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Forcing of Late Cenozoic Northern Hemisphere Climate by Plateau Uplift in Southern Asia and the American West

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Geologic evidence indicates that net vertical uplift occurred on a large (kilometer) scale and at accelerating rates during the middle and late Cenozoic in plateaus of southern Asia and the American west. Based on this evidence, General Circulation Model sensitivity tests were run to isolate the unique effects of plateau uplift on climate. The experiments simulated significant climatic changes in many places, some far from the uplifted regions. The basic direction of most of these simulated responses to progressive uplift is borne out by changes found in the geologic record: winter cooling of North America, northern Europe, northern Asia, and the Arctic Ocean; summer drying of the North American west coast, the Eurasian interior, and the Mediterranean; winter drying of the North American northern plains and the interior of Asia; and changes over the North Atlantic Ocean conducive to increased formation of deep water. The modeled changes result from increased orographic diversion of westerly winds, from cyclonic and anticyclonic surface flow induced by summer heating and winter cooling of the uplifted plateaus, and from the intensification of vertical circulation cells in the atmosphere caused by exchanges of mass between the summer-heated (and winter-cooled) plateaus and the mid-latitude oceans. Disagreements between the geologic record and the model simulations in Alaska and the Southern Rockies and plains may be related mainly to the lack of narrow mountain barriers in the model orography. Taken together, the observed regional trends comprise much of the pattern of "late Cenozoic climatic deterioration" in the northern hemisphere that culminated in the Plio-Pleistocene ice ages. The success of the uplift sensitivity experiment in simulating the correct pattern and sign of most of the observed regional climatic trends points to uplift as an important forcing function of late Cenozoic climatic change in the northern hemisphere at time scales longer than orbital variations; however, the modest amplitude of the uplift-induced cooling simulated at high latitudes indicates a probable need for additional climatic forcing.

1. INTRODUCTION

Kutzbach et al. [this issue] report results from sensitivity experiments using varying orography (Figure 1) to determine the projected impacts of Cenozoic plateau and mountain uplift on northern hemisphere atmospheric circulation. One major result was a redirection of the mean planetary waves in the northern hemisphere troposphere at middle and high latitudes. The extent of this impact can be gauged from the changes in the 500-mbar height field shown in Figure 2; progressive uplift rearranges the flow from nearly zonal in the no-mountain (NM) experiment to the meandering form observed today and shown by the full-mountain (M) experiment. Kutzbach et al. show that the impact of plateau and mountain uplift on this and other aspects of the large-scale circulation is roughly linear across the range from no orography to full plateaus and mountains.

The uplift experiments also simulated a rich variety of changes in lower atmospheric circulation and surface climate at middle and high latitudes of the northern hemisphere. This paper examines these model-simulated changes in surface climate and compares them with available evidence from the geological record. Section 2 discusses the experimental design and compares the model representation of orography to the actual geologic history of uplift. Section 3 summarizes the results of the uplift experiment simulations at a hemispheric scale. Sections 4 and 5 compare the simulated sensitivity of the climate system to uplift against the observed climatic record over land and over several ocean regions. Section 6 evaluates uplift as a forcing function of late Cenozoic climatic change and compares its influence to other possible climate-forcing mechanisms. Section 7 notes limitations of the current model and suggests future experiments to improve the climatic estimates.

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2. EXPERIMENTAL DESIGN

Ruddiman et al. [this issue] present the rationale for the uplift experiments. Rapid uplift on a large scale in southern Asia and the American west during the late Cenozoic has been inferred by many geologists. The scale of inferred uplift is large enough to affect climate in regions both proximal to, and remote from, the uplifted terrain.

On this basis, we ran a series of experiments (Figure 1): a full-mountain (M) experiment representing modern orography (and used as the control case against which other runs were differenced), a "half-mountain" (HM) experiment considered equivalent to the orography in these two key regions during the latest Miocene or early Pliocene (10–3 Ma), and a "no-mountain" (NM) experiment that is not strictly equivalent to any time interval but resembles the very low orography across most of the northern hemisphere during the late Eocene (40 Ma). We did not, however, correct for plate tectonic motion or other geographic changes in these

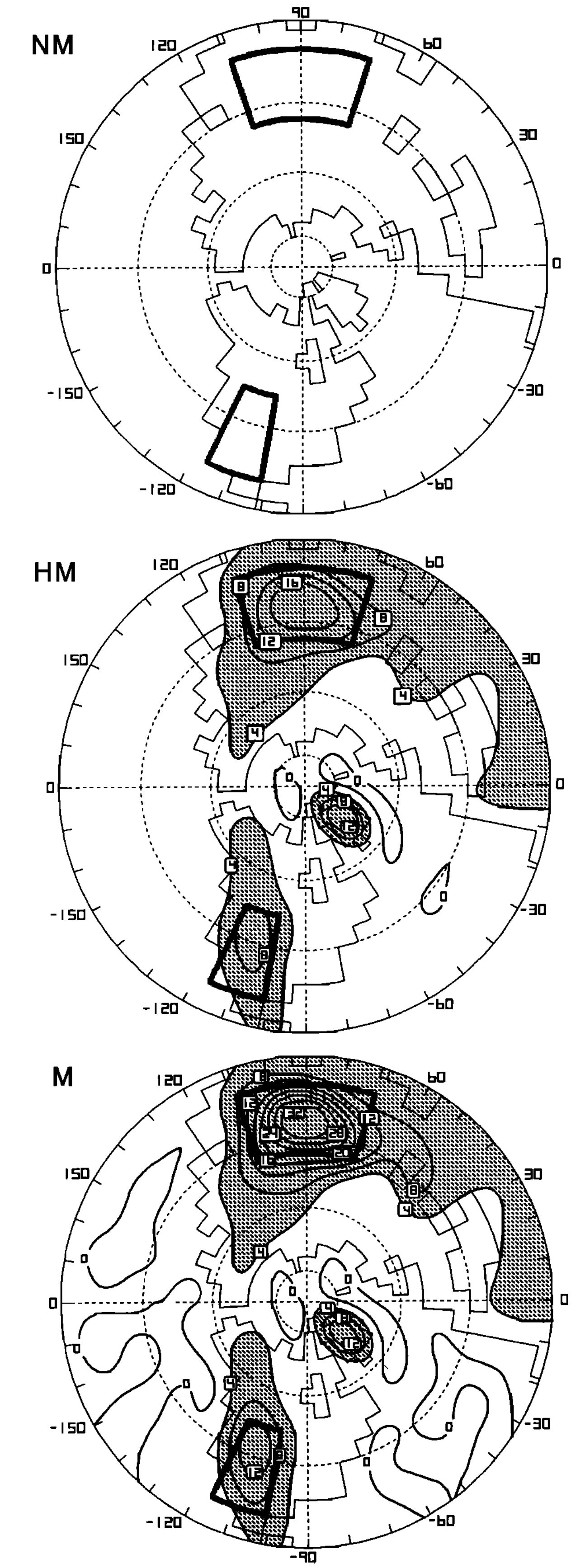

Fig. 1. Orographic heights for no-mountain (NM), half-mountain (HM), and full-mountain (M) experiments, in hundreds of meters. Contour interval is 400 m.

low-orography experiments; these are single-variable (orography) sensitivity tests rather than paleoclimate simulation experiments in which one may prescribe several geologic boundary conditions simultaneously [e.g., *Barron*, 1985]. We did not prescribe changes in snow cover, sea ice, or sea surface temperatures but kept them at their modern values.

