19th century AD. The worst has been from 1,920 to 1,929 AD, reaching its extreme in the years of 1,928 and 1,929 AD. During this period some major tributaries of the main rivers of northern China, including that of the Yellow River, went dry. Lakes in inner Mongolia dried up. Wells and springs dried up. Sand desertification extended. On top of this, floods, frost, hail and war caused serious famine in northern China. Almost no crops were harvested. The famine victims had to eat the bark of trees, roots of grasses, soil and even human corpses. In 1,929 AD more than 6 million people died in the provinces alone. Many people migrated from the region, many women were sold to other areas to “man-traders”. The most serious disasters occurred in the provinces of Gansu, and Inner Mongolia, next came Shaanxi and Henan, followed by Shanxi, Hebei and Shandong provinces.

A severe drought also occurred between 1,876 and 1,878 AD, causing the death of about 13 million people in northern China, The most recent occurred in the 1,970’s AD, causing Zhang et al. (1990) to come to conclude that there is a 50 year cyclical occurrence of droughts. Each 150 years, the authors maintain, there are chances for a severe drought like that of the 1920’s.

An analysis of the connection between the floods and droughts in north China and the Indian summer monsoon have been investigated by Guo (1990). This was achieved by using a 116 year data set (from 1,871 to 1,986 AD). The correlation coefficient of 0.39 is significant at a 99% confidence level, although unstable (0.15 for 1,921-1,950 AD and 0.58 for 1,951 to 1,980 AD). The variations both of floods and droughts over north China and of Indian summer monsoon rainfall are closely related to the general atmospheric circulation, especially to the Indian Low and the Western Pacific High. When the north border of the Indian Low moved to the north, its correlation with the floods and droughts over northern China became insignificant. At this time the effect of the Western pacific High on the floods and droughts in north China predominated. A correlation was also found between low summer rainfall over north China and high sea-surface temperatures of the eastern equatorial pacific, in summer, autumn and winter.

Interpreting records of autumn rains in the Hunan Province during the last 1,300 years, Xiong (1990) concluded that there is a remarkable periodicity of 90 years, and that autumn rain change is closely related to cold/warm fluctuations. In a warm period autumn rain is lower, while in a cold period it is higher. Since the late 15th century AD autumn rains show a clear increasing trend. Thus, with forecasted higher temperature, the author of this document forecasts a reduction in the autumn rains in the future, which will result in an increase in autumn droughts for the Hunan Province.

A study on the abrupt dry-to-wet changes in the middle Yellow River Valley during historical times has been carried out by Xu and Yin (1990). They analyzed data derived from tree rings for the last 280 years and royal court chronologies for the last 760 years. Abrupt changes have been found for the end of the 13th century AD, the middle and end of 15th century AD, the middle of the 17th century AD, the early 19th century AD, and early 20th century AD.

The climatic characteristics of the Little Ice Age in China has been investigated by Wang (1990). This was based on historical documents from which information on early first frost (and later last frost), summer snowing and bitter cold in summer were calibrated with temperature observations. Decade mean and seasonal temperature anomalies were estimated from the frequency of the events. Finally, a series of decade mean annual temperature anomalies was reconstructed. For the last 600 years three cold periods were identified, each of them consisting of two colder phases:

Phase 1: From 1,450s to 1,470s A. mean annual temperature -0.30°C (lower than the average of 1,880-1,979 AD)
Phase 2.: From 1,490s to 1,510s AD mean annual temperature -0.34°C (ditto)
Phase 3.: From 1,560s to 1,600s AD mean annual temperature -0.47°C (ditto)
Phase 4.: From 1,620s to 1,690s AD mean annual temperature -0.60°C (ditto)
Phase 5.: From 1,790s to 1,810s AD mean annual temperature -0.43°C (ditto)
Phase 6.: From 1,830s to 1,890s AD mean annual temperature -0.45°C (ditto)

The annual temperature average for the coldest phase (1,620s to 1,690s AD) was 1°C or more colder than the 20th Century average. Warming occurred from the 1,920s to the 1,940s. Although three cold periods were also identified for Europe and North America, great discrepancies in timing of the cold periods between China and these countries were observed. Those in China seem to occur earlier.

Sultan and Gaofa (1990) investigated the influence of climate change on peasant rebellions in 17th century AD, at the end of the Ming Dynasty. The years 1,627-1,643 AD were characterized by droughts of unprecedented durations over most of northern China. “A year by year comparison of the geographical distribution of rebel activity with the moisture conditions shows that 75% of the rebellions occurred in drought areas, 15% in areas affected by flood and only 10% in areas with normal weather. ”

Zhong-wei et al. (1990) investigated the Northern Hemispheric jump in the 1960’s AD. Their conclusions with regard to precipitation and temperature, in addition to that of atmospheric pressure, are that rainfall in the zone stretching from North Africa to China-Japan decreased abruptly, while that in the zones both south and north increased. They also concluded that temperatures decreased abruptly in most of the Northern Hemisphere and increased in some low latitude regions.

A preliminary study on the 7,500-5,000 BP climatic sequence in the Middle and Lower reaches of the Yangtze River by Tang et al. (1990), was based on calculating the temperatures according to palynological data from 50 sites. The calculations were based on the weighted distribution statistical formula. Their results imply that the annual average temperature from 7,500 to 5,000 BP was 0.1. 2.71°C higher than the present; the hottest period was between 6,500-6,000 BP, when the mean value was 1.7°C higher than the present; and the coldest period was 5,500-5,000 BP when the mean temperature was 0.4°C higher than the present.

Shi (1991) reports on the natural environment of the Neolithic Age in China. He divides the cultures of this age in China into three main cultures:

1. Paddy growing in east China and south of the Yangtze River. From 7,000 to 6,000 BP (Hemudu Culture), the deposit which bears evidence of an agricultural tribe, and contains an assemblage of pollen suggesting a warm and moist climate. Temperatures were higher than the present. From 6,000 to 5,000 BP (Qingdun and Songze Culture), the climate was an intermediate one between the subtropic and temperate zones. It was warm and moist, lakes and swamps were widely distributed. People were farming on terraces, hunting and fishing. From 5,000 to 4,000 BP the climate became dry and cool, the sea level went down, and man subsisted mainly on agriculture. In the Yangtze delta area the change of climate and sea level controlled the development. Around 8,50() to 7,500 BP the climate became cold, the sea level was 10 m lower than the present, and there was no human activity. From 6,500 to 5,000 BP the delta was formed and man migrated to it. From 5,000 to 4,000 BP the climate became dry and cool, the surface area of water reduced.

2. The millet type agriculture on the Loess plateau. From 8,000 to 7,000 BP (the pre-Yangshao culture). The climate was warmer (2-3°C higher than the present.) From 7,000 to 4,000 BP (the Yangshao culture), the climate was cooler and drier in the earliest part, then warm and humid in the middle period and dry and cold at the end.

3. The microlithic culture of north-east China, Inner Mongolia and Qinghai-Xizang Plateau. From 10,000 to 8,000 BP an arid forest prairie dominated. From 6,000 to 4,000 BP was the warmest period, the temperatures being 2-4°C higher than the present. From 4,000 to 3,000 BP the climate was arid, lakes shrank the prairie widespread and salinization intensified.

In contrast to the forecast of Li (1990) of a positive influence on the precipitation in China, Fu et al. (1990) forecast a dry trend in China due to global warming. They base this forecast on the trend they identified during the past 100 years. They "...applied a linear fitting to observed records of drought and wet indices, summer monsoon rainfall, length of plum rains (Meiyu) and the discharge of Yangtze river since 1887. Super imposed on this long term trend, there is an oscillation with a period of about 36 years, (the so called "Bruckner period") which is supposed to be related to solar activities and air-sea interaction. The most interesting feature is that the climate entered into the dry regime abruptly at about 1920's, which is detected by Mann Kendall rank statistic test.".

According to their analysis of the meridional profile of the global zonal mean precipitation anomalies during the peak period of global warming (1,930-1,940 AD), the changes of mean meridional circulation can be deduced to illustrate the above relationship: "There would be an enhanced Hadley circulation in the Northern Hemisphere during the warming episode, which has an enhanced descending branch over the subtropics and extratropics where most part of China is located. "

The conclusion of Fu et al. (1990) is that if the global warming continues, the dry trend in China would continue and be strengthened further due to the enhanced descending of the Hadley branch in the future.

A similar trend of the climate becoming warmer and drier in the Jilin Province of China is reported by Pan (1990). This is based on the analysis of climatic data of Changchun since 1,909 AD. The records show that the temperatures in Jilin has gradually risen since 1,909 AD. Pan (1990) also forecasts that in the middle of the next century the average atmospheric temperature of the year will be higher by 0.6°C than it is today. As the climate of the Jilin province is cold, then if the temperature turns warmer then the seeding time would be earlier, which may benefit the agriculture of the province. Yet, on the other hand, records show that the annual average rainfall has been reducing since 1,909 AD at the rate of 1.287 mm per year; that from May to September by the rate of 0.187 mm per year; and that of December to February by 0.155 mm per year. As a result, the forecast is that the semi-humid region in the middle part of the Jilin Province will gradually change to semiarid.

Liu and Wang (1990) used annual runoff data for the years 1,700-1,988 AD of the Shanmenxia tributary of the Yellow River and found a significant negative correlation between the runoff and the Antarctic ice cap, when the runoff is lagged one year behind the ice amount. They also found alternations between periods of floods and droughts. Although there was a period (1,922-1,932 AD) with severe drought, a trend towards a long term drought can not be seen. Accordingly it can be forecast that the valley is likely to be drier as the ice amount increases significantly, and vice versa. The driest period (1,922-1,932 AD) of the valley coincided with the peak period of the ice amount. The connection is due to the influence of the ice cover on the storm trajectories in the antarctic during winter. "When the ice cover is heavy, the cyclonic activities at high latitudes decrease and move equate-wards. Following it, the sea-surface temperature starts to fall in the eastern tropical Pacific Ocean, the Hadley Circulation weakens and the Walker Circulation intensifies, which cause the change of the high." (Liu and Wang, 1990).

Zhang (1991) reports also on climate changes in the last 1000 years in China on the basis of his interpretation of historical records. His conclusion is that the period of the annual mean temperature during the 12th, 13th, 17th and 19th centuries AD were the coldest for the last 5,000 years. In 1,111 AD, for example, the 2,250 km² surface area of Thaihu Lake was completely frozen, the ice so solid as to be able to bear traffic, In 1,110 and 1,178 AD the winter's coldness killed all the litchi fruit trees. From 1,131 to 1,260 AD, the 10 years' average date of the latest snowfall day in Hangzhou was April 9th. The temperature in April in Hangzhou was l-2°C lower than the present. The snowlines in the Tienshan mountains in the 12th and 13th centuries AD were possibly 200-300 meters lower than today. The 14th century AD was also colder than the present one. In the winters of 1,329 and 1,353 AD the Taihu Lake was frozen sufficiently to enable walking on it, and all tangerine trees were destroyed. These phenomena have not been seen in the present century.

The temperature variations in the recent 500 years display three dominant cold warm cycles with a periodicity of about 180 years. The three cold intervals are from 1,470 to 1,520 AD, from 1,620 to 1,720 AD and from 1,820 to 1,890 AD. The warm ones intervals are from 1,530 to 1,620 AD, from 1,730 to 1,810 AD and from 1,900 AD to the present. Not mentioned is the cold period of the 1920's AD. Investigating the mean annual temperatures along the Yangtze river, the author of this document has come to the conclusion that there has been no trend of warming during the last century compared to the last 500 years (variation in ranges between winter maximum and winter minimum of 1.2-1.8°C, while during this century it has been 0.6-0.7°C), and there has been no general warming in the last 20 years compared to the whole century (the highest winter temperature for Shanghai and Hankou for the 1,940s AD is 5.0 and 5.1°C, whereas during the 1,970s and 1,980s AD the value was 4.9 and 4.7°C respectively).

Zhang (1991) also suggests the use of historical records of spring and autumn frosts, that although not being able to infer the spring and autumn temperatures, could be used to mark the character of activities of cold air in spring and autumn. He notes "...that many annual first frost dates recorded in historical literature of the last 600 years are earlier than, and last ones later than the present". His statistical analysis shows that the killing frost years, come in clusters (for example in the Yangtze River Basin, they were the following: 1,420-1,460, 1,500-1,540, 1,620-1,690, 1,720-1,780 and 1,830-1,910 AD. By incorporating data reconstructed from the Qingyulu (a daily weather record of Qing imperial archives) and the data of modem observations, a July mean temperature sequence of Beijing was obtained for the last 260 years. The power spectrum analysis of this sequence reveals periodicities of 2-3, 7-8,21 and 80 years, hot summers being alternated with cool ones. The hot summer intervals are 1,730-1,780, 1,820-1,870, and 1,920-1,950 AD, whereas the cool ones were before 1,730, 1,780-1,820, and 1,870-1,990 AD.

