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MOUNTAIN BUILDING AND CLIMATE: MECHANISMS AND TIMING

ABSTRACT: OLLIER C.D., *Mountain building and climate: mechanisms and timing*. (IT ISSN 1724-4757, 2004).

Mountains affect climate in several ways. At a local scale, passive effects include simple elevation and rain shadow effects. On a broader scale and over a longer period, more substantial effects result from different mechanisms. In the past few million years, world wide uplift of mountains in the Neotectonic Period has actively forced climatic change. The uplift of the Tibet Plateau and its bordering mountains had global effects, through effects on the Asian monsoon, jet streams, and inter-hemispheric exchange. The hypothesis of the negative greenhouse effect is not supported by the dating of climatic change, or relationship between carbon dioxide, weathering and erosion. Antarctica has been long isolated by the Antarctic Circumpolar Current and does not share the same tectonic and climatic history as the rest of the world.

KEY WORDS: Mountains; Climate; Neotectonics; Weathering; Monsoon; Antarctica.

RIASSUNTO: OLLIER C.D., *Rapporti fra la costruzione delle montagne e il clima: meccanismi e tempi*. (IT ISSN 1724-4757, 2004).

Le montagne influenzano il clima in vari modi. A scala locale, stanno gli effetti passivi dell'altitudine e dell'esposizione sulle piogge. A scala più ampia e su periodi più lunghi, effetti più sostanziali derivano da alcuni differenti meccanismi. Negli ultimi milioni di anni, il vasto sollevamento neotettonico delle montagne alpine ha attivamente forzato i cambiamenti climatici. Il sollevamento dell'altopiano del Tibet e delle sue montagne periferiche ha avuto conseguenze globali negli scambi tra gli emisferi, attraverso gli effetti del monsone asiatico associato alle variazioni delle correnti a getto. L'ipotesi di un negativo effetto serra non è sostenuta dalla datazione del cambiamento globale del clima, come pure da una relazione tra la variazione dell'anidride carbonica, la degradazione meteorica e l'erosione. L'Antartide è stata a lungo isolata dalla Corrente Circumpolare Antartica e non ha partecipato alla stessa storia tettonica e climatica del resto del mondo.

TERMINI CHIAVE: Montagne; Clima; Neotettonica; Degradazione meteorica; Monsone; Antartide.

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INTRODUCTION

Mountains affect climate on local or regional scales, and also on the global scale. The passive effects of mountains on climate are well known and described in many books, such as Barry (1981) and Lamb (1972). These books take the mountains as a given fact, and even in discussion of older climates, such as the Cenozoic ice age, it is tacitly assumed that the mountains have always been with us Variations are seen as resulting from astronomical events, such as the Milankovich cycles, or in more recent years as related to atmospheric changes, especially in greenhouse gases.

The prevalent view today is that astronomical causes affect minor cycles, but the cause of the major changes, such as initiation of ice ages, remains controversial. Some advocate changing CO_2 content of the atmosphere, others changes in distribution of land and sea associated with plate tectonics.

In recent years new ideas have come along about the active effects of mountain uplift on climate, and even the effects of climate on mountain uplift. The age of uplift is often controversial, as are the mechanisms by which uplift affects climate. This paper reviews the climate/mountain inter-relationship. Basic concepts are briefly described, and the emphasis is on controversial topics.

PASSIVE EFFECTS

Since the air becomes less dense with increasing altitude it is able to absorb less heat, so increasing altitude results in cooling, and mountains are cooler at the top. If a mountain is high enough, its cold weather will affect its surroundings, and many high mountains will cause regional climatic effects. This is a simple passive effect. The simple temperature effect of elevation is complicated by the relationship of mountains to prevailing winds. Orographic precipitation is caused by the forced ascent of air over high ground, which leads to cooling, condensation and precipitation. Rainfall is concentrated on the windward side and a drier area, the rainshadow, appears on the lee side.

Mountain barriers over about three kilometres high have a significant effect on atmospheric circulation, especially if the mountain chain is oriented perpendicular to dominant winds. The Rockies, for example, effectively prevent a great deal of marine air from penetrating inland, and protect the western seaboard from cold air invasions originating in Arctic Canada. The Andes play a similar large role in the climates of South America. Rain shadows are a very significant factor, but will be discussed later.

ACTIVE EFFECTS

When new mountains appear they have a direct effect on local climate, and the impact grows as mountains increase in elevation and area.

Volcanoes

The most obvious change is associated with the construction of new volcanoes. A single volcano, such as Mount Etna, which was erupted entirely in the Quaternary, creates its own local climate. In many areas the Andes consist of a plateau, the altiplano, surmounted by many large stratovolcanoes such as Cotopaxi and Chimborazo, which are all of Quaternary age. The altiplano is itself high enough to create large climatic effects, and the volcanoes enhance this. They increase the elevation, often above the snowline, with a very marked effect on local and regional climate. This is especially so where the volcanoes are closely spaced and have a collective effect on climate greater than that of a single volcano.

Plateau uplift

We can also consider the change in climate caused by the appearance of new mountains. Pre-existing plains or lowlands may be uplifted, as on flanks of rift valleys or passive continental margins, and induce new climates. Partridge (1998) wrote: «Uplift of 1000 m [in southern Africa] is equivalent, in its effects on surface temperatures, to the cooling experienced during an Ice Age in higher latitudes».