The General Circulation Model (GCM) used for these experiments is the Community Climate Model (CCM), described by *Kutzbach et al.* [this issue]. The GCM class of climate models gives the most complete numerical representation available of physical processes acting in the climate system.

One serious model limitation is the smoothing necessitated by the spectral representation of the model's orography (and of climatic variables). Although plateaus and topographic swells broader than 1000 km are portrayed reasonably accurately, narrow mountain ranges are not well resolved. In each region, however, late Cenozoic uplift occurred on broad regional scales, and the model reproduces these broader features with better success.

Cenozoic uplift in southern Asia involved both the Himalayas and the vast Tibetan Plateau. The Himalaya Mountains are smoothed by the model and merged with the southern Tibetan Plateau as a topographic front. However, the Tibetan Plateau, the feature that is critical both in orographic diversion of the planetary waves [*Charney and Eliassen*, 1949; *Bolin*, 1950; *Trenberth*, 1983; *Held*, 1983] and in the development of the Asian monsoon [*Flohn*, 1968; *Hahn and Manabe*, 1975; *Tang and Reiter*, 1984], is well represented in the model.

Late Cenozoic uplift in the American west involves a broad region of high terrain encompassing the Sierra and Rocky mountains, the Colorado Plateau, the Basin and Range, and the High Plains. The prevailing view is that late Cenozoic uplift occurred as a broad epeirogenic swell that elevated this entire area [*Ruddiman et al.*, this issue]. If this view is correct, the model should give a reasonable representation of the effects of uplift in the American west, even though individual mountain ranges are not resolved.

3. Model Results

The uplift experiments indicate that uplift causes a wide array of changes in northern hemisphere atmospheric circulation, with numerous climatic effects at the Earth's surface (Figures 3–9). Here we give a brief overview of first-order changes in circulation that affect climate at the Earth's surface. We follow "time's arrow" by using the no-mountain (NM) experiment as the starting point and tracing the predicted changes due to uplift (from NM to HM to M). This sequence follows the sense of development of the climate system predicted for the Cenozoic.

For surface temperature and precipitation, we express the changes as M-NM (the difference between the no-mountain and full-mountain control case); this comparison also follows the sense of time's arrow by showing the direction of temperature and precipitation changes as a function of increasing uplift. We focus on regions where changes are statistically significant at the 99% level, based upon a *t* test [*Kutzbach et al.*, this issue, section 2]. Our emphasis here is primarily on the pattern and statistical significance of change, because other model limitations and biases preclude accurate estimates of magnitudes of change [*Kutzbach et al.*, this issue, section 3].

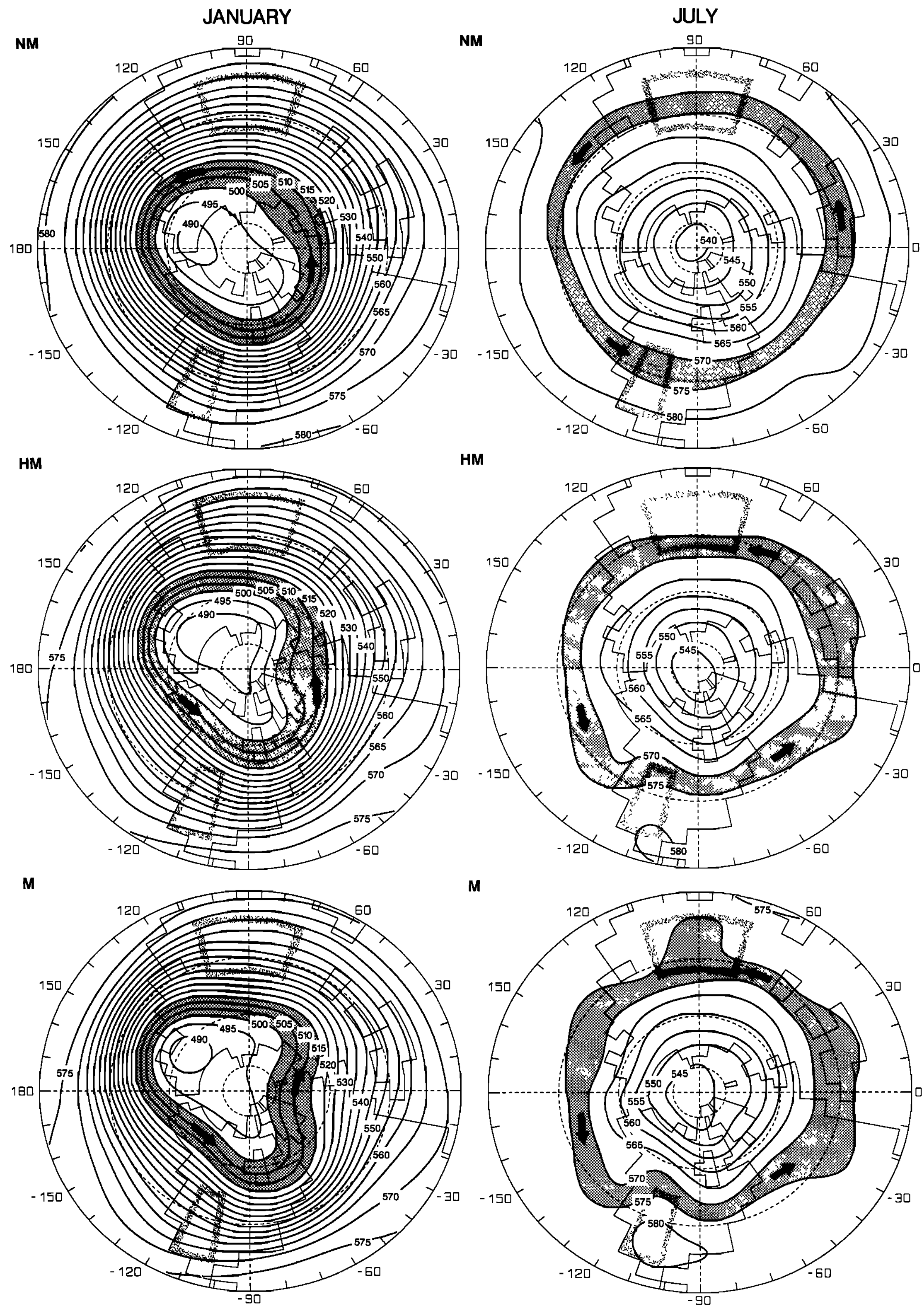

Fig. 2. Geopotential height field at 500 mbar in winter (left) and summer (right), for various stages of plateau and mountain uplift. Stippling follows contours chosen to highlight wavelike meander patterns induced by uplift. Arrows show direction of upper level circulation.

Similar spatial patterns of temperature and precipitation change and statistical significance were found for the two other difference comparisons (M-HM, HM-NM). The magnitude of changes between cases NM, HM, and M is in general proportional to the relative amount of uplift between these experiments [*Kutzbach et al.*, this issue, Table 1].

January

The orographic obstacles posed by the rising terrain in Asia and North America, combined with the progressive changes in winter cooling rates above these features [*Kutzbach et al.*, this issue], cause hemisphere-scale diversion of atmospheric flow and changes in surface climate in winter (Figures 3–5). With uplift, the near-surface (850 mbar) wind flow (Figure 3) acquires a stronger southerly (south-to-north) component west (upstream) of the plateaus along the western U.S. coast and over northern Arabia and a more northerly trend east (downstream) of the plateaus in the American midwest and in eastern Asia. A stronger southerly flow component also develops farther east, over the North Atlantic and over the North Pacific.