According to the same author, " ...an analysis of the historical literature of the last 2,000 years shows no linear trend, to be neither humid nor dry, in this period as for moisture variation, but rather repeated alternation of dry with humid or rainless with rainy intervals". He uses the frequency of large scale storms, plus flood records of the Yellow River, to determine the features of annual moisture variations for the last 1,000 years. This is compared also with the dust-fall frequencies, and with historical data from all over eastern China, including the Qingyulu (the daily weather record of Qing imperial archives) of Nanjing, Suzhou and Hangzhou (1,720-1,800 AD) which includes report on the Meiyu (plum rain).

Sun & Chen (1991) summarize the palynological research carried out during the last 30 years in China. Their main conclusions are:

1. The vegetation changes to more warm type around 10,000 BP are sudden and remarkable. This rise took only a few hundred years;

2. There was a Holocene warm period between 10,000 and 4,000 BP. This period can be divided into sub-periods 10,000 to 8,000 BP, the temperature rising period, and 8,000-4,000, the period of high temperatures;

3. The temperature decline in the Late Holocene, considered to be at 2,500 BP, may have started earlier, in most cases at about 4,000 BP. In contrast to the change at the beginning of the Holocene, this change was much slower and more gradual, taking some 1,000 to 2,000 years to reach a stable period.

Wu & Zhan (1991) extracted proxy data of climatic change from tree-ring widths in China. Their conclusions relate to three main regions:

1. The Xizhang Plateau of Tibet. This region had distinct fluctuations of temperature during the last 2,000 years. The climate was cold early in the 1st century AD. Then it became warmer, and was 1°C warmer during the 2nd and 3rd centuries AD. It maintained a longer cold period from the late 3rd century to the 5th centuries AD. Generally, it was about 1°C lower than that of the present. From the 6th century to the late 12th century AD, the temperature increase was more remarkable in the plateau. The 13th century AD was a distinct cold period. Around the 17th century AD the climate was the coldest for the preceding 1,000 years. It has been getting warmer since the mid 19th century AD. The variations of droughts and floods in the Tibet plateau, for the last 340 years derived from tree-ring series, for example, show that the major wet period appeared from the mid 17th century through the early 18th century AD, with another from the end of the end of the 19th century to the 1,930s AD. The authors also remark that according to the features of climate change in Tibet during historical times , it can be deduced that the plateau is an area more sensitive to climate change than other areas on the same latitude.

2. The Hengduan Mountains, in which a dry cold pattern was prevalent during the first half of the 17th century AD. In the second half of the same century the climate was warm and wet. In the 19th century AD both cold and wet years increased. The dry-warm pattern has been increasing since the beginning of the present century. on the whole, the variations in air temperature in this region are smaller than in Tibet and East China.

3. Northwest China. The general trend of temperature fluctuations is similar to that of the east part of China during the last several hundred years.

In general it can be said that the fluctuations in the Tibet area correspond very well with those of the Levant, and indeed the Tibet Plateau can be regarded as an important site to monitor global climate change.

Wang et al. (1987) investigated the drought/flood variations of the last 2,000 years in China in comparison with the global climatic change. Based on historical documentation they have extended the drought/flood series back to 951 BC, especially for the Changjiang River and Huanghe River. Generally they found four dry periods in the Huanghe River (4-6th, 11-13th, 15th, and 19-20th centuries AD) and four dry periods in the Changjiang River (4th, 9th, 1 l-13th, and 15-17th centuries AD). Analyzing the special situations of the Little Optimum (1l-12th century AD) and that of the Little Ice Age (] ,550-1,750 AD) they found that the first period was dry while the second was wet. This was correlated with the information from Japan (Yoshino, 1978) that the Little Ice Age was cold wet, which is explained by the fact that during the Little Ice Age the trough in east Asia may have been placed near Lake Baikal and the blocking Okhotsk High had occupied the east coast of Asia.

Xu & Yao (1990), from the Institute of Desert Research at Lanzhou in the north-western arid part of china, suggest three modes of division of the Holocene according to different time scales. They base their division on data from glaciers, geology, lakes, tree rings, ice cores and historical data:

1. On the general time scale they suggest that the Holocene be divided into three parts: Early Holocene, 10,000-7,500 BP, cold and dry; Middle Holocene 7,500-3,500 BP, warm and wet; Late Holocene, 3500 BP to present, cold and dry.

2. On the thousand year scale they suggest the following: 8,700-7,500 BP, 1st cold stage; 7,500-5,900 BP, 1st warm stage; 5,900-5,400 BP, 2nd cold stage; 5,400-4,200 BP, 2nd warm stage; 4,200-3,800 BP, 3rd cold stage: 3,800-3,000 BP, 3rd warm stage; 3,000-2,700 BP, 4th cold stage; 2,700-2,000 BP, 4th warm stage; 2,000-1,400 BP, 5th cold stage; 1,400-1,000 BP, 5th warm stage; 1,000 BP to present, 6th cold stage. During most of the stages the cold corresponded with dry, and the warm with wet.

3. On the hundred year scale for the last thousand years they found that the main cold spells appeared in the 1,100S, 1,310s, 1,470S, 1,680s, and 1,820S AD. The main warm spells were in the 1,200s, 1,390s, 1,570s, 1,780s, and 1,880s AD. The changes in precipitation in recent times, mostly in the arid and semi arid part of China, have been mainly a rainy and cold stage around 1,800 AD, and a warm and dry stage, with drought frequency increasing in the 20th century AD.

The present author suggests the consideration of the possibility that this area is a transition zone, influenced differently in different periods, either by the extreme north-western westerlies belt, or sometimes by the eastern rnonsoonal regime. Thus, correlation with other regions will be difficult without knowing from which part of the province the data had been secured.

Wan et al. (1991) report on natural hazards in the Gansu Province in China. Most of this province lies in the semi-arid and sub-humid zone, within the western margins of the monsoonal region of China. The summer wind pattern has a very strong influence on the climate of southern Gansu, the variation in precipitation being very large from year to year. As a result, droughts are frequent and very severe. Climatic records from the Han Dynasty, about 2,000 years ago, to the present show that some 315 droughts have occurred, of which 43 were very severe. A cyclic periodicity of 121, 64, 32, 15, 8-9, and 2-3 years is claimed. The cycles have become shorter and have smaller amplitude since about 200 years ago, and it seems clear that the climate of eastern Gansu has become progressively drier.

Fan & Shi (1990) investigation the impact of climate change on the hydrological regime of Qinghai Lake, which is a large inland lake, and may reflect well the impact of the climate change in northwest China and Central Asia. For the past 30 years the climate in the lake basin has had a tendency to warm, causing upland drying. Sixty percent of the years during this period were dry. From the early 1960s to the early 1980s AD the annual temperature increased by 0.3°C. Due to the warming and drying the lake level descended continuously. The historical climate of the lake basin is reflected in the lake sediments. On a thousand year time scale the climatic feature is warm-humid and cold-dry. On the hundred or decade scale it is warm-dry and cold-humid.

Bradley et al. (1987) studied the secular fluctuations of temperature over Northern Hemisphere land areas and mainland China since the mid-19th century AD. Records for the two regions show similar trends, namely temperatures increased from the late 19th century AD to a maximum in the 1,940s AD, followed by a cooling trend which reversed over the last 10-15 years. Extremely sharp drops in temperature, particularly in fall months, occurred after several major volcanic eruptions. High temperatures are sometimes associated with major El Niño years. When the two occur more or less simultaneously their influences are minimized.

Lough et al. (1987) have investigated the relationship between the Climates of China and North America over the past four centuries by comparing proxy data. They found that drought in China appears to be directly and significantly correlated with a weakening' of the North Pacific subtropical anticyclone. This is associated with above-average precipitation in the central and southern U.S.A. Floods in China are associated with the intensification and expansion of this anticyclone and the decrease of precipitation over central and southern U.S.A. A lag of one year between the two regions is obvious.

4.4.2 Japan and Korea

Takahashi (1987), through an analysis of historical and other relevant data from 1,650 to 1,983 AD of long term variation of storm damage in Japan, found a 70 years cycle, which tends to increase when the climate is warm. His analysis indicates that the influence of volcanic activity on climatic change (hence on storm damage in Japan) is considerable on a time scale of several decades.

Mikami (1987) reconstructed the climate of Japan during 1,781-1,790 AD, comparing it with that of China in order to derive data from Chinese processed historical documents. The climate of Japan in the late 18th century AD is estimated to have been very cool and wet, as severe famines due to crop failure occurred frequently. From 1,782 to 1,787 AD Japan experienced the most severe famine in its history, which reduced its population by approximately one million. The statistical examination of the relationships between Japan and China show simultaneity of rainfall in Japan and Central China, with opposite indications in north and south China, These are considered to be linked with global circulation patterns. Thus in 1,978 AD when it was extremely dry and hot in Japan and middle China, the axis of the subtropical high (North Pacific High) was located northward. Such a situation most probably occurred in 1,785 AD. In 1,980 AD when it was exceedingly wet and cool in Japan and middle China the axis of this high was shifted to the south and the polar frontal zone extended from middle China to Japan. Such a situation most probably existed also in 1,783 AD.

Kim & Choi (1987) report on a preliminary study of long term variations of unusual climate phenomena during the past 1,000 years in Korea. They found the following classification:

1,381-1,420 AD, cold period (winter warm, summer cold).
1,421-1,520 AD, warm period (winter and summer warm).
1,521-1,630 AD, first period of Little Ice Age, (winter and summer cold).
1,631-1,740 AD, second period of Little Ice Age (winter cold, summer very cold).
1,741-1,840 AD, third period of Little Ice Age. (winter and summer cold).
1,841-1,930 AD, warm period (winter and summer warm).
1,881-1,930 AD, cold period (winter and summer cold).
1,931-to the present, warm period (winter and summer warm).

They also observed that cold winters in the early 19th century AD corresponded to periods of weak sun-spot activity.

Cool weather phenomena in summer seasons in the Korean East Sea region occur owing to the intensifying of the Okhotsk High toward southern or south-westerly directions in summer seasons.

4.5 North At-mica

4.5.1 Western U.S.A

Pollen analyses by Davies et al, (1977) of Wildcat Lake, Whitman County, Washington State indicated: the Last 1,000 years have shown that although terrestrial pollen percentages are relatively stable prior to the introduction of horses, sheep and cattle, changes occurred in the Wildcat Lake Aquatic environment following deposition of Mt. St. Helens volcanic ash (400 - 500 BP). Two periods of intense erosion followed the introduction of grazing and subsequent range deterioration. It is obvious from the studies at Wildcat Lake that no climatic event of the past 1,000 years has resulted in vegetation changes as great as those brought about by European agricultural and grazing practices.

4.5.2 California

The Californian depression, situated between the Coastal Range and the Sierra Nevada Mountains with a length of 580 km inside American territorial boundaries, is a product of the sinking of a large closed basin onto which sediments from the surrounding mountains were deposited. Later, a lake developed in this area. The present Lake Tulare is a remnant of the old lake. The zone has a moderate climate with an average annual precipitation in excess of 1000 mm. During the summer, temperatures vary between 27° and 28°C; in winter, between 7° and 8°C.

On the Pacific coastal plain (in Mexico), the climate is dry and very hot. Here, cold winds from the northwest blow for eight months of the year. During the summer, the southwesterly winds bring torrential rains.

The most detailed record for the climate changes which took place in this region during the Holocene was derived from the study of the tree rings of bristolecone pine growing near the upper tree line on the white mountains in California (Lamb, 1982). The variations in the width of the rings is suggested by Professor V.C. La Marche (from the Laboratory of Tree Rings Research, University of Arizona in Tucson) to indicate variations of summer warmth and/or its seasonal duration. This means that wider rings indicate warmer periods, If this is correct, then there is a negative correlation between this curve and the base section (Fig. 18). On the other hand, when the width of the tree rings is interpreted as an indication of precipitation, i.e. wider rings indicating more precipitation and vice versa, and if more precipitation means, as in the Mediterranean, periods of colder climate, then the correlation with the base curve of the Levant is rather good.