He thinks the close coincidence of South African uplift with a global period of cooling and increasing aridity between 2.8 and 2.6 Ma would have amplified its effects. Among the landscape responses to this were the creation, for the first time, of the Kalahari Desert, expansion of grasslands, and the changes in antelope and other faunas that occurred in this area in the late Pliocene.

Many other places, to be described later, moved into high elevation in the Quaternary, affecting both landforms and climate.

Rain-shadow effects in active areas

Mountains induce orographic rainfall, which also leads to reduction of rainfall on the lee side of the mountain barrier. Fossil evidence may reveal the climatic change. A classic example is provided by the Sierra Nevada in California. This is a tilted fault - block mountain range, about 100 km by 600 km, high on the eastern fault scarp edge and sloping gently to the west (fig. 1).

Axelrod (1962) used biogeographical evidence to determine the time of uplift of the Sierra Nevada. In the early Pleistocene a similar vegetation, a pine - fir ecotone, was established right across the whole region, a situation that could only exist if the present climatic barriers were absent, so the major uplift occurred later in the Pleistocene. At present the Owens Valley, east of the Sierra Nevada, has a very different climate from the lowlands to the west of the range, and the arid Death Valley is a little further east. These contrasting climates were created by Quaternary uplift of the Sierra Nevada tilt block.

It must be noted here that some workers think the uplift of mountains in the western USA occurred earlier. Wolfe & *alii* (1997) suggest high altitudes in Nevada in the Miocene, and Wolfe & *alii* (1998) suggest high altitudes in the Eocene and Oligocene. The conclusions are based on a technique of using leaf shapes as a surrogate for altitude. Botanists I have asked about the method consider it very dubious, and I feel it should not over-ride the species based conclusion of Axelrod.

Aspect is important in climatic and erosional effects on large volcanoes. The north-eastern slopes of the larger Hawaiian islands may be incised by deep canyons because of the high rainfall, whereas leeward slopes remain relatively undissected. The rate of erosion on a volcano may be reduced by the growth of another volcano to the windward, as happened when the Koolau dome cut off the trade winds from the Waianae dome, on Oahu, the main island of Hawaii (Ollier, 1988).

Snow lines and tectonic uplift

In principle the snowline is at high altitude at the equator, and progressively lower towards the poles. The timber line, and other climatic indicators, follow at a lower elevation. Fleming (1979) has documented timberline and snowline changes between Antarctica and New Guinea via New Zealand. He drew diagrams showing the location of the snowline at different times (fig. 2).

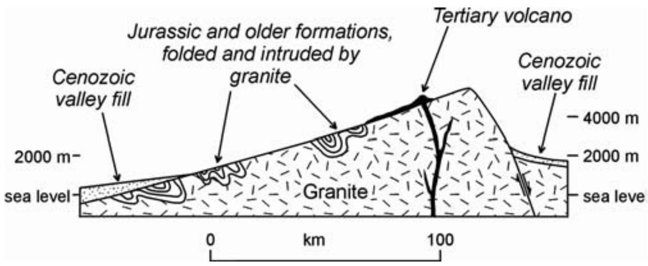

FIG. 1 - Cross section of the Sierra Nevada. Death Valley is to the right of the section. Two million years ago there was a single ecosystem across the whole region (after Axelrod, 1962).