Associated with these wind field changes are shifts in intensity and position of the main features of sea level pressure (Figure 3). With uplift, the Icelandic low weakens and shifts to the west, where it partly merges with and reinforces a strengthened trough off the east coast of the United States. In contrast, the Aleutian low strengthens considerably, but it also moves west and merges with a similarly strengthened trough off the east Asian coast. Similar results were obtained from a barotropic model by *Held* [1983]. Uplift also causes a northward migration of the Asian anticyclone (Siberian High) and the development of an anticyclonic ridge in western North America.

With uplift, the January storm tracks (Figure 3) take on a more wavelike shape, with a prominent northward meander over Alaska and a southward displaced track over east central North America and the east coast of Asia. Similarly, the 250 mbar (tropospheric jet stream) wind maxima follow more southerly paths off the east coasts of Asia and North America (Figure 4). In addition, the regions of maximum upper-level winds also shift westward nearer to Asia and build a strong maximum back into the American southwest, with maximum velocities in both jet maxima intensifying by about 10–20%. The shift of maximum upper westerly flow to lower latitudes over the eastern portions of both continents reflects the formation of upper-air low-pressure (cyclonic) systems to the east of the two uplifted plateaus [see *Kutzbach et al.*, this issue, Figure 8]. Similar results were obtained in GCM orography experiments by *Manabe and Terpstra* [1974].

January precipitation (Figure 5) increases across southern and eastern North America beneath the stronger and southward displaced jet and storm track. An analogous increase occurs from the east coast of Asia to the central Pacific beneath the jet maximum. Precipitation decreases over central North America in the rain shadow of the rising mountains and in the region of increased northwesterly wind flow, and it also decreases over Asia and to the north of the Tibetan Plateau. Similar changes in precipitation were observed in the GCM orography experiments of *Manabe and Terpstra* [1974].

January surface temperatures (Figure 5) cool considerably in the uplifted regions, due mainly to lapse rate effects and the changed surface energy budget over the prescribed uplifted terrain [*Kutzbach et al.*, this issue]. Temperatures are also colder in regions remote from the uplifted areas, including most of northern Asia, much of Europe, parts of the Arctic Ocean, and middle latitudes of east central North America. In most cases, these changes are linked to stronger northerly flow associated with the altered planetary wave pattern. In contrast, Alaska, easternmost Siberia, and Greenland become warmer due to the stronger southerly flow.

July

Uplift-induced changes in circulation and climate are also large in July (Figures 6–8). Although the orographic barrier of the rising topography has some direct effect on the circulation, many summer changes are connected with strengthened atmospheric heat sources over the Tibetan and Colorado plateaus [*Kutzbach et al.*, this issue], resulting in increased midtropospheric warmth, upward vertical motion, cyclonic (summer monsoonal) circulation in the lower atmosphere, and increased anticyclonic circulation in the overlying high-pressure systems that develop in the upper troposphere.

Although low-level winds in July are generally weaker than in January, the changes in flow direction caused by uplift are large (Figure 6). On the west coast of North America, stronger northerly flow develops on the eastern flank of the strengthened subtropical high. With formation of a small (monsoon) low over the Colorado Plateau, southerly flow increases along the entire Gulf of Mexico coast. A sharp downwind trough over Hudson Bay brings more northerly flow into the American midwest and a stronger southwesterly return flow along the east coast that merges with the gulf flow.

In Asia the strengthening of the monsoon low over Tibet (Figure 6) rearranges the zonal westerly and trade wind flow into a cyclonic swirl around the Tibetan Plateau, with stronger northeasterlies over Arabia and the Mediterranean and stronger southwesterlies over southeast Asia, Japan, and the Phillipines (W. L. Prell et al., manuscript in preparation, 1989). Similar results in Asia were obtained by *Manabe and Terpstra* [1974]. Over both the North Atlantic and eastern North Pacific oceans, the subtropical highs strengthen and move northward. Due to lower pressure over Asia and higher pressure over the North Atlantic, winds over the Mediterranean shift from westerly to northerly and even northeasterly, similar to changes along the west coast of North America (Figure 6).

In association with the large southward meander in low level winds across east central North America, similar trends develop in the July storm track (Figure 6) and the 250-mbar wind maximum (Figure 7), along with increased upper wind velocities. Over southernmost North America the upper-level winds decrease. These changes (increased westerlies in the north, decreased westerlies or increased easterlies in the south) are associated with the development of an upper anticyclone over the Colorado Plateau [*Kutzbach et al.*, this issue]. In Asia the response to uplift is similar, with stronger upper level westerly flow north of 40°N and weaker westerly flow to the south extending out into the Pacific Ocean. Over India, the Arabian Sea, and