4.6 Southern Hemisphere

4.6.1 South America

The Late Quaternary climate of the western Amazon basin was studied by Colinvaux and Liu (1987) and shows that from 33,000 to 26,000 BP it was 4.5°C cooler in the Amazon lowlands, and from 1,300 to 800 BP it was a wet period. There is clear evidence from lake sediment records of greatly increased precipitation for the western Amazon basin in the period 1,300-800 BP, dated by radiocarbon (Colinvaux and Liu, 1985). The above period corresponds to about 350-1,150 AD in historical years and nicely fits the two warm periods as designated above (300-400 AD and 500-1150 AD) observed in the Levant.

Ice cores from the tropical Quelccaya ice cap in Peru, situated at an elevation of 5,670 m, show that both the onset and termination of the Little Ice Age were abrupt in tropical South America and happened around 1,490 and 1,880 AD, respectively (Thompson and Mosley-Thompson, 1985. The implication is that for this part of the world, there was no warm period around 1,705-1,880 AD. The warm period 1,920-1,963 AD can be largely associated with a relatively wet climate in the Sahel. The only significant dry phase in this period occurred from 1,944 to 1,948 AD (according to Rognon, 1987) or 1,946-1,949 AD (according to Lamb, 1983).

The greatest drought recorded in southern Peru occurred from 1,933 to 1,945 AD. The level of Lake Titicaca, the highest (3,812 m) large lake in the world, dropped by almost 5 m This drought is very clearly reflected in the tropical Quelccaya ice cap in Peru. Ice layers for the above dry period show the $d^{18}O\%$ ratios to be clearly less negative (- 18 to - 14), while a substantial increase of dust can be observed in the ice (Thompson and Mosley-Thompson, 1987).

The beginning of a major drought in the Sahel in 1,968 AD has been associated with the global cooling trend which began in the middle 1,960'S AD, starting about 1,964 AD (Kukla et al., 1977; Newell and Hsiung, 1987). However, the drought also continued after 1,977 AD, when global temperatures began to show a warming trend. Zonal mean air temperatures for the 300-700 mb layer in the tropics became clearly higher after 1977 (Newell and Hsiung, 1987).

4.6.2 New Zealand

Climate data from stalagmites growing in a cave in New Zealand were obtained by $d^{18}O\%$ measurements on samples taken at regular intervals down the axis of the stalagmite and a time base was obtained by ^{14}C dating (Wilson et al., 1979). The resulting isotope-temperature curve has been compared with the curve given for central England by Lamb (1965). The isotope curve is scaled in terms of temperature by calibrating it with modern meteorological records from a site near the cave. The two curves appear to be broadly similar suggesting that such climate fluctuations as the Medieval Warm Period and the Little Ice Age are not just local European phenomena.

5. SUMMARY, CONCLUSIONS AND RECOMMENDATIONS

The rational Mind the work presented in this report was that the knowledge of the impact of climate changes during historical periods on the paleo-environments can be used as a key for estimating impacts of climate changes in the future. It may also serve well for the calibration of global circulation models (GCM), in view of the fact that these models are very complex, and up to date their simulations with climatic scenarios of the glacial and interglacial have been too extreme when expectations of a greenhouse effect are presented.

The investigation of these changes has involved a general survey of climate dependent records in the Mediterranean region and China during the historical period. This included isotopic data collected from lakes, seas, glaciers and caves, palynological and dendrochronological data, paleo-levels of the sea and lakes as well as historical and archaeological records on desert settlements.

The first part of the research concentrated on the climate variations during the Holocene in the Eastern Mediterranean region.

The climate of this region is affected during the summer by the subtropical high pressure belt, which results in hot and rainless weather conditions. During winter the region is dominated by the by mid-latitudinal depressions, connected with the westerlies regime. This brings cold weather and rainfall.

It was hypothesized, that in the past, during warm periods, summers were longer and winters were milder. This may have caused droughts all over the region and the movement of the desert belt northward. During colder periods the conditions would have been reversed.

In order to trace the climate changes during the Holocene, the conditions which prevailed during the Last Glacial and post glacial periods were investigated. It was found that during the glacial period the region was more humid due to the southern migration of the westerlies belt. This brought more rains which caused the level of the Dead Sea to rise. During this period the precipitation was isotonically lighter, as was found in the Paleo-water under the Sinai and Negev deserts. Also during this period loess layers were deposited, a function of the deposition of dust brought in by dust storms followed by rain storms. As the global climate became warmer the paleo Dead Sea receded, and sand dunes invaded the coastal plain. In the south they covered the loess.

In order to trace climate changes during the Holocene the changes in the isotopic composition of the precipitation was investigated. This was obtained from cores in lacustrine carbonates of the Sea of Galilee and cores in stalagmites in caves of the Galilee, northern Israel. This was accomplished by the development of a palynological time series from various lake deposits in the Levant. All these data were correlated with archaeological data which were derived from surveys in the Sinai and Negev Deserts and complemented by the results from studies on the paleo levels of the Mediterranean sea. The hydrological data were mainly obtained from the studies of the Dead Sea paleo levels.

The general and most important conclusions which can be derived from the interpretation of the above mentioned sources of data were the following:

- A cold and humid period existed during the Pre Pottery Neolithic A period (ea. 10,000 and 9,500 BP),
- A warm and humid period existed as a consequence of the monsoon influence during the Pre Pottery Neolithic B period (ea. 9,500-8,500 BP),
- A cold and humid period existed during the Pottery Neolithic A period (ea. 8,500 and 7,500 BP),
- A warm and dry period existed during the Pottery Neolithic B period (ea. 7,500 and 6,500 BP),
- A humid and most probably cold period existed during the Chalcolithic period (ea. 6,500-5,000 BP), except at the end of the period when the climate became warm and dry,
- A short warm period of the Upper Chalcolithic existed around 5,000 BP,
- A cold and humid period existed during most of the Early Bronze Period (ca 5,000 BP to 4,200 BP),
- A warm period started around 4,200 BP (Early Bronze VI) extending to around 3,400 BP,
- From around 3,400 BP to 2,600 BP (the Iron Age) there was another cold and humid period,

- From 2,600 to 2,300 BP (mainly during the Persian and Hellenistic empires) the climate was rather warm and dry,
- During the Roman period (ea. 2,300 BP to 1,400 BP, with a short interval lasting about 100 years around 1,700 BP) a colder and more humid period prevailed,
- A warm and dry period started ca. 1,400 BP (The Arab warm period) reached a peak of warmth and drought ca. 1,200 BP and abated ca. 1,000 BP,
- From ca. 1,000 BP to 800 BP the climate was rather cold and humid (The Crusaders cold period),
- At 800 BP a warm and dry period started which continued about 150 years (The Turkish warm period),
- From 650 BP to about 150 BP (the Industrial period) the climate was cooler and more humid, due to The Little Ice Age,
- From about 150 years ago the climate became more warm and dry.

As numerous observations showed that the regime of the Nile did not conform with that of the Levant, it was suggested that the unconformity might be an indication that the Nile (which originates in the monsoonal and tropical regions) reflects the influence of climate changes on the hydrology in these regions. In that case it is suggested that warm periods will generally equate with more humidity, while cold periods with more dryness.

In order to investigate the suggested hypothesis a detailled investigation of climate changes and their influence on the hydrology of China was carried out.

The climate of China is affected in winter by the polar continental air mass (PCAM) causing a northerly flow of cold dry air masses. In the summer the region is dominated by the tropical-subtropical oceanic air mass (TOAM) and a tropical continental air mass causing a southerly monsoon which draws oceanic warm and moist air from the low latitudes, causing rains.

The study of the Chinese records of climate changes during historical periods showed a good correlation with that of the Mediterranean region. Cold dry periods in China correspond with the cold wet periods in the Mediterranean region, while the warm wet periods of China corresponds with warm dry periods in the Mediterranean region. The 1,000 BC period is that of the Iron Age, followed by a period of warmth and dryness which had started to change in the Middle East by 300 BC to a colder and more humid period, but had no impact until 0 BC/AD. This continued during the entire Roman cold period. The Arab dry and warm period in the Levant was warm and humid in China. The Crusaders cold and wet period in the Levant was again a cold and dry period in China. The same happened again during the Little Ice Age.

Thus, although the observations described in the present document are but a small part of the ample data which exists from the historical period in the Mediterranean region in China and which Chinese scientists have accumulated and processed, the general, conclusion which may be drawn from this data is that indeed in the case of a global rise in temperature, the

Mediterranean region (which belongs to the westerlies climatic system) will become more dry and arid. On the other hand in China and probably all South East Asia a warm period will mean substantial rains, and most probably floods. This is in agreement with the results obtained by the simulation runs of the GCMs.

In the past the price that the people had to pay for good years in China was the loss of life and damage caused by floods. This might also be true for South-East Asia, e.g. Bangladesh, India, Burma, Thailand and Sri Lanka. And as it is in these countries that live the majority of the agricultural population of the world, the green house effect seems to also have its positive side. This would happen, however, only if precautions are taken in time to reduce the damages inflicted by the floods. The recommended strategy in this case, should be based on the “win or gain principle”, i.e. that flood control projects should be carried out. Thus, if the greenhouse effect takes place, these projects would mitigate its effects; while if it should not happen, then the projects would help to promote the economy of the region. In China the main precautions are the system of dams proposed to be built on its main rivers. There are many reasons suggested why these projects should be avoided. Most reflect a fear of the negative impacts on the environment. From a climate-change perspective, however, there are arguments for their construction.

In the Middle East the greenhouse effect would probably reduce the quantites of water available at present. As water in this region is already a commodity in shortage, such a forecast would demand that immediate steps be taken in order to avoid further severe shortages, and thus reduce the potential for conflicts and even wars between the countries of the region.

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Fig. 1- The base section of the middle and upper part of the Holocene