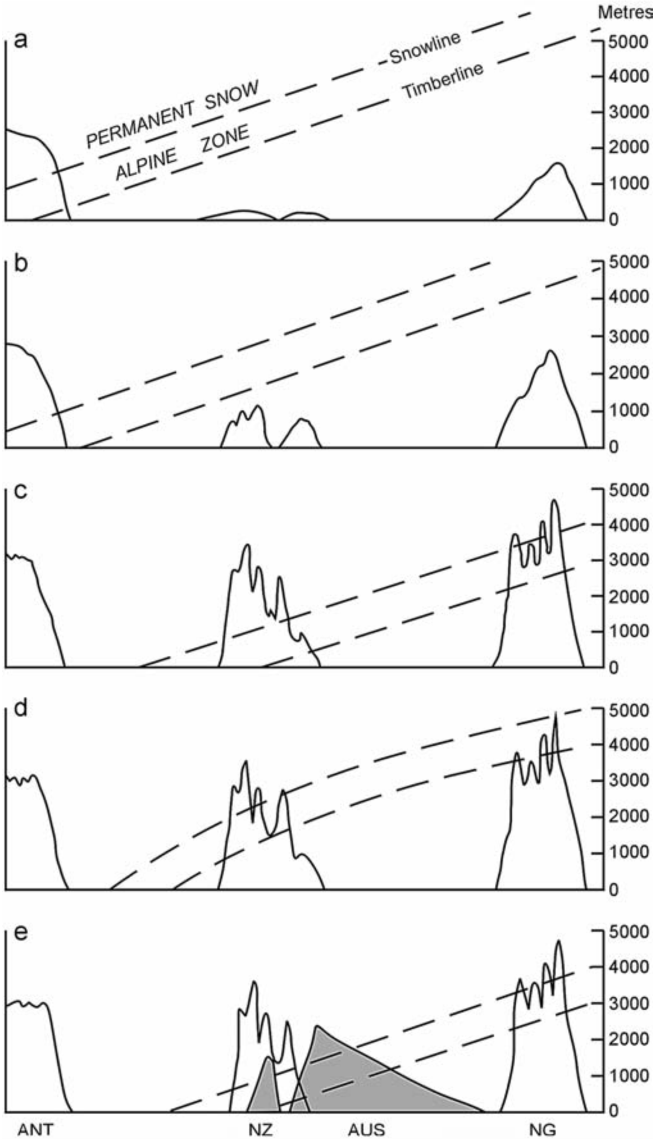

FIG. 2 - Changes in the snowline and timberline elevation with latitude, and its variation through time (after Fleming, 1979). The horizontal scale is diagrammatic. a. mid-Tertiary; b. Pliocene; c. Last Glaciation; d. Present; e. Present, with Australia fitted on to the Fleming snowline (after Ollier, 1986). For explanation see text. A = Antarctica; NZ = New Zealand; AUS = Australia; NG = New Guinea.

In mid-Tertiary time Fleming's climatic zones at sea level are based mainly on marine evidence, because at this stage New Zealand and New Guinea had hardly appeared above sea level. In the Pliocene the climatic zones were almost unchanged at sea level, but New Zealand and New Guinea had grown higher, reaching different climatic zones. By the time of the last glaciation the islands had achieved their full present height, and furthermore the climatic belts had moved, producing extensive glacial and alpine zones in the islands. The uplift of New Zealand and New Guinea clearly had a dominant effect on their climate in the Pleistocene.

The climatic gradients determined from New Zealand and New Guinea fail to account for the situation in Australia. If Fleming's snowline is drawn across Australia it is evident that Australia is high enough to have been glaciated throughout the Quaternary cold periods. Yet mainland Australia experienced only the last glaciation. The suggestion is that it was not high enough during earlier glaciations, supporting the idea of Neotectonic uplift in the late Quaternary (Ollier, 1986).

WEATHERING AND NEGATIVE GREENHOUSE EFFECTS

Ruddiman and his colleagues (Ruddiman & Kutzbach, 1991; Raymo & Ruddiman, 1992; Ruddiman, 1997) have been major proponents of the hypothesis that mountain building controls climate through a negative greenhouse effect. The proposal may be summarised as follows:

Increased tectonic uplift over the last 40 million years leads to greater erosion and weathering. Weathering of rocks is by carbonation and removes carbon dioxide from the atmosphere. The reduction in carbon dioxide causes a negative greenhouse effect and leads to global cooling, and eventually to glaciation.

There appear to be three major misconceptions in this proposed interrelationship of erosion, weathering and carbon dioxide.

1. Erosion and chemical weathering are two different processes that do not necessarily occur together. Erosion can proceed, as it often does in mountainous regions, with little chemical alteration of the rock or mineral fragments that are loosened and transported. Indeed, immature, so-called labile sediments with large amounts of unweathered mineral and lithic fragments are generally assumed by sedimentologists to be derived from the rapid erosion of mountains.

A different view is expressed by Riebe & *alii* (2001), who claim chemical weathering correlates strongly with physical rates but only weakly with climate. Their conclusions are based on elaborate chemical assumptions, but their understanding of deep weathering seems suspect when they write: «The average [Zr] in saprolite (i.e. the chemically altered but physically intact bedrock at the base of the soil is roughly the same as average [Zr] in rock outcrops… indicating that the saprolite has undergone little chemical depletion…» But saprolite is formed from fresh rock by constant-volume alteration, and the difference in density between rock and saprolite can be used to calculate losses. This is the basis of the standard cell method of calculation of Barth (1948). More primitive calculation methods often use Zr as a basis to calculate losses of other elements, as it is generally regarded as a very immobile element.

2. Weathering is not synonymous with carbonation, that is the interaction with CO_2, as expressed by Raymo & Ruddiman (1992) in the following simple formula:

chemical weathering

$$CaSiO + CO_2 \rightarrow CaCO_3 + SiO_2$$

This oversimplification is found in many elementary texts, but is incorrect. In most situations *hydrolysis* is by far the most important process of silicate weathering (Ollier, 1984; Trescases, 1992; Eggleton, 1998). Weathering produces clays, not carbonates. Eggleton writes: «During weathering protons are added to the solid phases as other cations and oxygen are lost in solution». The following is one of the formulae provided by Eggleton.

$$2KAl_2[Si_3Al]\ O_{10}(OH)_2 + 4H_2O \rightarrow 3Al_2Si_2O_5(OH)_4 + K_2O$$

muscovite water kaolinite

Ollier (1984) represents the weathering of an alumino-silicate mineral as follows, where M represents any metal ion and SiO represents any silica group:

$$MalSiO + H_2O \rightarrow M^+ + HalSiO + OH^-$$

Al-silicate water cation clay hydroxyl ion

The real-world evidence from deep chemical weathering seems to be the exact opposite of that postulated by Ruddiman and colleagues. The greatest deep weathering profiles all around the world are of Mesozoic or early Tertiary age. This was a period of broad plains with very little mountain building on a global scale. Since then uplifted plateaus have been associated with stripping of the old regolith to make etchplains (Ollier & Pain, 1998). In general there appears to have been a reduction in chemical weathering since the mid- or early Tertiary, and present-day deep chemical weathering is confined to the humid tropics.