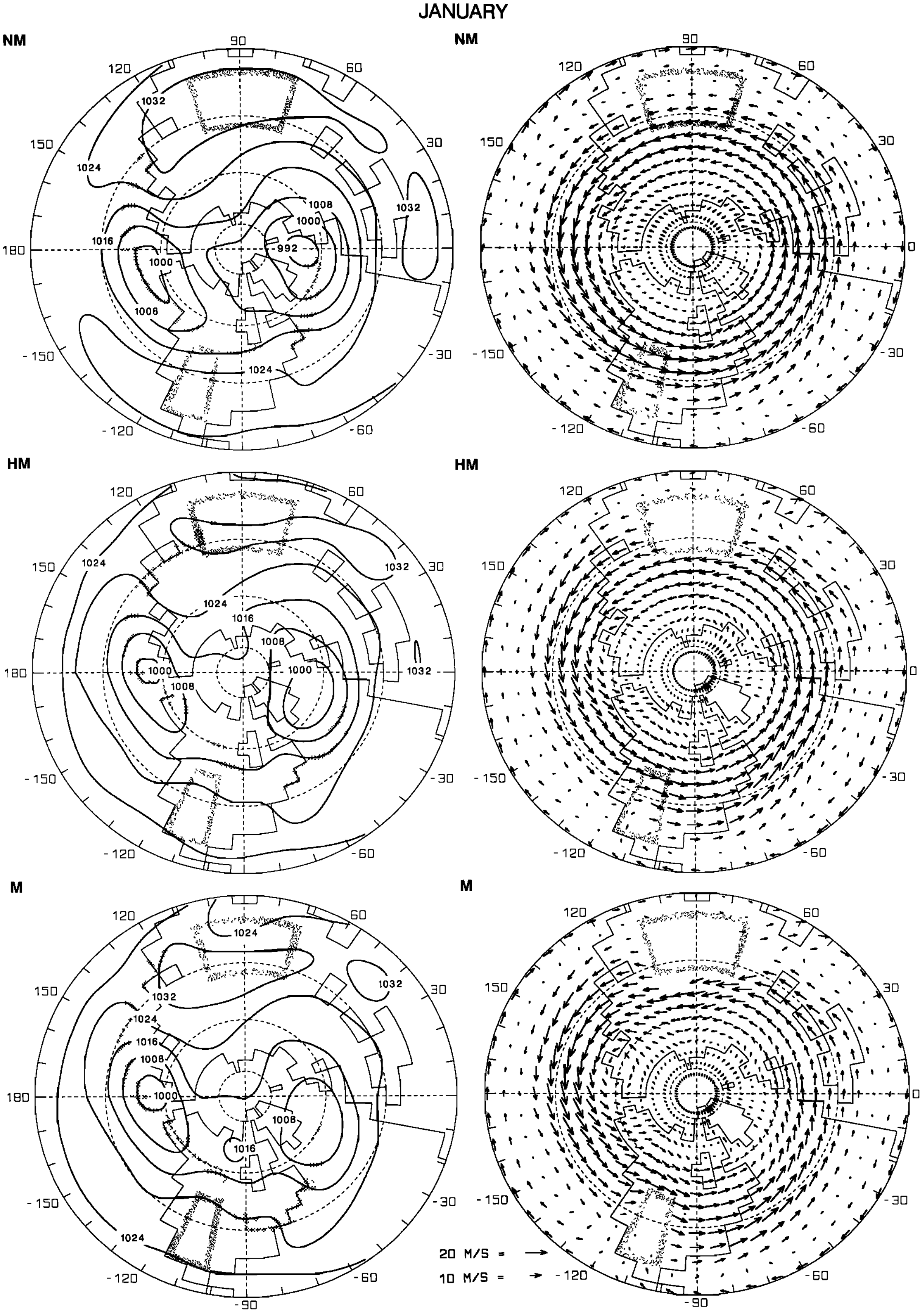

Fig. 3. January sea level pressure (left) and low-level (850 mbar) winds (right) for no-mountain (NM), half-mountain (HM), and mountain (M) experiments. Contour interval for sea level pressure is 8 mbar. Storm track axis (indicated by wide stippled arrow) derived from tracing the maximum values of standard deviation of daily surface pressure after band-pass filtering the daily data to "pass" features with periods of 2–6 days [*Blackmon et al.*, 1983]. Wind vectors at the 850-mbar (1.2 km) level calibrated to standard wind arrows below map. Uplifted areas of southern Asia and the American west are lightly outlined by stippling.

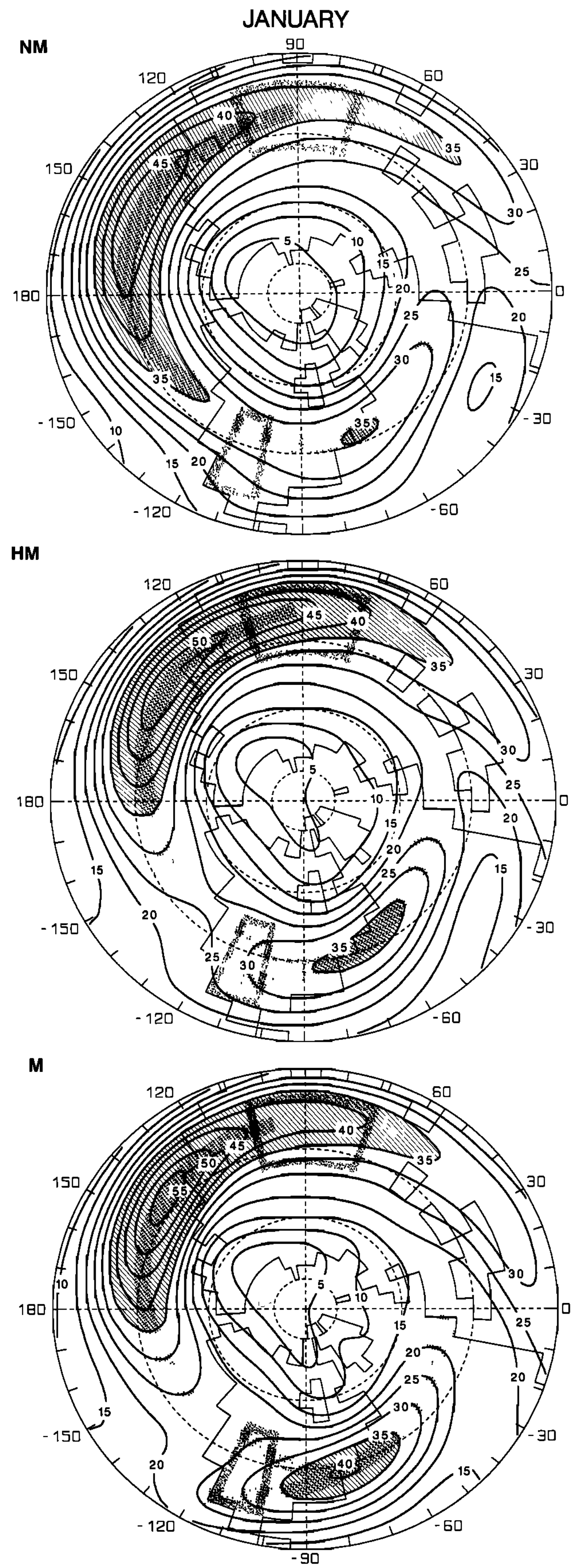

Fig. 4. January wind speed (in meters per second) at the 250-mbar level for no-mountain (NM), half-mountain (HM), and mountain (M) experiments.

North Africa, the easterly jet strengthens (W. L. Prell et al., manuscript in preparation). Similar trends were noted at the 190 mb level in experiments by *Hahn and Manabe* [1975].

July precipitation (Figure 8) increases over southern and eastern coastal Asia due to the stronger monsoon (W. L. Prell et al., manuscript in preparation, 1989). It also increases over south central and southeastern North America, apparently due to stronger southerly flow of moisture-bearing winds from the Gulf of Mexico and the southerly flow to the east of the strengthened low over the plateau. Stronger northerly surface winds west of the Rockies (Figure 6) cause decreased precipitation in that region. Precipitation also decreases over the western interior of Asia, Middle East, and the Mediterranean due to more northeasterly winds.

Surface temperatures in July (Figure 8) cool mainly in the uplift regions due to lapse rate effects and the changed surface energy budget over the prescribed uplifted terrain. Slightly cooler temperatures also extend northeast of the Colorado Plateau (in connection with the strengthened upper-air trough over central North America) and west of the Tibetan Plateau into Arabia and North Africa (due in part to increased northeasterly surface winds).

4. Regional Comparisons: Land

For the following comparisons between geologic data and the uplift sensitivity experiment results, we focus on northern hemisphere continental regions where highly significant climatic changes are simulated over large geographic areas. For each region we specify the uplift-induced climatic changes simulated by the model and compare these simulations to paleoclimatic evidence from the late Cenozoic. These comparisons are summarized in Table 1. We focus only on uplift-induced changes, holding discussion of other climatically important factors until section 6.

These are not conventional data/model comparisons, because other boundary conditions (plate position, sea level, isthmus barriers) have changed during the late Cenozoic interval from which we have compiled geologic evidence of climatic changes. To the extent that major uplift has occurred in the very recent geologic past, however, changes in other geographic factors must have been relatively small [*Ruddiman et al.*, this issue]. In this context, to the extent that our uplift sensitivity experiments approach the status of being climate simulation experiments (i.e., other geologic boundary conditions not differing significantly from the present), our data/model comparisons can be viewed in the conventional sense.