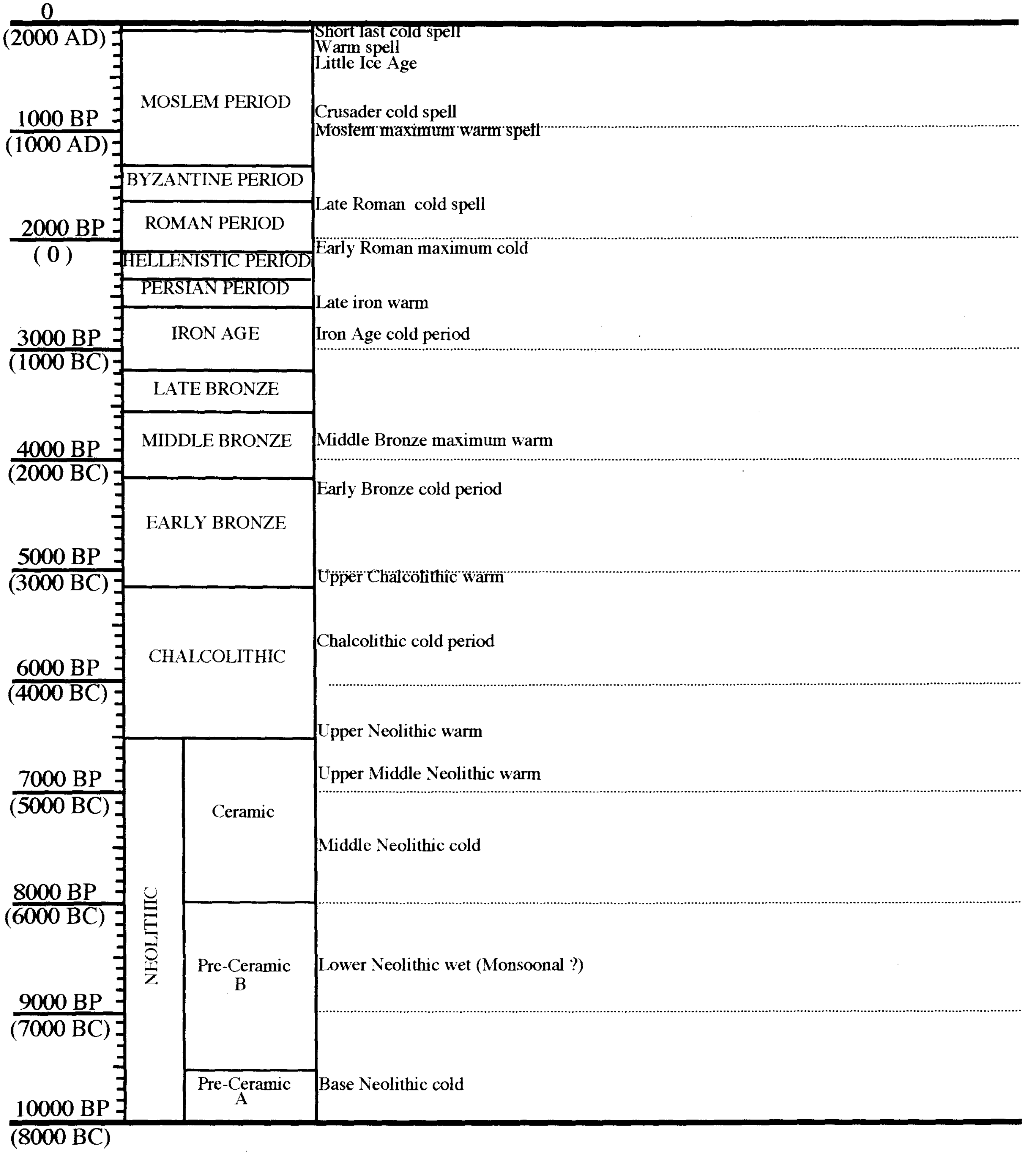


Fig.2. Correlation between curves of time - series in the Levant

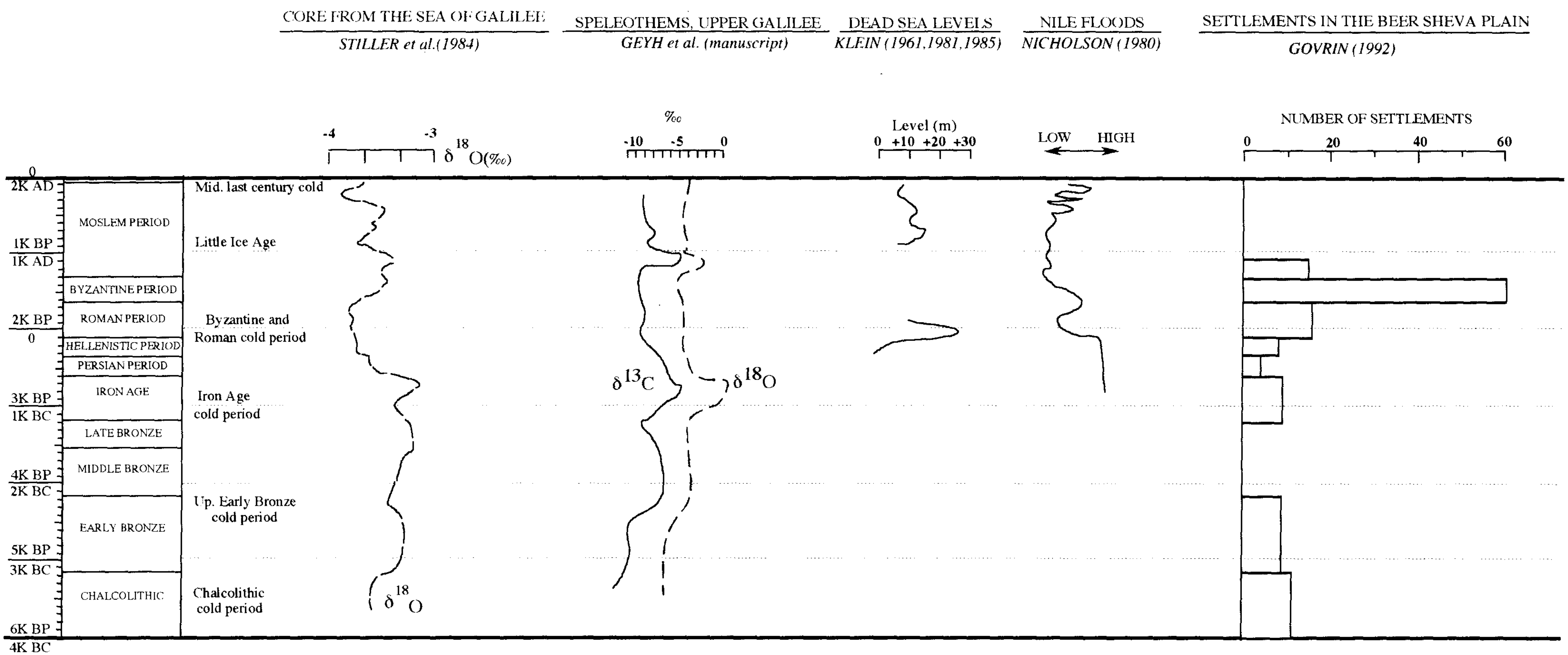


Fig.3-Dead Sea lake levels (Frumkin et al. 1991)

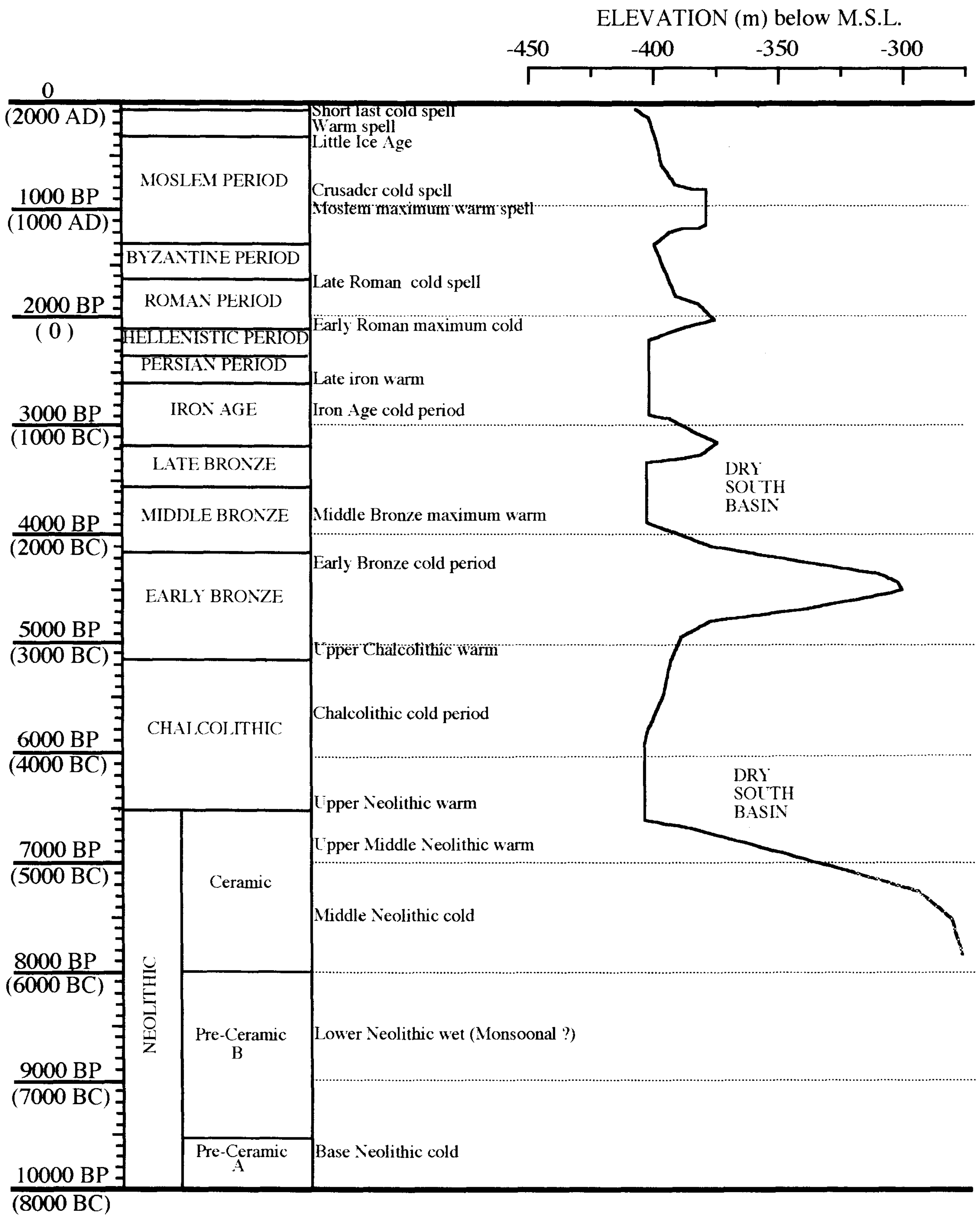


Fig. 4. Mediterranean Sea levels

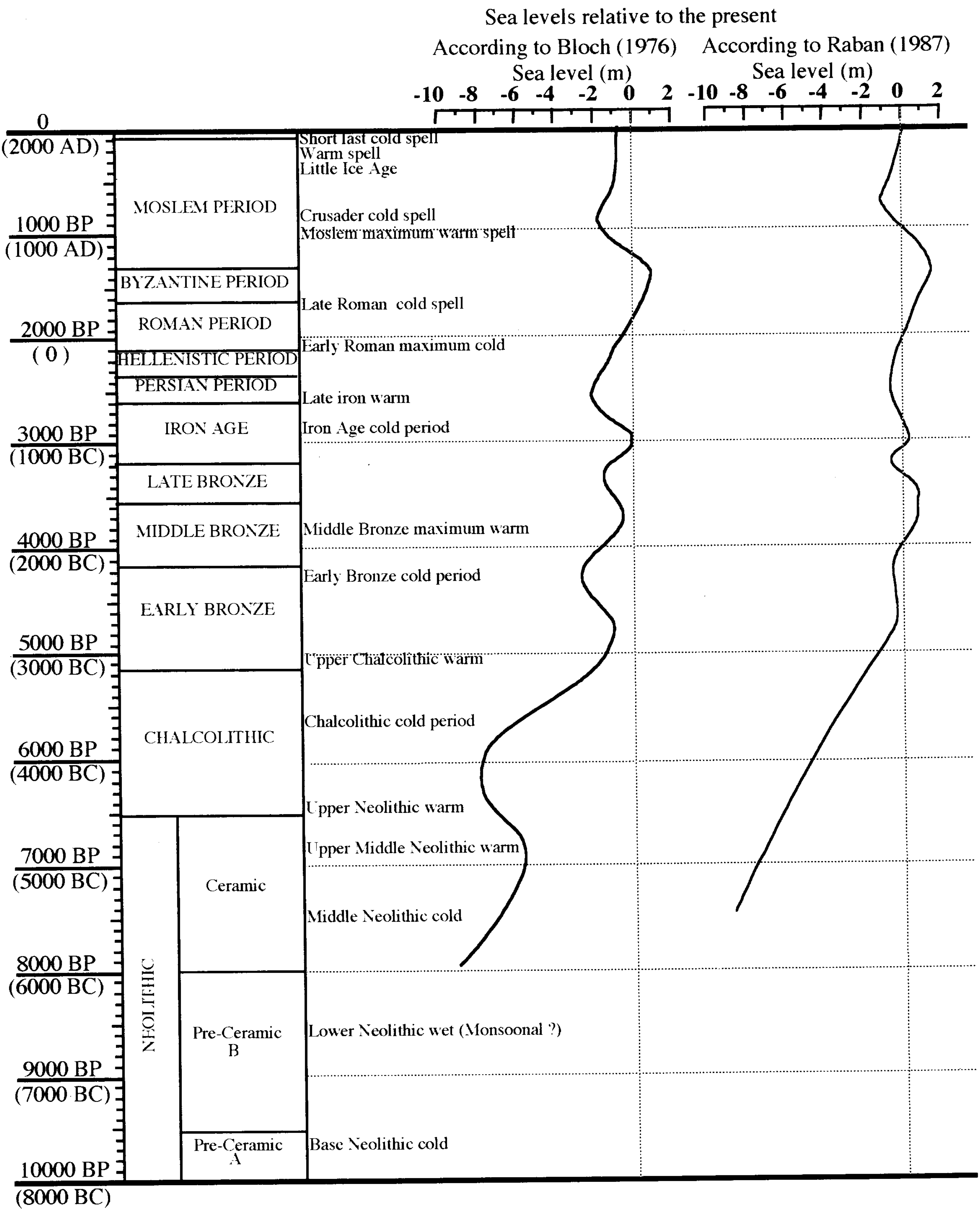


Fig.5- Palaeo levels of Lake Van and the base section (Degens et al. 1984)