3. A further problem arises from Ruddiman's erroneous time scale. Originally Ruddiman & Kutzbach (1991) wrote: «Prior to 40 million years ago most of the world was warmer and wetter than it is now». «During the past 40 million years, and particularly during the past 15 million years, this warm wet climate largely disappeared». «Significantly the three rivers that currently carry the highest loads of dissolved chemicals into the ocean… all drain regions that have been uplifted in the past 40 million years». They repeatedly stress 40 million years. They maintained that uplift of Tibet, the highlands of western North America such as the Colorado Plateau, and the Andes started about 40 million years ago.

But the Cenozoic ice-Age is essentially a Quaternary event, with precursors in the late Neogene. Glaciation in the northern hemisphere started about 3 million years ago (Imbrie & Imbrie, 1979). Mountain building 40 million years ago, or even 15 million years ago, would seem to be a remote relationship. Their suggested timing now seems wrong, as we shall see, and they were not aware of widespread young uplift in many other parts of the world. The connection between climatic cooling and mountain building is correct in principle, but a time scale of 5 million years seems more probable.

THE NEOTECTONIC PERIOD AND GLOBAL CLIMATIC EFFECTS

It is commonly assumed that folding and mountain building go together, but this need not be true. Mountains occur not only on folded rocks, but on horizontal rocks, granite, and lava flows. Ollier & Pain (2000) assembled evidence that most mountains are the products of uplift of a plain to form a plateau, which may or may not be extensively dissected. The age of a mountain or mountain range is then the age of plateau uplift, not the last age of folding of rock. On this basis, a table of time of uplift of mountains and plateaus from around the world (tab. 1) shows a preponderance of uplift in the last few million years. This uplift of mountains in the past few million years appears to be a global phenomenon. It affects so-called Alpine mountains, mountains on passive continental margins, and in deep continental interiors. The period of uplift is known as the Neotectonic Period (Ollier & Pain, 2000). Earlier, Morner (1992), after summarising much earlier work, wrote «an important neotectonic-paleoclimatic linkage is hereby advocated». Zhu (1997) is another who believes there is a «coupled climatic-tectonic system, because of the close relationship between climatic change and neotectonic movement». According to him the first uplift of the Tibet Plateau occurred between about 3.5 and 2.4 Ma.

Ollier & Pain (2000) proposed that since the Neotectonic Period coincides roughly in time with the Cenozoic ice age, perhaps orographic change caused the climatic change. If the uplift of Tibet, the Andes and many other highlands occurred mainly in the Pliocene and Pleistocene the correspondence between uplift and climatic change is much greater than Ruddiman and his colleagues proposed. The orographic time scale matches the glaciation time scale much better than the 40-20 Ma of Ruddiman and colleagues. The correspondence of Tibetan uplift with the onset of the monsoon climate and the deposition of loess is the best example of detailed correlation.

As an example, the timing of uplift in Tibet and its bordering mountains will be looked at in a little more detail.

Gansser (1991) wrote: «... we must realise that the morphogenic phase is not only restricted to the Himalayas but involves the whole Tibetan block. This surprising fact shows that an area of 2,500,000 km^2 has been uplifted 3-4,000 m during Pleistocene time and that this uplift is still going on». In places the uplift rate is 4.5 mm/y (five times the maximum in the European Alps).

From the Pliocene to the early Quaternary (5-1.1 Ma) the Kunlun Pass area of the Tibetan Plateau was no more than 1500 m high and the climate was warm and humid (Wu & *alii*, 2001).

«The extreme geomorphic changes in the Kunlun Pass area reflect an abrupt uplift of the Tibet plateau during the Early and Middle Pleistocene. The Kunlun-Yellow River tectonic movement occurred 1.1-0.6 Ma». Zheng & *alii* (2000) concluded from sediments at the foot of the Kunlun Mountains that uplift began around 4.5 Ma.

For the Himalayas, Gansser puts the uplift as Pleistocene. Japanese workers on the Siwaliks, deposits in a

TABLE 1 - Some suggested ages for mountain uplift. In many areas there are precursor movements, and the ages in Table 1 generally refer to the major or latest uplifts