The key indicator for reconstructing climatic change over the continents is fossil vegetation. Early Cenozoic interpretations tend to be based on macrofossil remains such as leaves, seeds, cones, and wood; late Cenozoic interpretations also incorporate pollen deposited in lakes or streams. For comparison with our modeling experiments, we focus only on major, first-order vegetation trends that persist in time over millions of years and in space over large portions of continents. Only the most fundamental vegetational changes (e.g., loss of a major floral type over a broad region) provide the kind of unambiguous climatic signals needed for comparison with the simulations of the climatic effects of uplift. And only changes that persist over millions of years overcome the probable shorter-term responses of Cenozoic

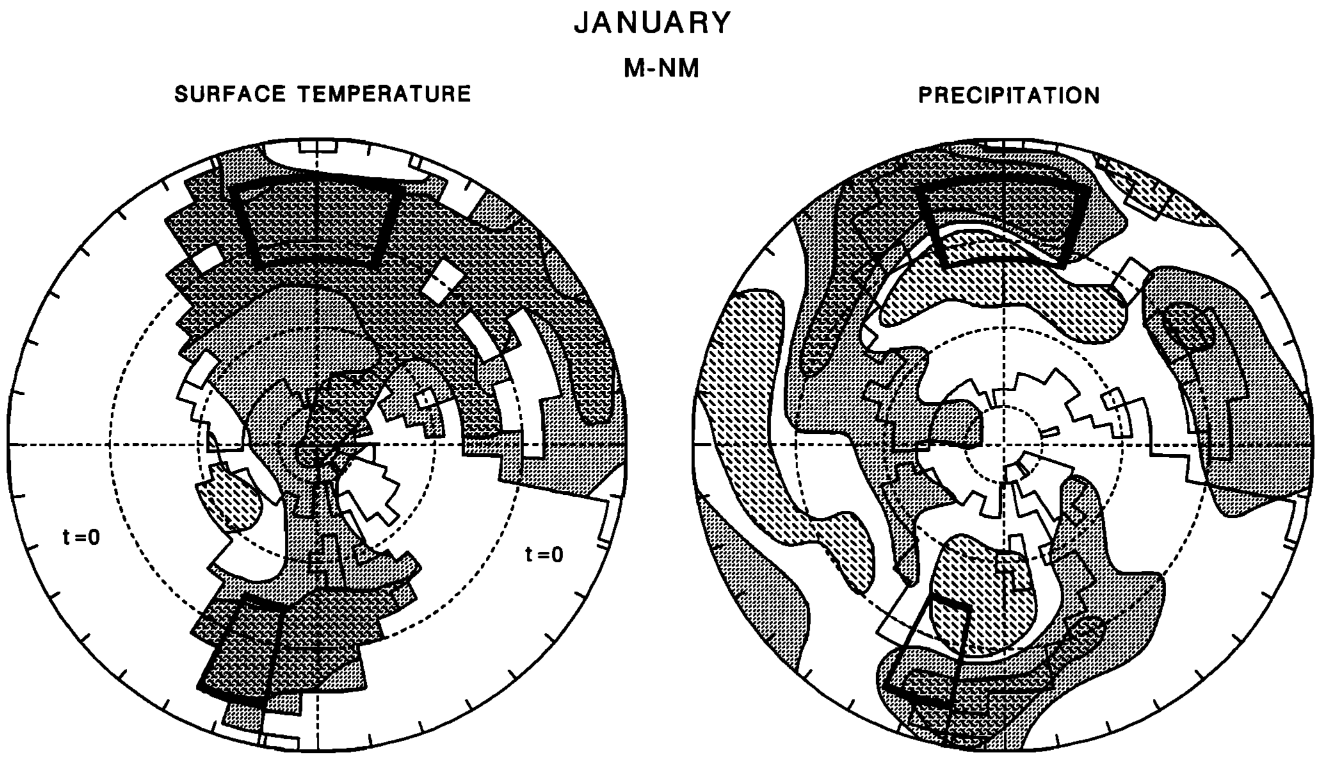

Fig. 5. January changes in surface temperature (left) and precipitation (right) with uplift from no-mountain to mountain experiment (M-NM). Dark shading shows regions that become cooler and wetter with uplift, respectively. Diagonally dashed pattern shows regions where changes are significant at 99% significance level based on the t statistic. Because sea surface temperature is held at modern values, the t statistic is equal to zero over the oceans.

vegetation to orbitally induced variations in climate [*Barnosky*, 1984] or other short-term events such as volcanic eruptions.

Drier Summers Along the North American West Coast (Mid-Latitudes)

The biggest change simulated by the uplift experiments along the American west coast is a transition toward a summer-dry "Mediterranean" climate from the NM to the M case. Drier summers (Figure 8) are caused by a shift of the low-level winds from westerly to northerly along the entire coast from southern California to British Columbia (Figure 6). This change mainly reflects stronger cyclonic flow in the western limb of the deeper low over the Colorado Plateau and increased subsidence and anticyclonic flow in the eastern limb of the North Pacific subtropical high [*Kutzbach et al.*, this issue].

The portion of this large-scale drying which occurs over the California coast agrees with the major trend in late Cenozoic vegetation history along the west coast: development of a Mediterranean type of vegetation able to withstand summer drought [*Axelrod*, 1966*a*; *Ballog and Malloy*, 1981]. During the late Miocene and Pliocene, this vegetation progressively replaced elements of the flora (e.g., Magnolia) that need sustained summer precipitation. This drying trend continued during Plio-Pleistocene interglaciations [*Axelrod*, 1966*a*] but was interrupted by periodically wetter intervals during Plio-Pleistocene glaciations (due to effects of ice sheets on atmospheric circulation).

Drier Winters in the Northern Plains

The uplift experiment simulates drier winter and mean-annual climates in the northern Plains (Figure 5). This results in part from blocking of moist Pacific air masses by the rising topography, but it also reflects the change in wind direction from westerly to northwesterly, bringing in colder, drier air from the north.

The simulated effects of uplift agree with studies over many decades citing the gradual development of drier vegetation types in the northern plains during the late Cenozoic [*Elias*, 1942; *Axelrod*, 1950; *Leopold*, 1967; *Thomasson*, 1979; *Axelrod*, 1985; *Wolfe*, 1985; *Leopold and Denton*, 1987]. Most of these studies ascribe this trend to the lee-side rain shadow formed behind rising mountains in the west; this can be contrasted with comparable latitudes of western Europe, where winter storms from the ocean bring in abundant moisture. The model results add a second important explanation for this drying: the change in winter wind directions provided by large-scale changes in atmospheric circulation around the rising plateau.

Wetter Summers and Winters in the Southern Rockies and Southern Plains

The uplift experiment simulates a broad region of increased precipitation in the southern Rockies and southern plains (Figures 5 and 8). The winter increases are caused mainly by the stronger (southward shifted) jet stream and associated winter storms (Figure 4), and the summer in-