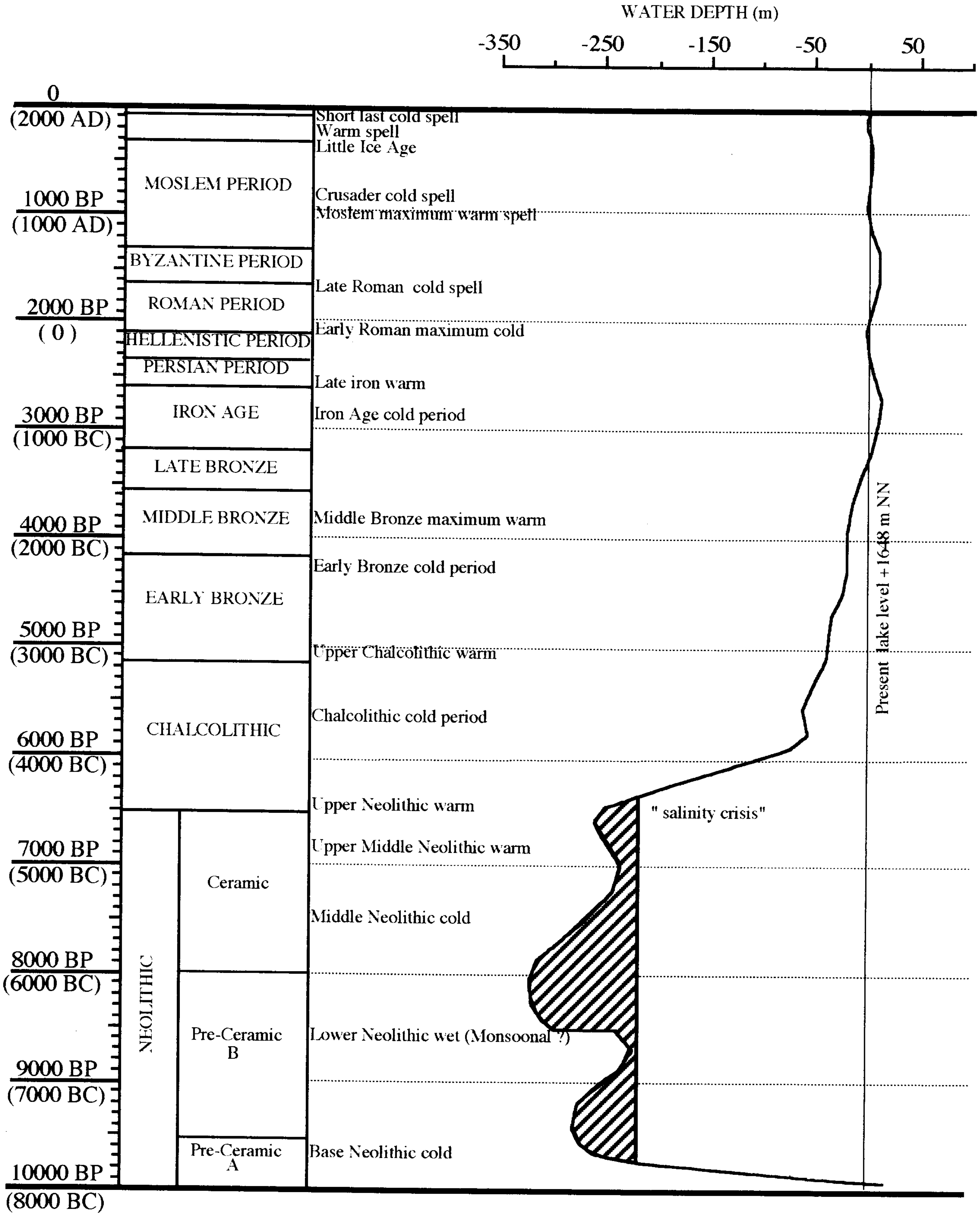


Fig.6. Correlation between sedimentation sections in Greece (Hatzious et al., 1984, cited in Papae 1984) and Levant curve.

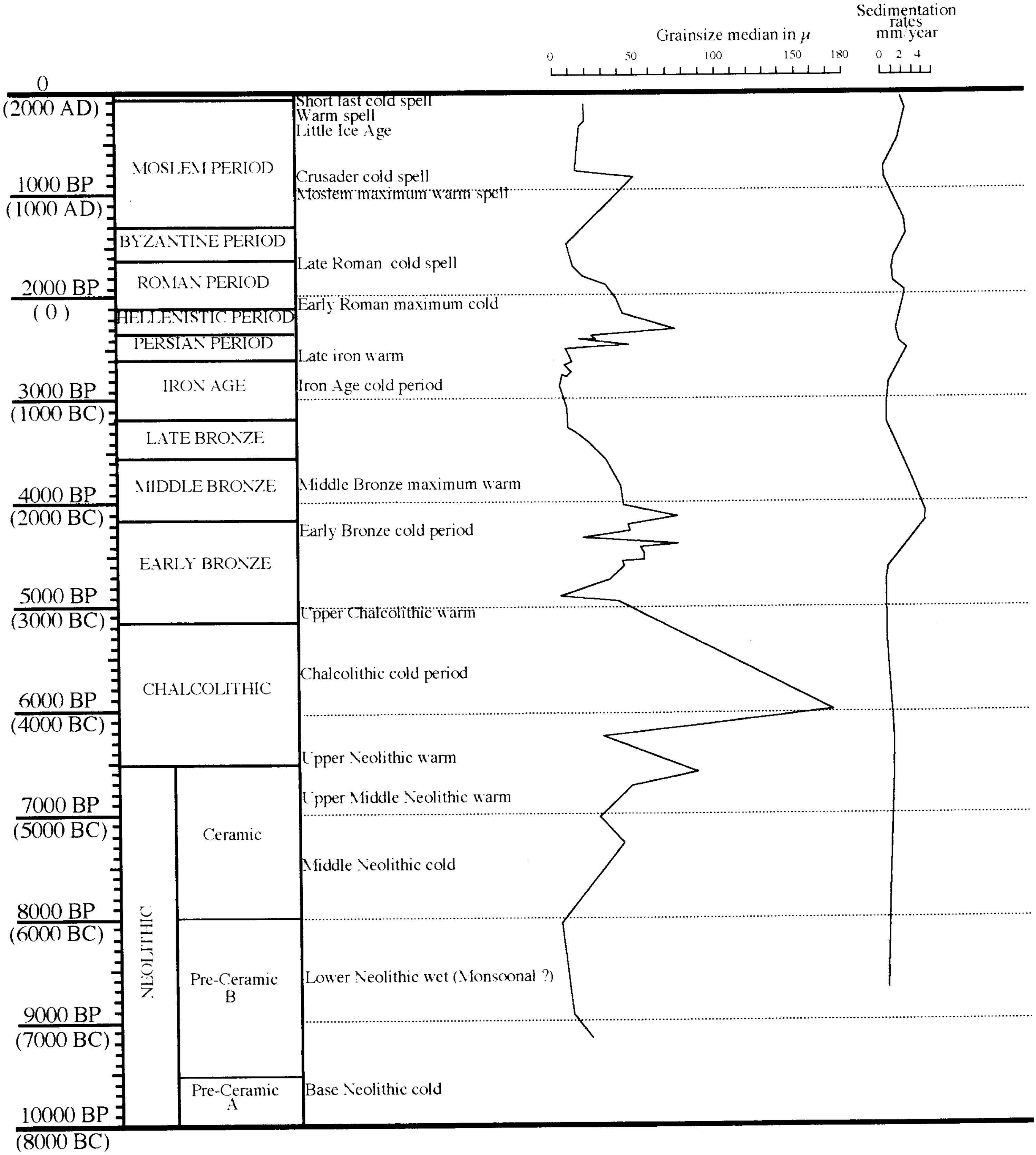


Fig 7 . Correlation between the Levant climate base section and the upper part of the curve of the levels of Lake Fucino-Italy (Giraudi, 1989)

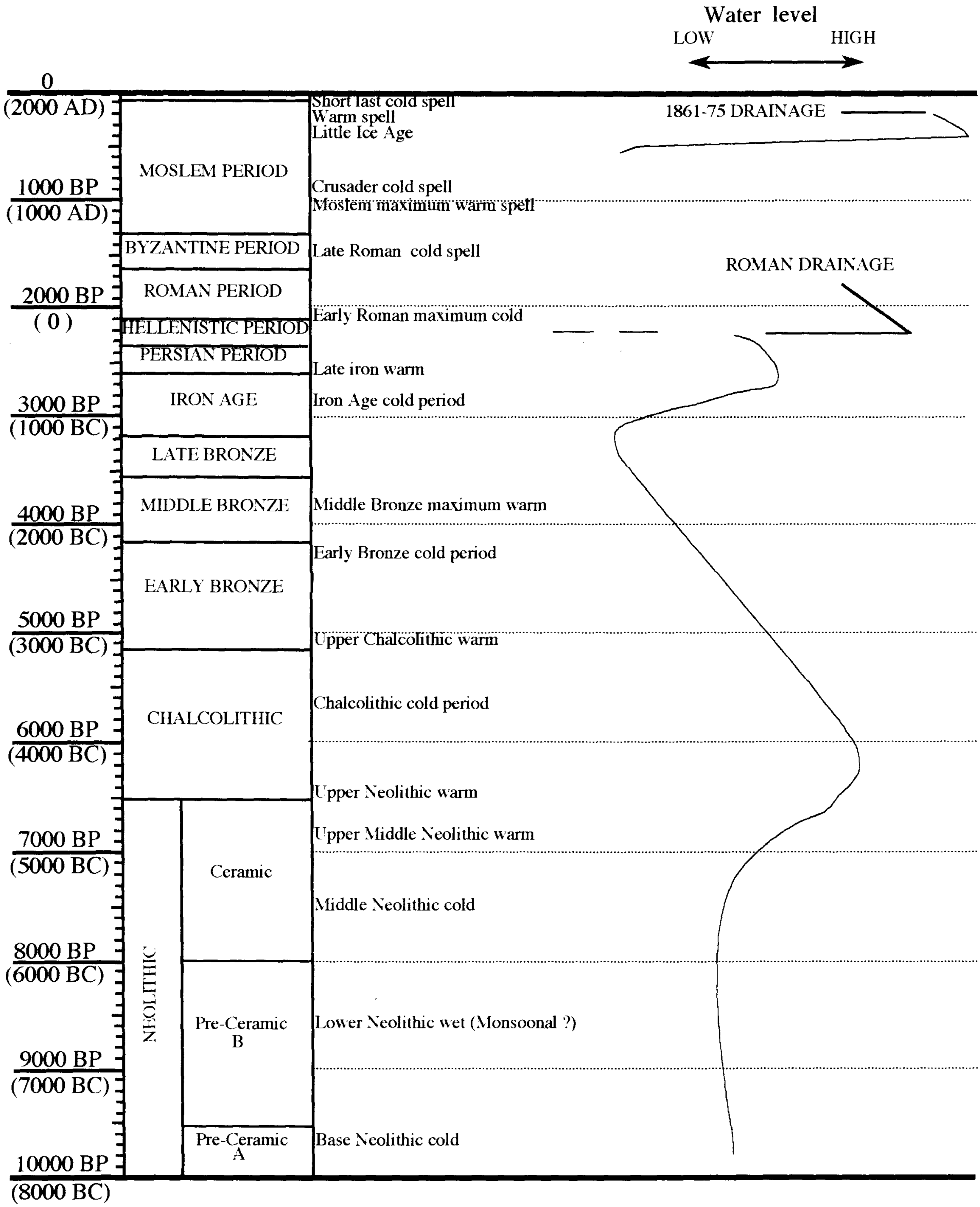


Fig. 8- Pollen diagram from Castillo de Caltarava (Garcia Anton M, et al., 1986)