EUROPE		
Swiss Alps	(Trumpy, 1980)	Pliocene - Quaternary
Jura	(Holmes, 1965)	Pleistocene
Apennines	(Coltorti & Pierruccini, 2000)	latest Pliocene - Middle Pleistocene
Pyrenees	(Sala, 1984a; Calvet, 1994)	Pliocene
Central Cordilleras of Spain	(Sala, 1984 b)	Plio-Pleistocene
Baetic Cordillera	(Choubert & Faure-Muret, 1974; Sala, 1984c)	upper Miocene - Pliocene
Western Carpathians	(Földvary, 1988)	Upper Miocene - Pliocene
Eastern Carpathians	(Zuchiewicz, 2000)	Pliocene
Southern Carpathians	(Radoane & *alii*, 2003)	2,500 m about 12Ma; 1000 m about 2 Ma
Caucasus	(Bridges, 1990)	Upper Pliocene
Ural Mountains	(Bridges, 1990)	Pliocene-Pleistocene - Middle Pleistocene
Sudeten	(Migon & Lach, 1999)	Pliocene-early Quaternary
ASIA		
Tibetan Plateau	(Wu & *alii*, 2001)	Pliocene - Quaternary
Himalayas	(Zhang, 1998; Kalvoda 1992; Gansser, 1991)	Pliocene - Quaternary
Kunlun Mountains	(Wu & *alii*, 2001; Zheng & *alii*, 2000)	late Pliocene-Quaternary
Tien Shan	(Holmes, 1965)	Quaternary
Pamir	(Strecker & *alii*, 2003)	late Cenozoic
Altai Mountains	(Suslov, 1961)	Tertiary
Transbaikal Mountains	(Ufimtsev, 1990)	mid - Tertiary
Karakoram	(Schroder, 1993)	Late Neogene to present
Shanxi Mountains	(Li & *alii*, 1998)	Miocene - middle Pleistocene
Japanese Mountains	(Hoshino, 1998)	Pliocene - early Pleistocene
Taiwan	(Chai, 1972; Ho, 1986)	Early Pleistocene
NORTH AMERICA		
Sierra Nevada	(Axelrod, 1962)	post-Pliocene
Basin and Range	(Nitchman & *alii*, 1991)	4 Ma
Colorado Plateau	(Lucchita, 1979; Trimble, 1980)	Late Pliocene to Recent
Bighorn Mountains	(Thornbury, 1965)	middle Tertiary - Pleistocene
Rocky Mountains	(Eaton, 1987)	5 million years
Coast Ranges	(Thornbury, 1965)	late Pliocene
Canadian Cordillera	(Mathews, 1991)	Late Miocene - Pliocene
Canadian Coast Ranges	(Farley & *alii*, 2001)	modern topography post 2.5 Ma
Cascade Range	(Priest & *alii*, 1983)	4-5 Ma
SOUTH AMERICA		
Colombia	(Kroonenberg & *alii*, 1990)	Plio-Pleistocene
Chile	(Holingworth & Rutland, 1968)	Pliocene and Pleistocene
Bolivia	(Walker, 1949)	Plio-Pleistocene
Ecuador	(Coltorti & Ollier, 1999)	Upper Miocene - Plio-Pleistocene
OTHER REGIONS		
Ethiopian Rift	(Partridge, 1997)	2.9 and 2.4 Ma
Western Rift	(Pickford & *alii*, 1993)	3 to 2 Ma
Ruwenzori	(Partridge, 1997)	within the last 3 million years
New Guinea Mountains	(Ollier & Pain, 1988)	Plio-Pleistocene
New Zealand Mountains	(Suggate, 1982)	Pliocene
PASSIVE MARGINS		
East Australia	(Ollier & Taylor, 1988)	Pleistocene
Appalachians	(Pazzaglia & Gardner, 2000; Stanford & *alii*, 2001)	Miocene or younger
South Africa	(Partridge, 1998)	Pliocene (7-900m)
Western India	(Widdowson & Gunnell, 1999)	late Tertiary
Greenland	(Weidick, 1976)	late Miocene or younger
Antarctica	(Behrendt & Cooper, 1991)	Pleistocene
Scandinavia	(Lidmar-Bergstrom & *alii*, 2000; 2002).	Neogene
Brazil	(Martins & *alii*, in press)	post-Miocene

sedimentary basin filled with erosion products from the Himalayas, found that fine sediments give way to a boulder conglomerate at about 1 million years, indicating a time of major uplift (Prof. T. Kosaka, *pers. Comm.*).

The Tibet plateau area had two earlier uplifts in the middle Eocene and early Miocene, recorded in planation surfaces. The strongest uplift of the plateau and its bordering mountains, the so-called Qinzang (Tibet) movement, occurred between 3.6 and 1.7 Ma (Li & *alii*, 1996) and had three phases commencing:

A. 3.6 Ma
B. 2.5 Ma
C. 1.7 Ma

It was phase B of the Qinzang movement at 2.6 Ma that raised the plateau to the critical height of 2,000, triggering the onset of the monsoon and loess deposition. Phase C of the Qinzang movement at 1.7 Ma caused a large geomorphological adjustment and produced the present geomorphic configurations including the large rivers such as the Huang He and the Chang Jiang.

Still later movements are the Kunlun-Huang He (or Kunhuang) movement which occurred between 1.1 and 0.6 Ma, and the Gonghe movement which occurred after 0.15 Ma (Li and Fang, 1999). The Kunhuang movement lifted the plateau to an average height of 3,000 m with mountains to over 4000 m, a critical height for glacial development. Since then the plateau has undergone several glaciations. The Gonghe movement raised the plateau to its present height. The Himalayas rose to over 6000 m and became a major barrier for the inflow of the Indian monsoon onto the Plateau, leading to further drying of northwestern China.

Overall, the Quaternary uplift of the Tibet Plateau and the Himalayas introduced a powerful new geographical factor in the pattern of climate. In the Early and Middle Pleistocene when the average elevation of the Himalayan range was about 4400m, the evidence from interglacial deposits shows that the north side was as warm as the south side at similar elevations. The uplift of the range to its present 6000m average elevation made the Himalayas a much more effective climatic barrier, preventing warm, moist air from entering the Tibetan Plateau.