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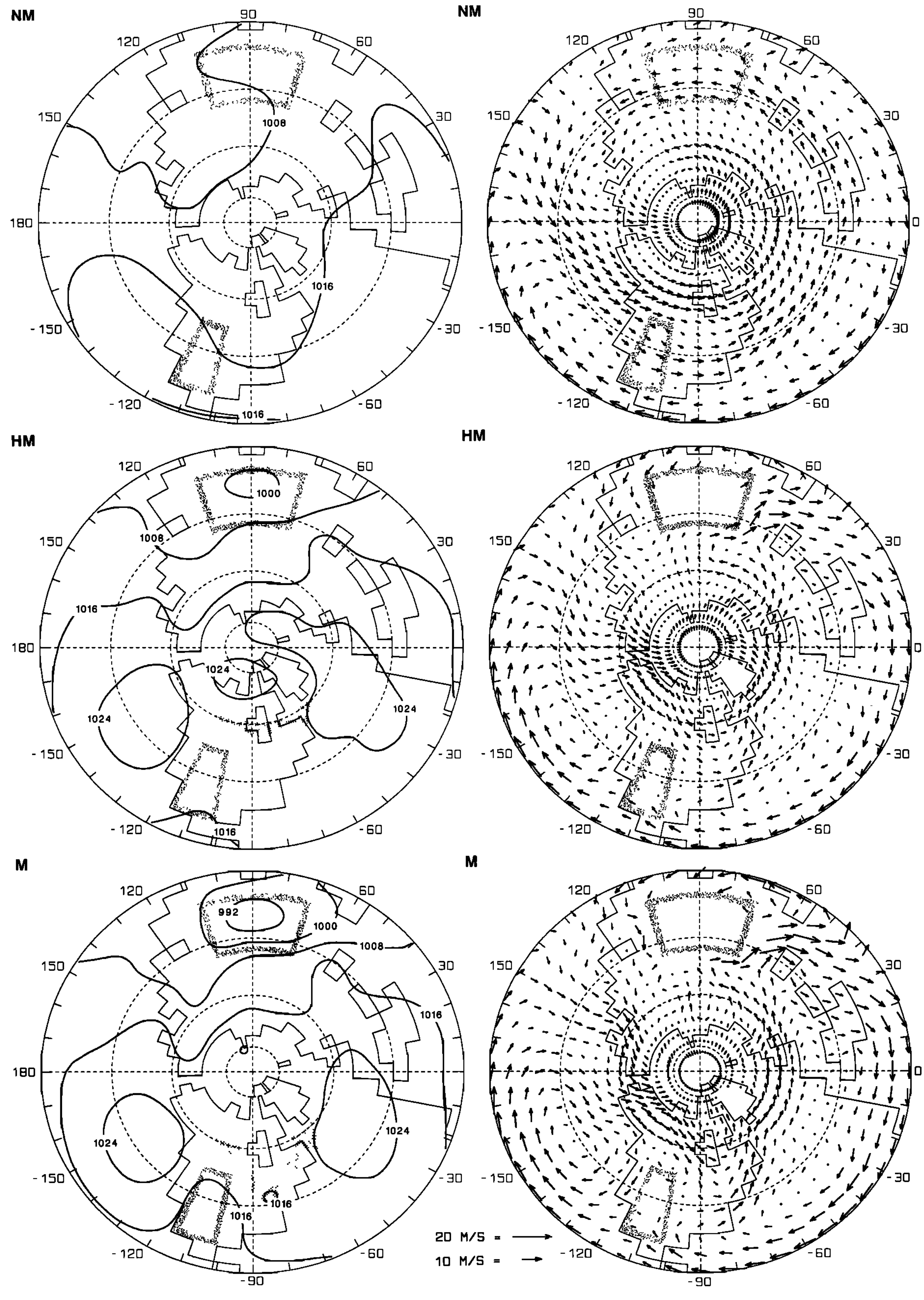

Fig. 6. July sea level pressure (left) and low level (850 mbar) winds (right) for no-mountain (NM), half-mountain (HM), and mountain (M) experiments. Contour interval for sea level pressure is 8 mbar. Storm track axis (indicated by wide stippled arrow) derived from tracing the maximum values of standard deviation of daily surface pressure after band-pass filtering the daily data to "pass" features with periods of 2–6 days [*Blackmon et al.*, 1983]. Wind vectors at the 850-mbar (1.2 km) level calibrated to standard wind arrows below map. Uplifted areas of southern Asia and the American west are lightly outlined by stippling.

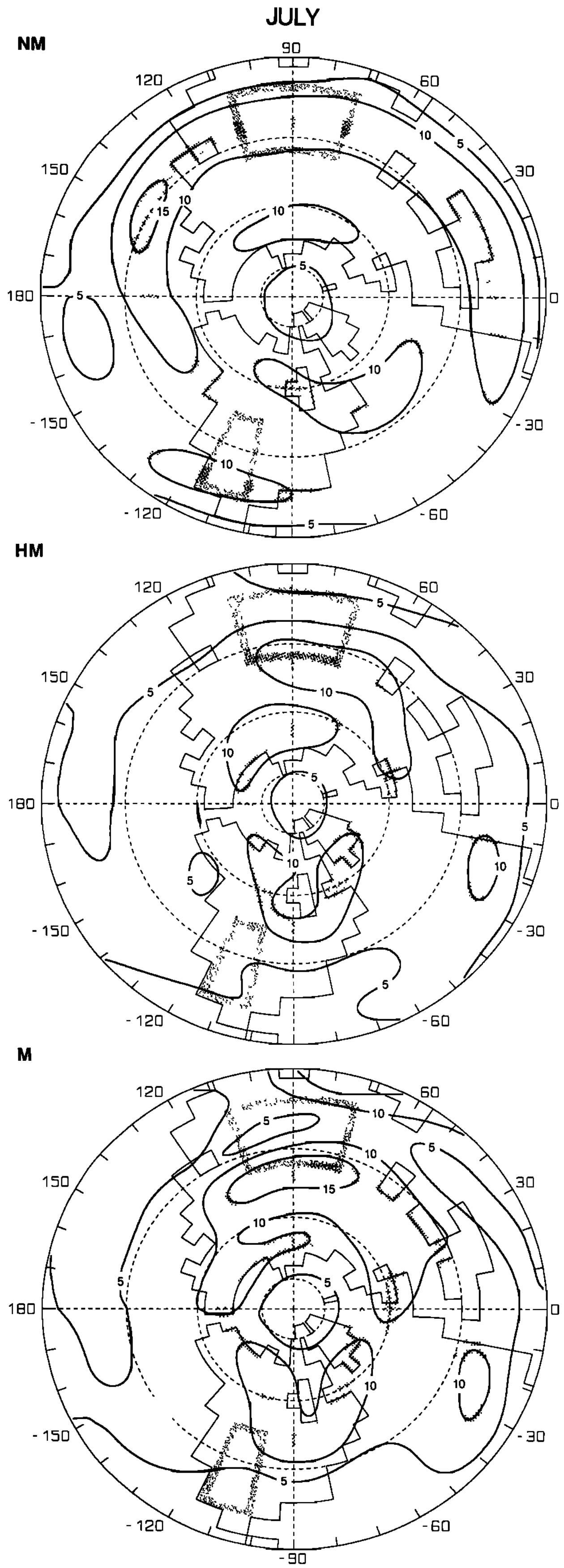

Fig. 7. July wind speed (in meters per second) at the 250-mbar level for no-mountain (NM), half-mountain (HM), and mountain (M) experiments.

creases by increased monsoonal flow from the south (Figure 6).

These simulated effects of uplift disagree at least in part with paleobotanical evidence. Although dry-adapted vegetation occurs in the southern Rockies as early as late Oligocene [*Axelrod*, 1987], and while there is evidence of increased precipitation at higher altitudes of the Rockies in the late Cenozoic [see *Ruddiman et al.*, this issue], the pervasive late Cenozoic trend at lower elevations of the southern Rockies and in the Southern Plains is toward drier conditions [*Axelrod*, 1950, 1966*b*].

One likely reason for this data/model disagreement is smoothing of the high orography of the Rocky Mountains and Sierra in the model. This mutes the effect of winter rainshadow drying on the southern plains and in basins in the Southern Rockies. It also makes it difficult to compare the coarse, regionally integrated output of the model with the spatially complex trend of altitudinally segregated precipitation patterns that actually developed during the late Cenozoic in the Rockies [*Axelrod*, 1988].