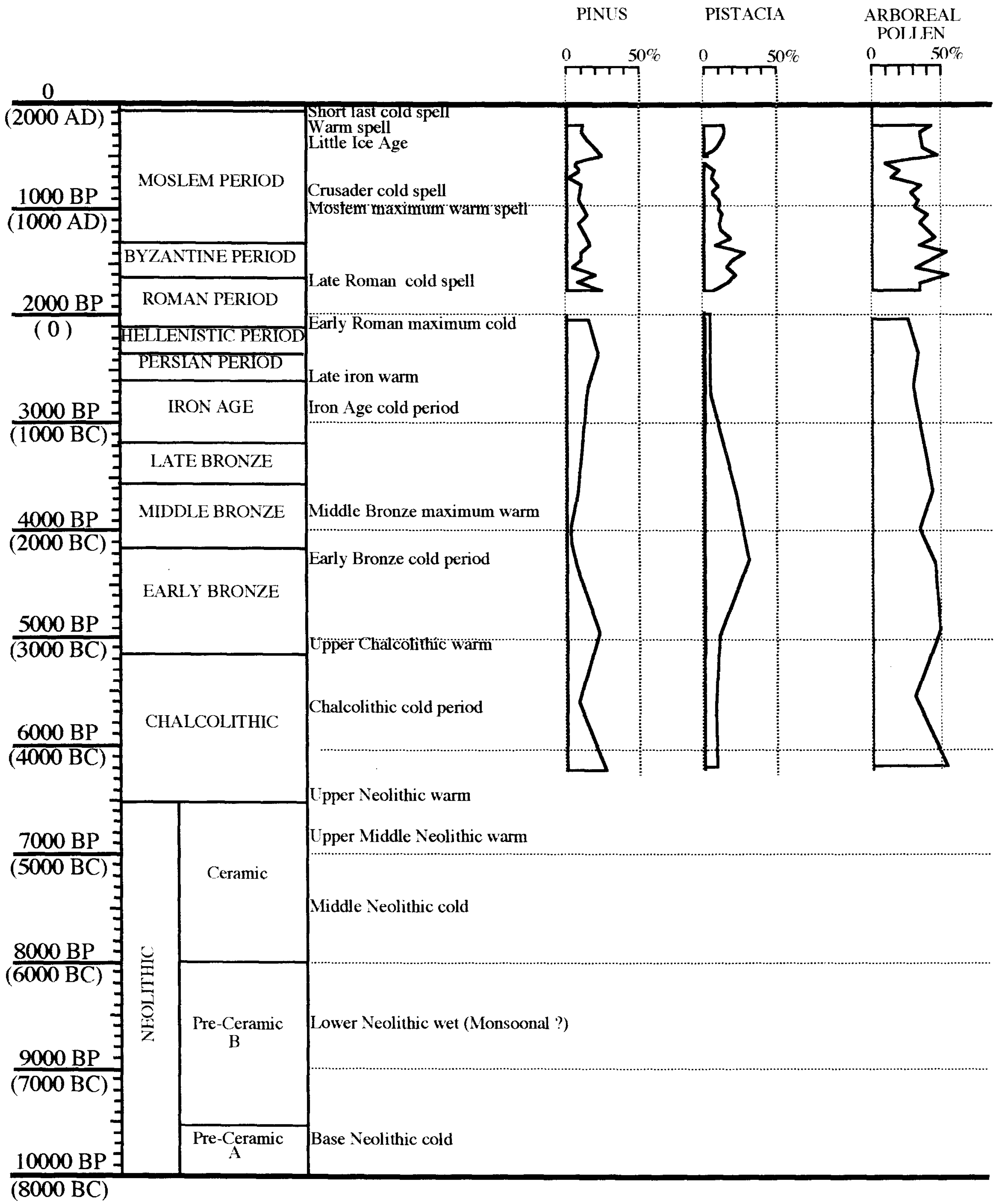
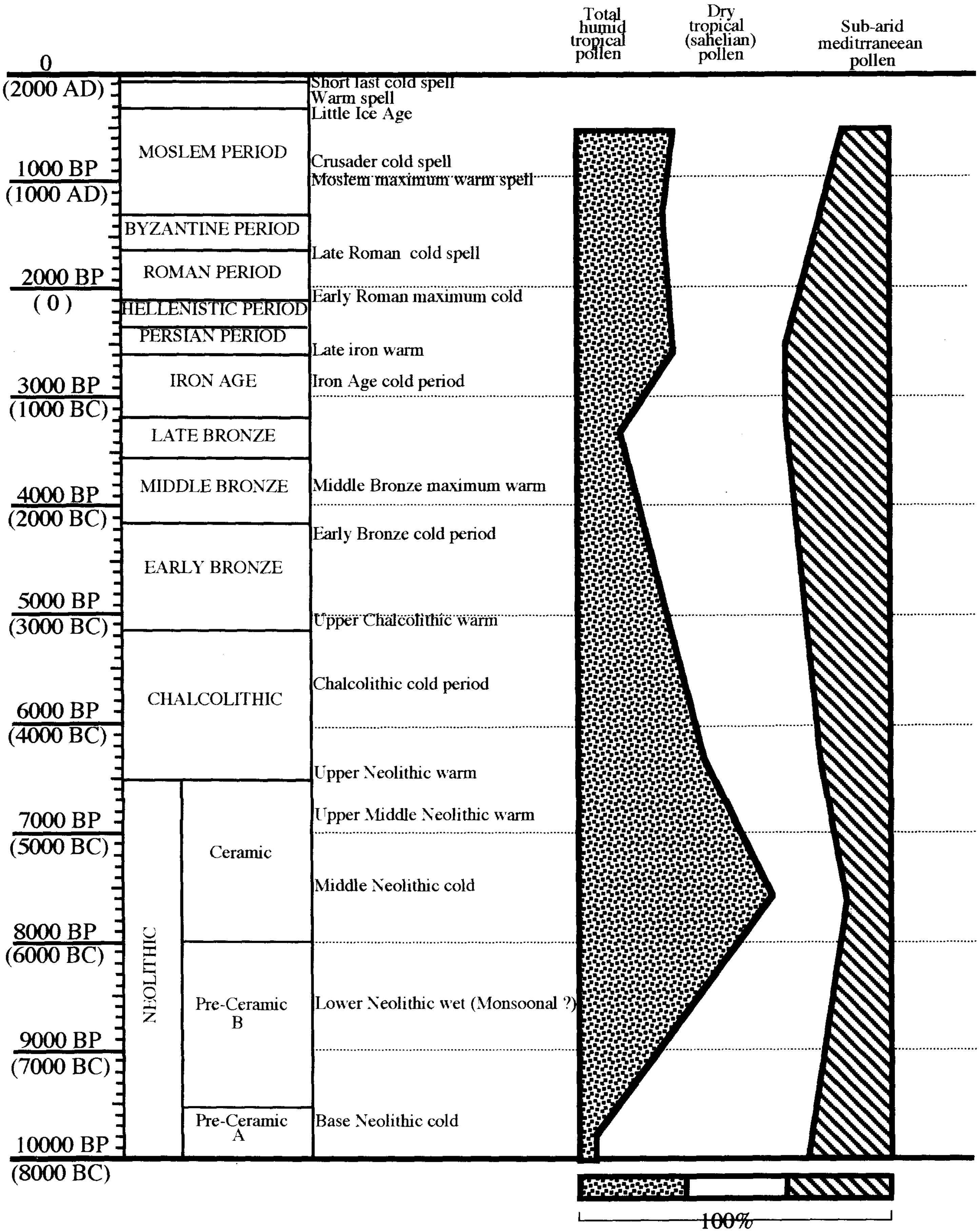


Fig.9. Correlation of pollen record from a core in the Arabian Sea with the Levant base curve (Van Campo et al., 1982)



TOTAL HUMID TROPICAL TAXA INCLUDE HUMID TROPICAL, TROPICAL MONTANE AND SUDANO-SAHELIAN TAXA

Fig.10- Correlation of Paleoclimates of Central Sahara (Maley, 1977,b) and the Levant base section

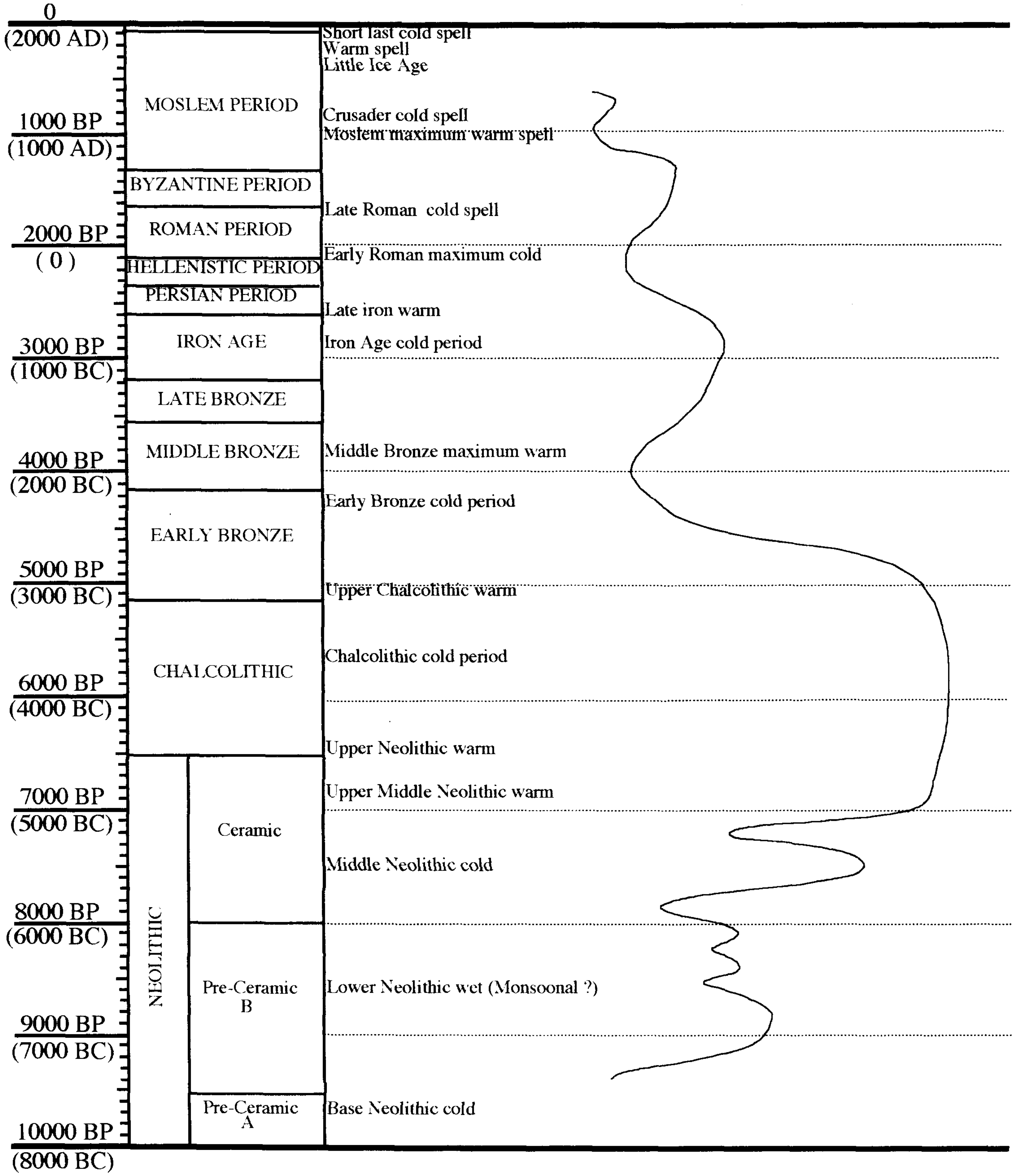


Fig. 11. Correlation of glaciers' advance and retreat in Scandinavia and the Levant base curve (Karlen, in Starkel et al., 1991)

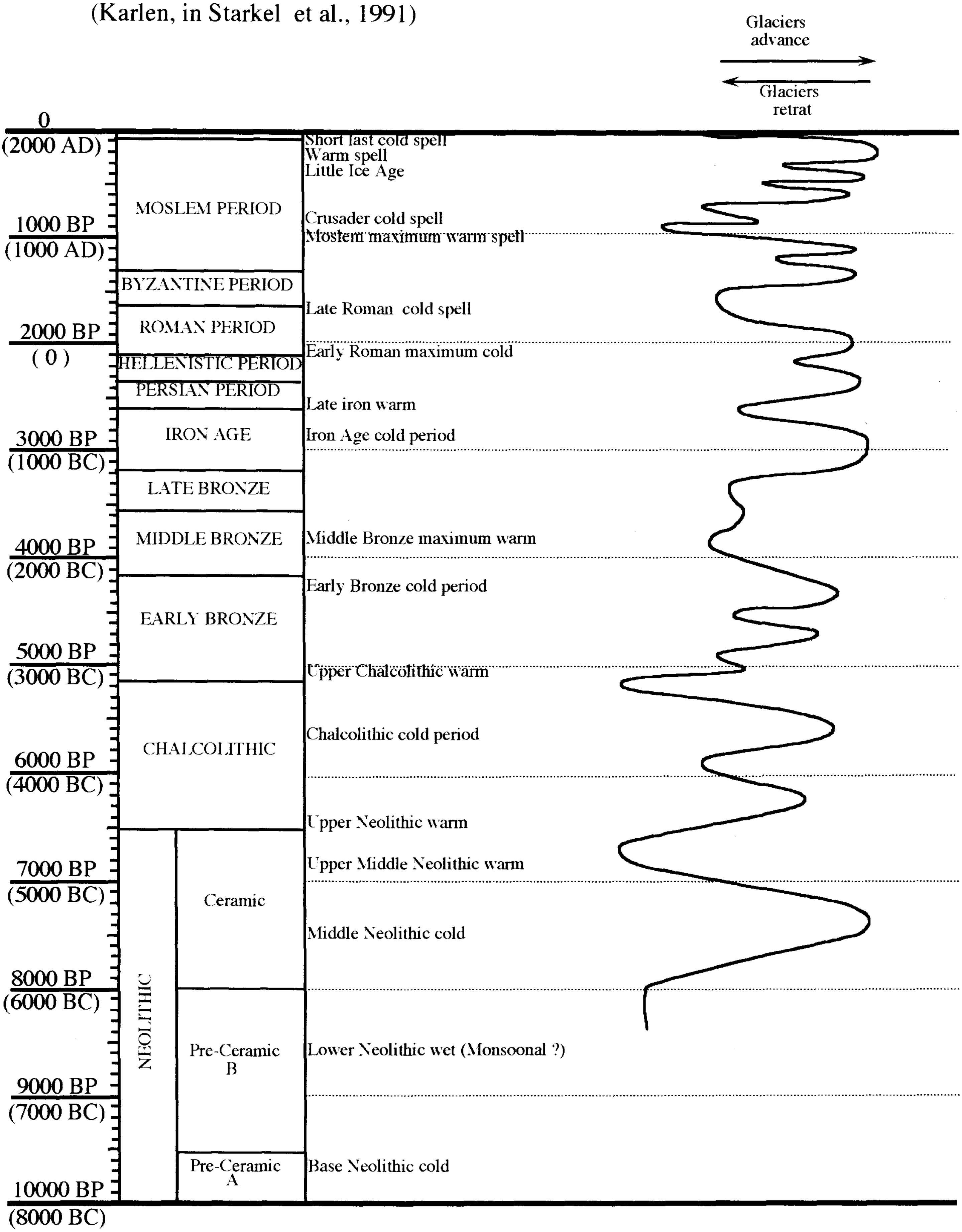


Fig.12 - Correlation of abundance of hazel pollens (Bennett, 1983) with the Levant base curve.