Mountains on passive continental margins differ from others in several respects (Ollier, 2004). They may have been originally high, like the high plateaus bounding many present-day rift valleys, or uplifted at the time new continental margins were formed. In some models, the passive margin is dominated by flexture of an old land surface (palaeoplain), with uplift on land and subsidence offshore. The palaeoplain may be equated with the breakup unconformity beneath marine sediments offshore (Ollier & Pain, 1997). In this situation offshore sediments may be used to determine time or times of uplift. Most passive margins appear to have two periods of movement, one around the time of creation of the new continental margin, and a later one.

Partridge (1998) proposes Mesozoic uplift in the creation of the southern African landscape, and also Pliocene uplift of the South African high plains of up to 800 m. He wrote: 'The evidence for large-scale Neogene uplift is now beyond question». In Australia the eastern highlands are associated with a palaeoplain of Trias-Jura age (Hills, 1975). Uplift was attributed once to the Plio-Pleistocene "Kosciusko Orogeny" (Andrews, 1910). This idea was replaced by general belief in early Cenozoic uplift, but some movements of up to 1 km may have occurred in the Pleistocene (Ollier & Taylor, 1988). In western India Widdowson and Gunnell (1999) showed several phases of laterite formation on the coastal plain. The elevation of the coastal laterite (up to 200 m) together with associated development of an entrenched drainage indicates that widespread uplift has affected the margin during late Tertiary times. Martins & *alii* (in press) present evidence of late Miocene or younger uplift in southeastern Brazil.

In the Appalachians the palaeoplain might date back to the Cretaceous but there is also evidence of Miocene or younger uplift, especially in the Piedmont province (Pazzaglia & Gardner, 2000; Stanford, 2001). The Scandinavian margin had continuous ulift from the Mesozoic, but in southern Norway there was renewed uplift of about 1000 m in Neogene time (Lidmar-Bergstrom & *alii*, 2000; Lidmar-Bergstrom and Näslund, 2002). In Greenland the highest and oldest planation surface cuts across late Miocene basalt, so uplift is later than that (Weidick, 1976).

The Transantarctic Mountains may have experienced major uplift since the early or middle Pleistocene (Behrendt & Cooper, 1991), or may have remained at their present level since the Miocene (Kerr & *alii*, 2000).

MONSOON EFFECTS

The monsoon is a climatic type which some regard as an exaggerated form of the tropical summer rain climate. Major wind reversals are part of the monsoon system. The Indian Monsoon is important south of the Himalayas, and the East Asian Monsoon east of the Tibetan landmass.

On a homogenous globe the planetary circulation would be the same in both hemispheres. Mainly because there is more land in the northern hemisphere, the two hemispheres have different circulation. The northern hemisphere has greater seasonal contrast. The southern circulation is more vigorous, and when the thermal equator migrates northwards in June and July the southern hemisphere circulation is able to encroach over the equator. In the reverse season migration of the inter-tropical convergence zone into the southern hemisphere is limited.

Monsoon-type wind and seasonal changes are found in several continents, but the Asian one is the most intense and has an apparently global impact. Much of the literature concerns the Indian Monsoon, but the East Asian Monsoon may be even more significant. This monsoon is related to the deposition of wind-blown loess in China, and this uniquely stratified material provides a splendid record of climatic change.

The classic explanation of the monsoon goes back to Halley (1686) and remained the dominant idea for almost three hundred years. The idea is almost like the concept of sea and land breezes extended to a continental scale. In brief, a summer low-pressure cell develops over land because of solar heating, and this induces moist air to blow inwards from the ocean. In winter the system is simply reversed, and cold air blows outwards from the Asian continent. In reality the explanation of the monsoon is very complex (see for instance Lockwood, 1974; Krishnamurti, 1978). A simplified general description is provided below.

Besides the winds near the ground, which constitute the monsoon, the atmosphere is affected by jet streams at higher level. They flow generally from west to east, but the location of the principal jet streams appears to be substantially influenced by the topography of the Tibet Plateau and its associated mountains. Since the jet streams mainly occur at an elevation of about 10 km and the mountains only rise to half this elevation, it is not direct topography but possibly a thermal high over Tibet that deflects the jet streams. One jet stream occupies a position just south of the Himalayas and another skirts around the north of the Tibetan Plateau. These two jet streams become confluent east of the Himalayas in China.

In spring the westerlies begin their northward seasonal migration. The northerly jet stream intensifies at the expense of its Himalayan counterpart, but they both retain their mean geographic location because of topographic constraints. The southern jet stream declines until finally, over a period of a few days it disappears, and the summer monsoon of India finally advances. In late September or October the equatorial trough of low pressure weakens and retreats south. At high level the Himalayan jet stream reappears rather suddenly in mid-October.

Fang & *alii* (1999 b) provide a detailed explanation of the monsoon climate, and their map (fig. 3) shows the location of the westerly jets. The pattern is clearly influenced by the Tibetan block. They say that uplift of the Tibet Plateau not only generated the Tibetan Plateau monsoon, but also intensified the northern hemisphere temperature gradient. Fang & *alii* (1999 a) further suggest that before the uplift of the Tibet Plateau there was a single jet stream with a course right across the position of the present plateau (fig. 4). The final uplift that caused the last change in winds (and a change in the composition of loess) occurred about the Bruhnes-Matuyama magnetic reversal, about 800,000 years ago.