The uplift-produced increase in southerly flow in summer that enhances precipitation over the southern plains agrees with the modern monsoonal flow pattern [*Tang and Reiter*, 1984]. In addition, paleobotanical evidence indicates that vegetation requiring summer precipitation has persisted in broad low-lying regions of the Southern Rockies, Arizona, New Mexico, and west Texas during the late Cenozoic, despite being eliminated from the west coast by summer drying [*Axelrod*, 1977]. In a relative sense, this differential trend in moisture regimes to the west and east of the rising plateau agrees with the simulated effects of uplift.

Colder Winters East of the Rockies

Progressive uplift causes a January cooling in excess of 6°C in the area of uplift [*Kutzbach et al.*, this issue, Table 2]. It also causes a significant cooling across a broad region of eastern and central North America (Figure 5), resulting from orographic diversion of lower-level wind flow from westerly to northwesterly (Figures 2 and 3).

This simulation of the effects of uplift agrees with evidence in central and eastern North America for a long-term replacement of broad-leaved evergreen forests by deciduous and boreal forests [*Axelrod*, 1966*b*; *Wolfe*, 1979]. These warm early Tertiary floral types were common in lowland regions of east central North America during the Eocene and Oligocene, still persisted during the middle Miocene in New England [*Fredericksen*, 1984], became scarce by the late Miocene in the middle Atlantic states [*Rachele*, 1976], and vanished from northeastern North America during the Pliocene. A diverse broad-leaved evergreen forest survives today in southern China, where winters are temperate. The progressive late Cenozoic loss of this floristic element from North America has thus been ascribed to increasingly severe winters, with stronger outbreaks of polar air [*Axelrod*, 1966*b*; *Wolfe*, 1979].

Wetter Summers and Winters in Southeastern United States

Most uplift-produced changes in temperature and precipitation in the eastern gulf and southeast coast of North America are relatively small. Precipitation increases signifi-

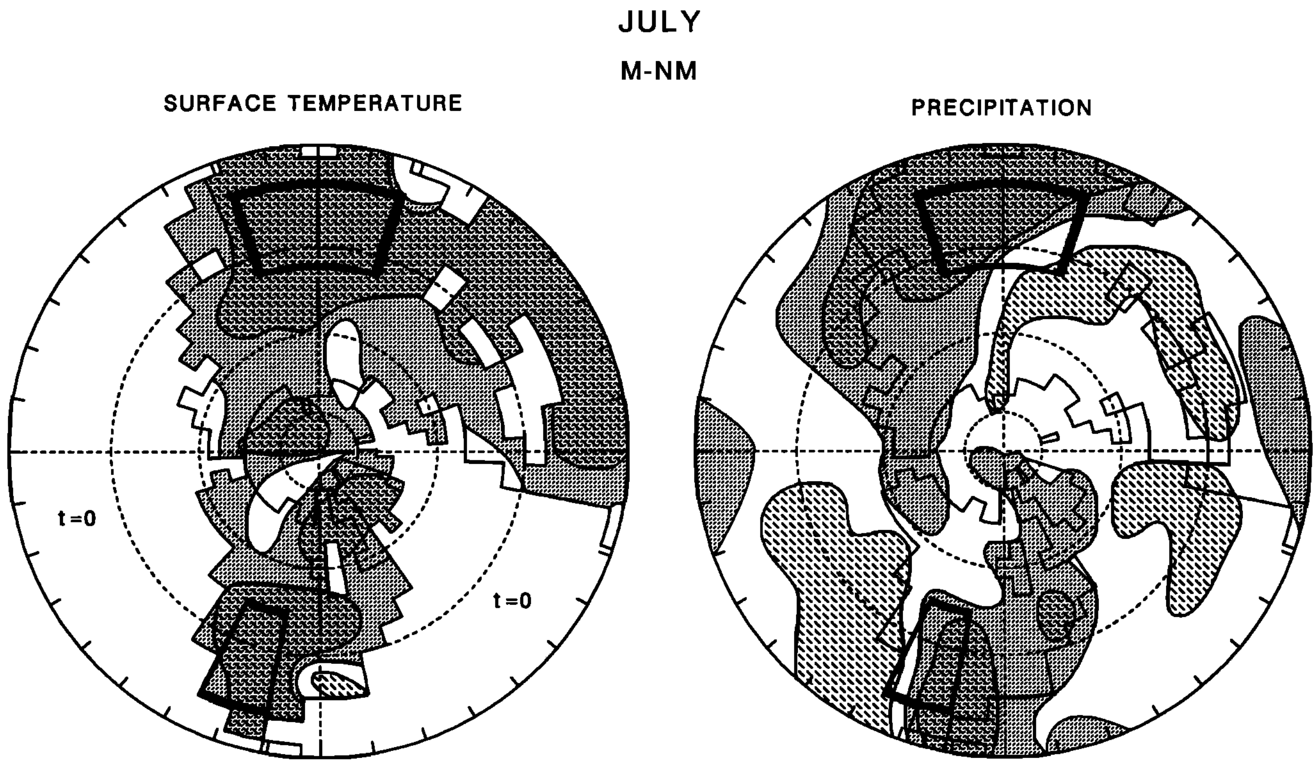

Fig. 8. July changes in surface temperature (left) and precipitation (right) with uplift from no-mountain to mountain experiment (M-NM). Dark shading shows regions that become cooler and wetter with uplift, respectively. Diagonally dashed pattern shows region where changes are significant at 99% significance level based on the *t* statistic. Because sea surface temperature is held at modern values, the *t* statistic is equal to zero over the oceans.

cantly in January (Figure 5) due to deepening of the east coast trough in response to uplift (Figure 3). The small precipitation increases in July are not significant nor is the small warming in summer (Figure 8) or cooling in winter (Figure 5).

Paleobotanical evidence from the margins of the gulf embayment indicates relatively little change in vegetation during the late Cenozoic. Some elements that require a warm, wet climate persisted in this region while being eliminated from the rest of North America [*Graham*, 1964; *Axelrod*, 1966*b*]. The lack of major late Cenozoic changes in vegetation across this region is consistent with the relatively small changes in the model simulations and contrasts with changes toward cooler and drier climates in many mid-latitude regions.

Cooler Summers and Warmer Winters in Alaska

The uplift sensitivity test simulates slightly cooler summers but significantly warmer winters across Alaska (Figures 5 and 8). The warmer winters are due to stronger southerly wind flow at low levels in the eastern limb of an intensified Aleutian low-pressure system (Figures 5 and 3).

Both the lack of significant summer cooling and the strong winter warming disagree with floral evidence of major Cenozoic cooling in Alaska. Broad-leaved deciduous or even subtropical forests were widespread during the Eocene but later gave way to colder deciduous and boreal forest and tundra [*Wolfe*, 1985].

There are several possible reasons for this data/model mismatch. First, the HM and M model orography lacks high mountain barriers that rose during and after the Miocene along the southern Alaskan coast and in the Brooks Range to the north [*Miller*, 1959]. Instead, the topography is smoothed into a low dome in the Alaskan interior (Figure 1). As a result, the northward flow toward the southern Alaskan coast in winter, comparable to that observed today, is not barred by coastal mountain barriers but penetrates into and across all of Alaska. More realistic simulation of this rising barrier along southern Alaska should have reduced this flow, along with the anomalous low-level warming. More accurate simulation of the rising orography might also have caused orographic trapping of snow sufficient to explain the onset of glaciation in the Elias and Wrangel ranges during the late Miocene noted by *Denton and Armstrong* [1969].