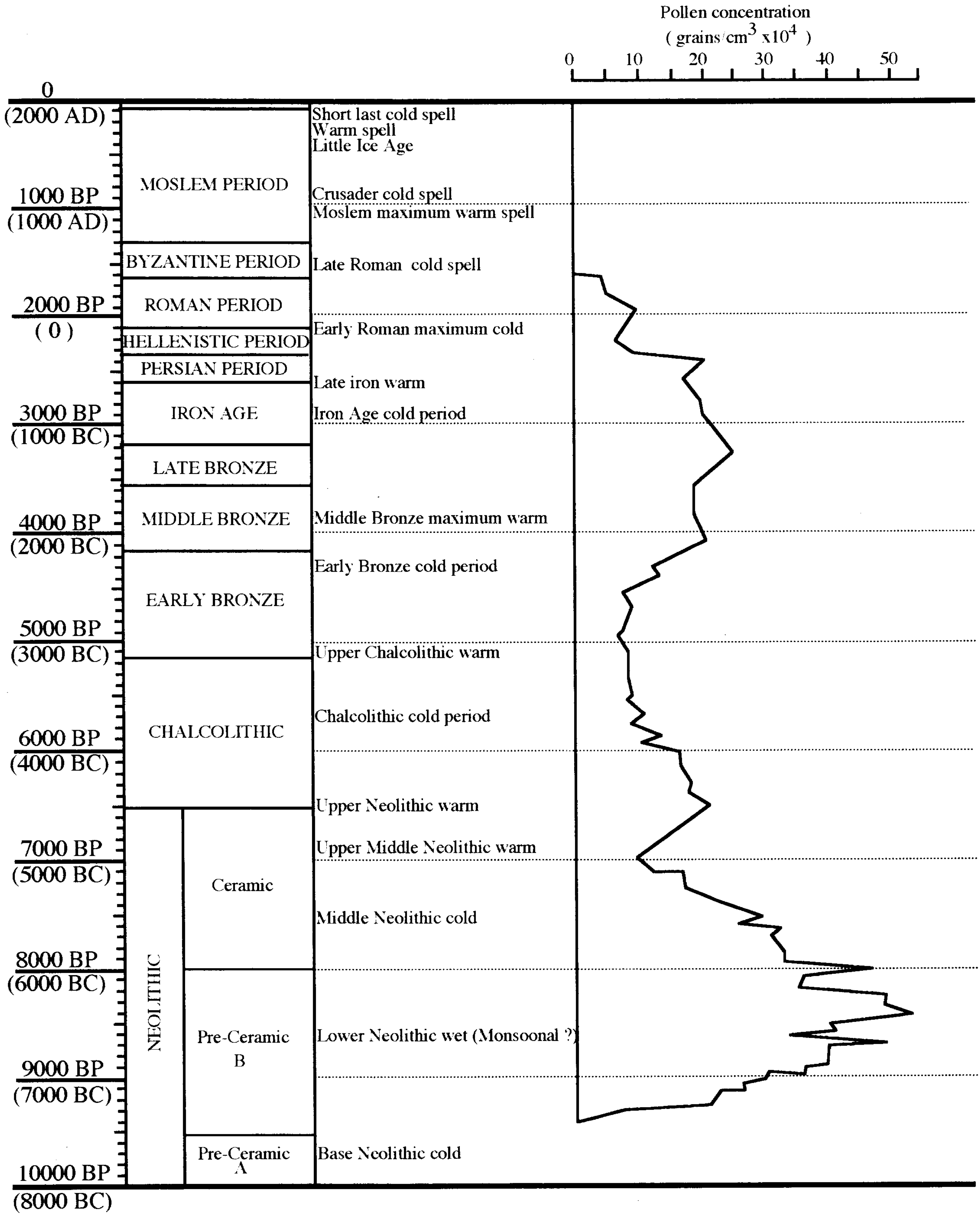


Fig.13. Correlation between coastal changes in the Netherlands and the Levant base curve (Louwe-Kooijmans, 1980)

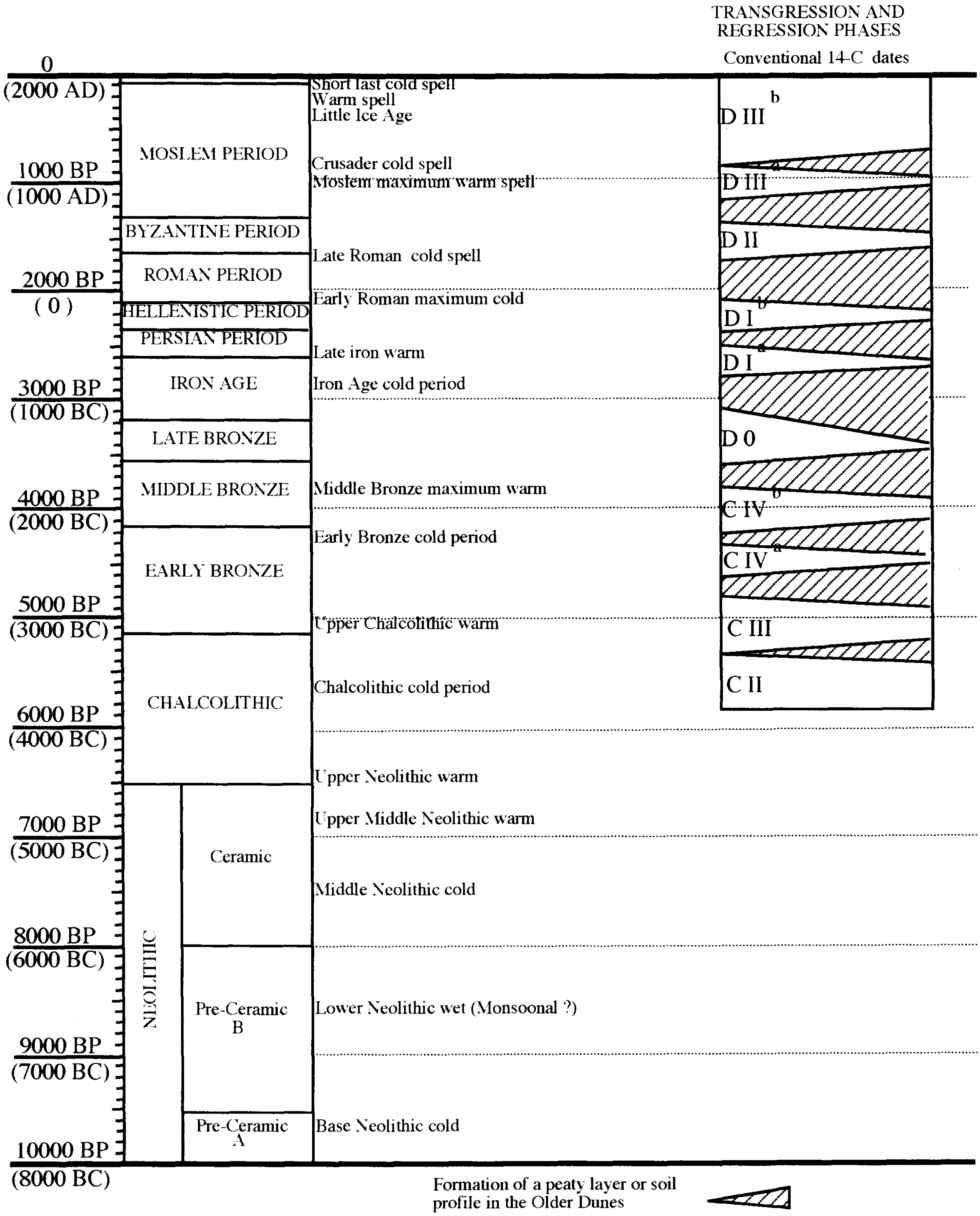


Fig.14 - Correlation of Black Sea levels and the Levant base curve, (according to Ghenea and Mihailescu, 1991, combined with Degens and Ross, 1972, and Panin et al., 1983)

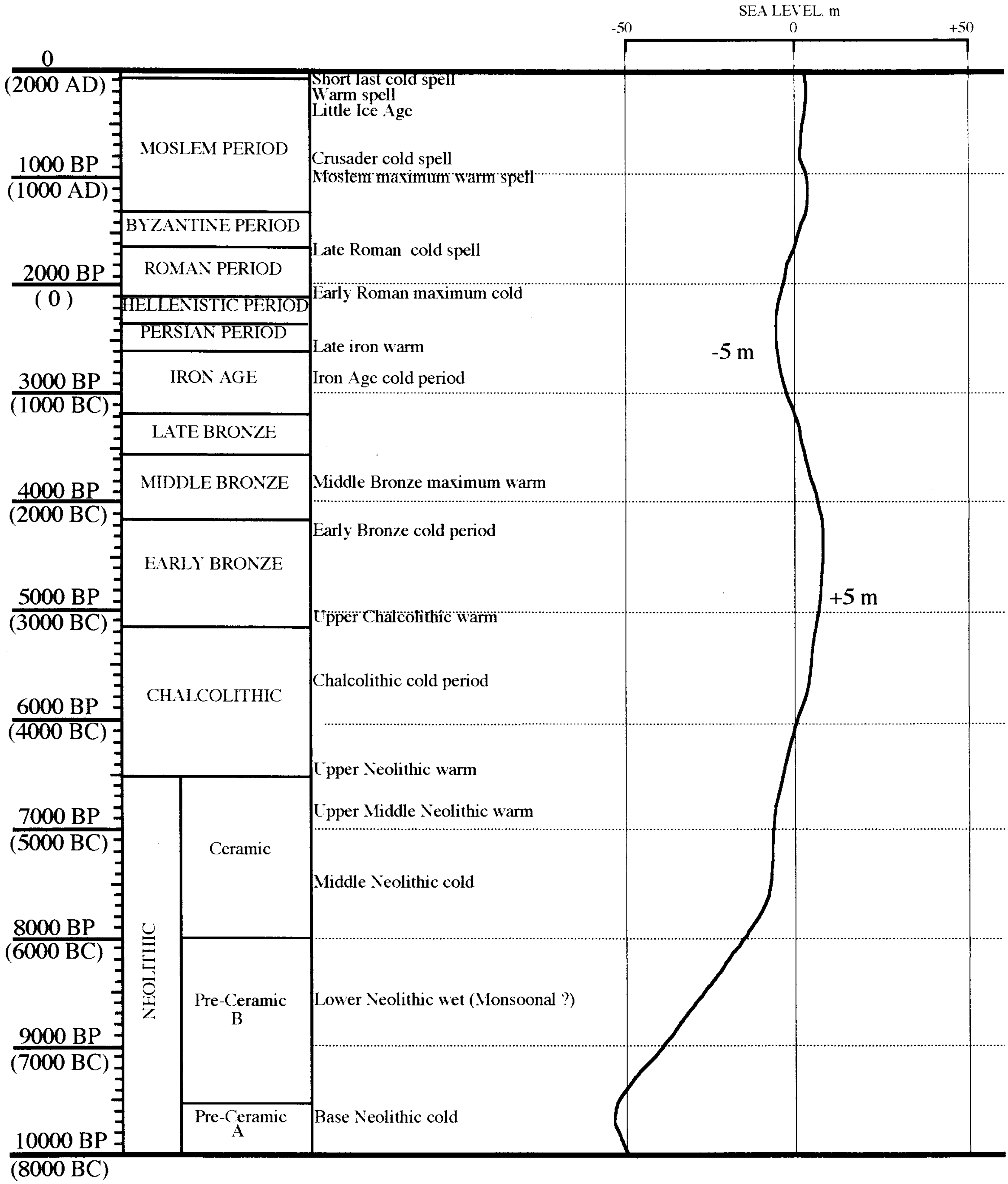


Fig.15 .Correlation of temperature variations of the Qinghai-Xizang Plateau (Fu Bau and Fan, 1987) and the Levant base curve.