Manabe & Terpstra (1974) claim that upheaval of the Plateau essentially created the monsoons of east and south Asia. Once monsoon conditions existed, exclusion of the Indian monsoon made the north side of the Plateau colder and drier and this restricted the growth of glaciers.

Liu & Ding (1998) described the palaeomonsoon as determined from the loess record. In the past 2.6 million years the palaeomonsoon record can be divided into 166 events. The monsoonal rainfall belt has experienced a wide, repeated advance-retreat change during the glacial-interglacial cycles of the Pleistocene. Both temporal and

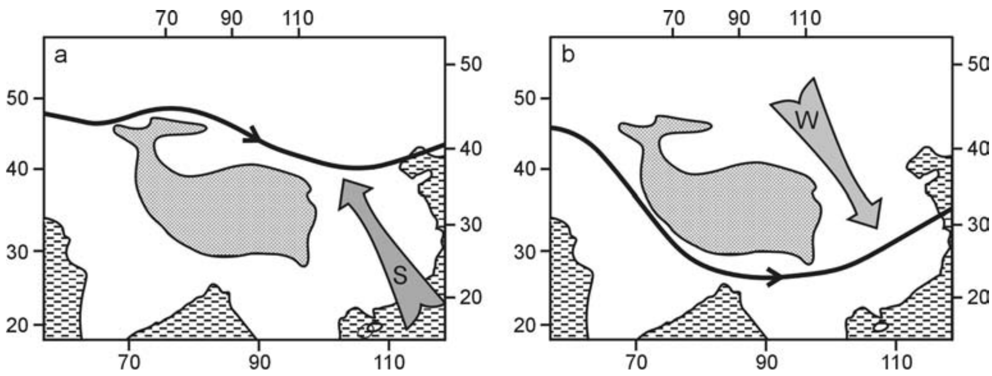

FIG. 3 - Maps of Tibet showing the variations of the jet streams associated with the East Asian Monsoon (after Fang & *alii*, 1999a). a. July. Summer monsoon (northern hemisphere); b. January. Winter monsoon. The shaded area is bounded by the 3000 m contour.

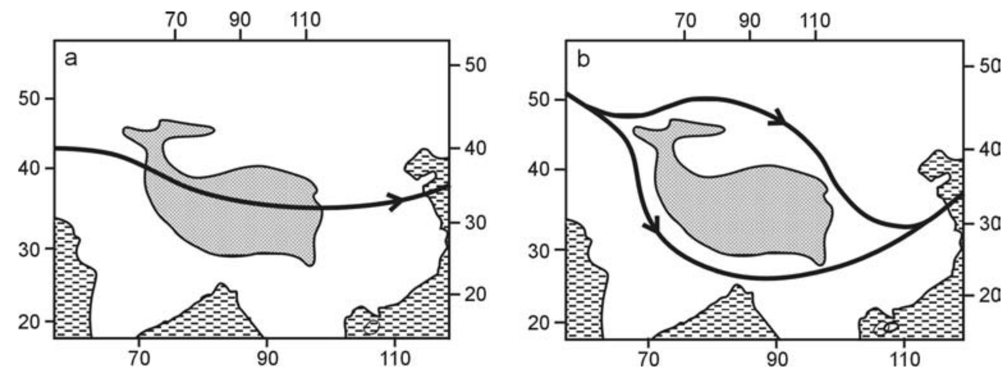

FIG. 4 - The jet stream over and around Tibet, before and after uplift (after Fang & *alii*, 1999b). a. Before 800 000 years ago; b. Since 800 000 years ago. The shaded area is bounded by the 3000 m contour.

spatial changes of the monsoon system in the Quaternary can be linked closely to global ice-volume relationships. It therefore seems that although uplift of the Tibet Plateau may be vitally concerned with the onset of the monsoon, minor variations are controlled by orbital variations.

They also stress the important feature that the monsoon links the low pressure cell over the Tibet Plateau (India Low) to both the Pacific High and the Australia High, leading to inter-hemispheric temperature exchange, and so plays a part in global climate changes. «Because the winter monsoon originates in the high latitudes of the Northern Hemisphere and can transfer climatic signals to low latitudes and even across the equator, we speculate that it may have played a part in the inter-hemispheric connection of climatic changes».

Inter-hemisphere interaction might also work the other way, and some authorities have concluded that the annual pattern of the southern circulation is fundamental to the precise timing and extent of the Asian monsoon.

ANTARCTICA AND CLIMATE

The climatic and tectonic history of Antarctica is so different from the rest of the world that it places important constraints on all global hypotheses of mechanisms controlling climate. Antarctica is unique because the work of running water has been suppressed for millions of years. It also differs from the rest of the world in having a very much longer history of glaciation. To develop a large ice-cap it is necessary to have large precipitation. The climate of the present day probably could not grow an ice cap, and our major ice caps have to be inherited from a previous situation which combined high precipitation with low temperatures in high latitudes. Furthermore, only when sea-ice has grown quite large does it have a significant effect on ocean temperatures, which then works to sustain an ice age.

The cause of glaciation is debatable, as is the time of its initiation. Studies of marine sediments indicate that Cenozoic Antarctic ice sheet activity dates back to 45 Ma, and continental scale glaciation to about 40-36 Ma (Hambrey & *alii*, 1989; Cooper & *alii*, 1991).