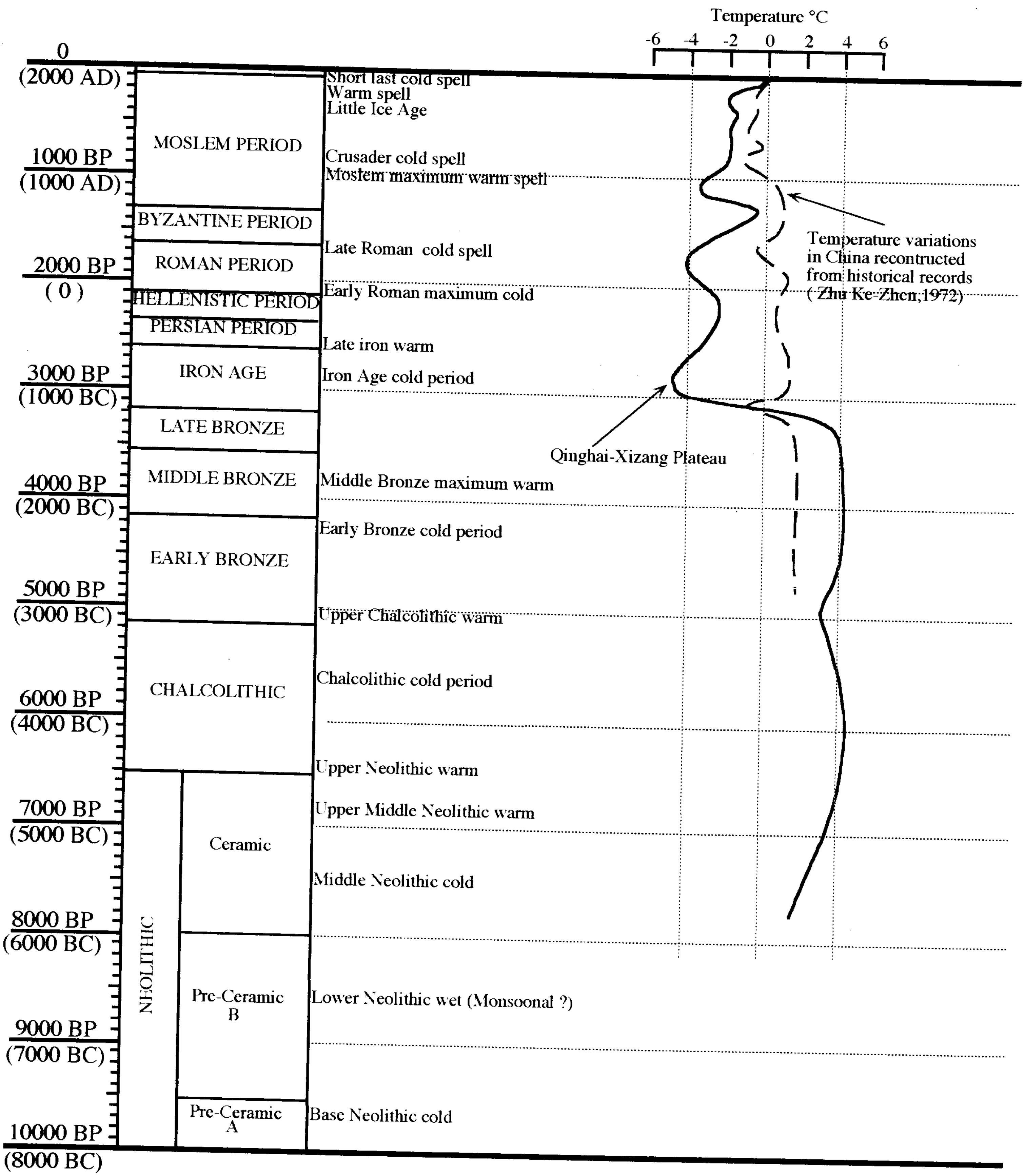


Fig.16. Correlation of the nomadic invasions from north with temperature, frequency of droughts, and intensive desertification stages since 200 BC in China (Fang Jin Qix,1990)

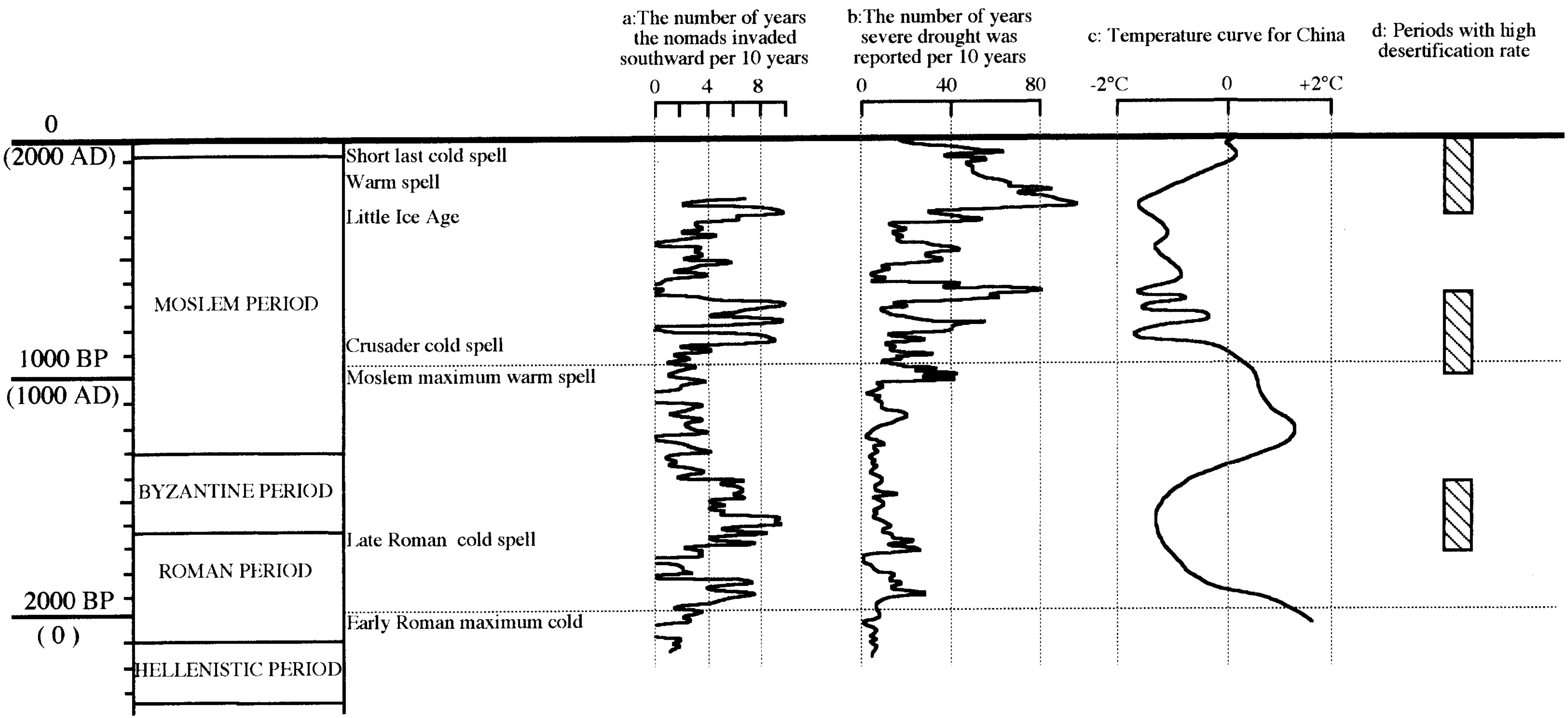


Fig.17. Holocene temperature changes at some sites in China (Wang et al., 1990)

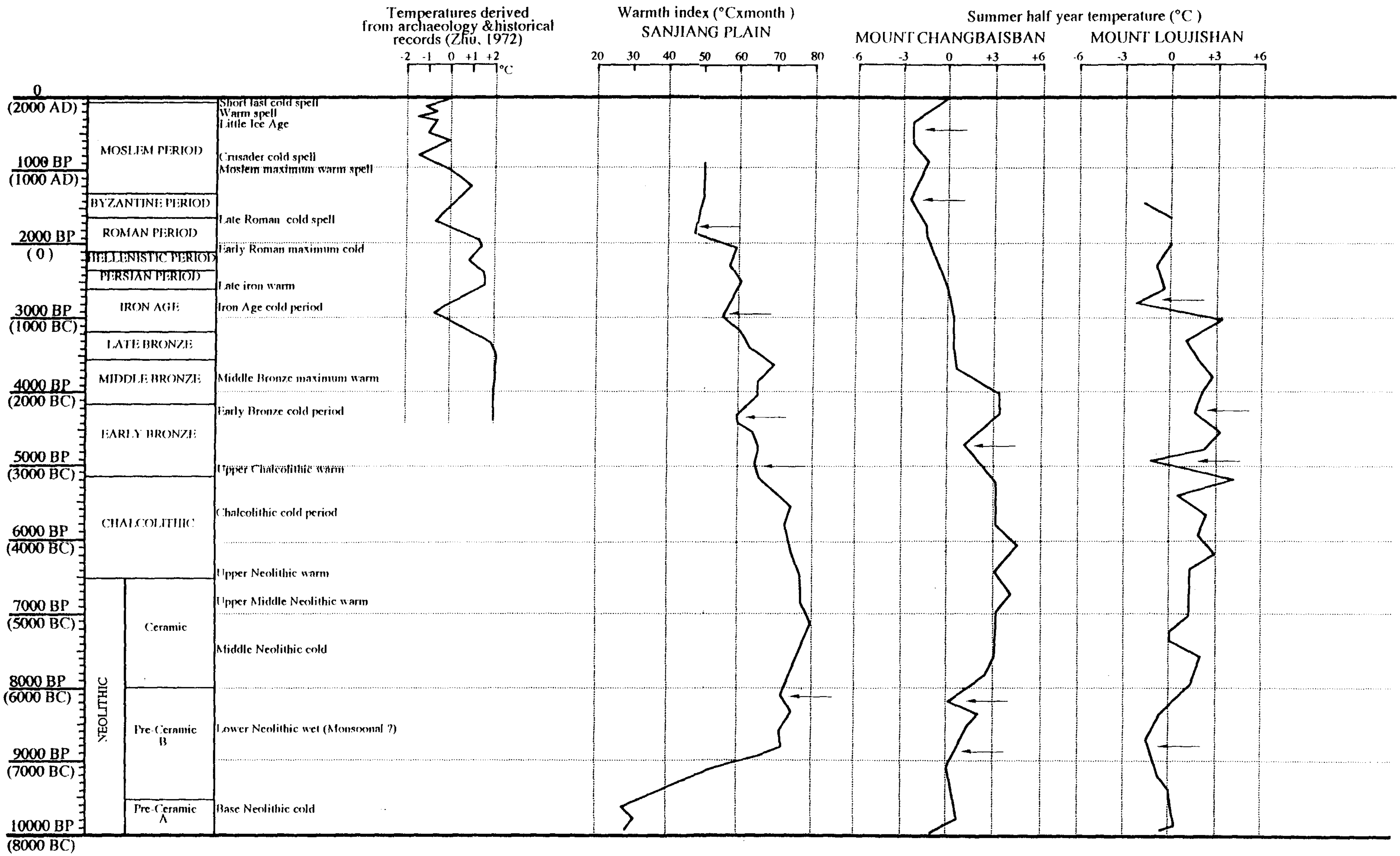


Fig.18. Correlation between tree rings' curve of the bristole cone pine in California (Lamb, 1982) and the Levant base curve.