Glaciation started about the Eocene-Oligocene boundary, at about 34 Ma according to Deconto & Pollard (2003).

In the seventies Kennet and others proposed that climate cooled and an Antarctic ice sheet developed as the Antarctic Circumpolar Current (ACC) increasingly isolated Antarctica from warm surface circulation of the Southern Hemisphere oceans (see Kennett, 1977). This current originated as Australia and other southern continents drifted north from the stationary Antarctica (fig. 5). The vital breaks were the opening of the Tasman Gateway between Australia and Antarctica in the latest Eocene about 37-33.5 Ma (Exon & *alii*, 2000) and opening of the Drake Passage in the earliest Neogene (Lawver & *alii*, 1992). Deconto & Pollard (2003) propose an alternative hypothesis, that the Antarctic cooling is caused by changes in atmospheric CO_2 This seems unlikely because Antarctica would require a different atmosphere from the rest of the world for 30 million years.

A major debate concerns whether the Transantarctic Mountains experienced uplift since early or middle-Pleistocene (Behrendt & Cooper, 1991), or whether the mountains have remained at their present level since the Miocene (favoured by Kerr & *alii*, 2000). In either case Antarctic glaciation commenced well before major uplift.

It should be noted that only Antartica is exceptional. Glaciation elsewhere in the southern hemisphere, such as New Zealand, South America and Australia coincides roughly with that of the northern hemisphere, apart from some early Oligocene mountain glaciation in Tasmania, which was then close to Antarctica (Macphail & *alii*, 1993).

CAN CLIMATIC CHANGE CAUSE UPLIFT OF MOUNTAIN RANGES?

Molnar & England (1990) pose the most radical question of all: did climate change bring about the uplift of mountains rather than the other way around? They suggest that erosion is a driving force of uplift, and since erosion is climatically controlled to some degree that climate may be a driving force of uplift. «An alternative cause of these phenomena [deep incision and high mountain altitudes] is late Cenozoic global climate change».

The Swiss Alps of today were produced by strong uplift in the Pleistocene, accompanied by severe glaciation. Molnar & England (1990) suggest that although the relief

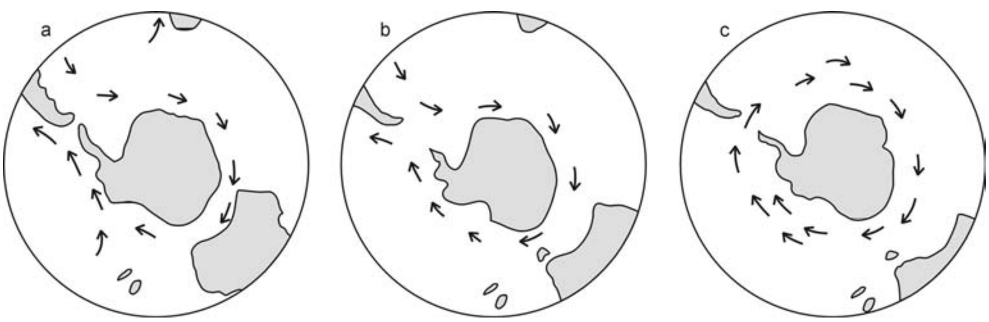

FIG. 5 - Evolution of the Antarctic Circumpolar Current (much simplified after Lawver & *alii*, 1992). a. 50 Ma middle Eocene; b. 30 Ma Early Oligocene; c. 20 Ma Early Miocene.

is 4000 m the mean elevation of the Alps is only 2000 m, and the inferred uplift of 2000 m could simply be the isostatic response to the removal of material by the incision of deep valleys.

Whether an erosional plain graded to sea level is in isostatic equilibrium depends on the prior history. If a considerable overburden has been removed, there could be some isostatic uplift in store. The effect may be enhanced by loading on the bordering seafloor by the deposition of material eroded from the plain area. The uplifted area would of course be subject to further erosion, and the ultimate landscape depends on the balance of different tectonic and erosional processes.

Most mountains and plateaus tend to have very distinct edges, suggesting uplift of distinct blocks, and to raise such blocks by isostasy alone seems improbable. Antecedent rivers draining high plateaus show that although erosion can enhance uplift in some areas (by isostatic compensation of interfluves as well as valleys), it cannot be the prime driving force of plateau uplift, which is the real basis of mountain building. In simple terms, it is very unlikely that the threshold for isostasy can be reached starting with a plain. If somehow a plain came into existence, graded to sea level and in isostatic equilibrium, there is no way that erosion alone could cause uplift. There has to be some force from within the earth to push up the land and induce valley erosion before isostasy can start to have an effect.

Finally, it is clear from Table 1 that mountain uplift in the Neotectonic Period occurred in a wide variety of climates, and it would be difficult to find a common climatic element that could induce uplift in such diverse regions.

CONCLUSIONS

The Neotectonic Period saw many mountain regions arise in the past few million years. This had major effects on climate through several mechanisms, especially orographic cooling and the creation of the monsoon system, but not the negative greenhouse effect. Antarctica has been long isolated by the Antarctic Circumpolar Current and does not share the same tectonic and climatic history as the rest of the world, even though the Transantarctic Mountains may be affected by Neotectonic uplift.

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