Second, the model's lack of interactive sea ice, snow cover, and ocean temperatures makes data/model comparisons difficult in both seasons in Alaska, particularly along the maritime-influenced North Slope. Finally, other climatic forcing may well be needed to explain the large Cenozoic cooling observed in Alaska.

Colder Winters in Northern Europe

In parts of Europe, uplift causes a statistically significant winter cooling of several degrees Celsius (Figure 5). This occurs in association with the orographically induced west-southwestward shift and weakening of the winter Icelandic low (Figure 3) and increased northerly flow aloft (Figure 2 and Figure 8 of *Kutzbach et al.* [this issue]), which brings more continental climatic conditions to Europe.

TABLE 1. Comparison of Uplift Sensitivity Test Results With Geologic Observations of Climate Change on Land

Region	Climate Change Simulated by Model	Data Versus Model*	Observed Climate Change	Evidence for Observed Climate Change	Age of Observed Climate Change	Source of Evidence
West Coast of North America	drier summers	√	drier summers	loss of flora requiring wet summers	late Miocene to Pliocene	*Axelrod* [1966*a*]
American northern plains	drier winters (colder)	√	drier winters (colder)	change from forest to grasslands	late Miocene to Pliocene	*Thomasson* [1979] and *Axelrod* [1985]
America, Southern Rockies and southern plains	wetter summers and winters	X	drier at low elevation; wetter (?) at high elevation	shift from grasslands to desert scrub at low elevation (persistence of some summer-wet vegetation)	Eocene to Pliocene	*Axelrod* [1950, 1966*b*, 1988]
Eastern northern America	colder winters and summers	√	colder winters (and summers?)	loss of flora sensitive to extreme winter cold	middle Oligocene to Pliocene	*Axelrod* [1966*b*] and *Wolfe* [1979]
SE United States and Gulf Coast	wetter, little temperature change	√	little temperature change	persistence of flora eliminated from rest of eastern North America (i.e., little change)	Eocene to Pliocene	*Graham* [1964] and *Axelrod* [1966*b*]
Alaska	warmer winters (cooler summers)	X	colder, drier	loss of flora sensitive to cold; mountain glaciation	Eocene to Pliocene	*Wolfe* [1985]
Northern Europe	colder winters (and summers)	√	cooler	elimination of subtropical flora	Miocene to Pliocene	*Dorf* [1955] and *Van der Hammen et al.* [1971]
Mediterranean	drier summers (wetter winters)	√	drier summers	first "Mediterranean" flora	late Pliocene	*Suc* [1984]
			increased salinity	marine $\delta^{18}O$ and ostracod fauna	early Pliocene to Pleistocene	*Thunell et al.* [1977]
Northern Asia	colder winters (and summers)	√	colder	change from subtropical forest to Taiga	Eocene to Pliocene	*Tanai* [1972] and *Wolfe* [1985]
Eurasian interior	drier summers and winters	√	drier	change from forest to steppe	late Miocene to Pliocene	*Traverse* [1982] and *Wolfe* [1985]
				increased dust flux to Pacific Ocean	Oligocene to Pliocene	*Leinen and Heath* [1981] and *Janecek* [1985]
Coastal SE Asia	wetter, little temperature change	√	little change	persistence of flora eliminated from rest of Eurasia (little change)	Eocene to Pliocene	*Axelrod* [1966*b*] and *Li* [1985]

*Check indicates qualitative agreement of uplift experiment with geologic observation; cross indicates qualitative disagreement of uplift experiment with geologic observation.

This simulation of the effects of uplift agrees with observations showing a long-term trend from subtropical vegetation in the Eocene to a warm temperate Miocene flora and to cool temperate vegetation by the Pliocene [*Dorf*, 1955], although this sequence is not unidirectional [*Wolfe*, 1985]. *Van der Hammen et al.* [1971] noted that the last subtropical taxa disappeared by the late Miocene, apparently in association with climatic cooling. *Suc and Zagwijn* [1983] summarized pollen evidence that climate in Europe further changed from relatively warm and moist during the early Pliocene (5.2–3.2 Ma) to an interval of alternating cool and warm fluctuations during the late Pliocene (3.2–2.3 Ma), just prior to the start of large-scale northern hemisphere glaciation.

Drier Summers in the Mediterranean

The uplift experiment simulates summer drying in the Mediterranean (Figure 8) due to a change in lower-level winds from westerly to northeasterly in the western limb of an intensified low-level cyclonic flow around the uplifted Tibetan Plateau and in the eastern limb of the intensified North Atlantic subtropical high (Figure 6). Increased subsidence to the west of the uplifted plateau is also a factor [*Kutzbach et al.*, this issue, Figure 3]. Winters become slightly wetter (Figure 5), but the largest and most significant precipitation change is the summer drying.

Tectonic isolation of the Mediterranean basin during the late Miocene [*Ryan et al.*, 1973] complicates interpretations of climate in this region. We focus mainly on climatic changes subsequent to the reconnection between the Atlantic and Mediterranean at about 5 Ma.

Consistent with the simulated uplift effects, *Suc* [1984] noted that the classical "Mediterranean" vegetation, which depends on dry summers and cool wet winters, first developed at about 3.2 Ma, and that summer drought became more persistent after 2.8 Ma. These trends slightly precede development of large-scale northern hemisphere ice sheets, which periodically brought steppe vegetation to the shores of the Mediterranean at and after 2.3 Ma [*Suc*, 1984].

Also consistent with the uplift simulations, *Thunell et al.* [1987] summarized geochemical and biotic evidence that

TABLE 2. Comparison of Uplift Sensitivity Test Results With Geologic Observations of Climate Change Over the Ocean

Region	Climate Change Simulated by Model	Data Versus Model*	Observed Climate Change	Evidence for Observed Climate Change	Age of Observed Climate Change	Source of Evidence
Arctic Ocean	colder winters and summers	(√)	colder	loss of subtropical/ deciduous forest along margins	Oligocene to Pliocene	*Sher et al.* [1979] and *Wolfe* [1985]
			more sea ice	decreased marine productivity		*Herman and Hopkins* [1980] and *Clark* [1982]
Mid-latitude North Atlantic	increased southerly wind flow	(√)	increased northern source deep water	increasing Atlantic/Pacific $\delta^{13}C$ differences	late Miocene to Pliocene	*Woodruff and Savin* [1989]
Mid-high-latitude continents	colder (air advected over ocean)					
subtropical North Atlantic	increased subsidence; decreased P/E					
East North Pacific	more northerly winds in summer	(√)	increased upwelling	increased deposition of diatom silica	middle Miocene to Pliocene	*Ingle* [1973]

*Check indicates qualitative agreement of uplift experiment with geologic observation (parentheses indicate agreement with oceanic changes inferred from atmospheric changes simulated by model).

circulation in the Mediterranean changed from a pattern of deep-inflow/surface-outflow to the modern pattern of surface-inflow/deep-outflow sometime after 2 Ma, culminating a trend that had begun well before the onset of northern hemisphere glaciation. This change requires a shift toward a more negative local precipitation/evaporation (P/E) balance and hence a drier climate.

Colder Winters and Cooler Summers in Northern and Central Asia

Uplift causes a statistically significant winter cooling of several degrees Celsius over the interior and east coastal regions of Asia north of 40°N (Figure 5). Away from the immediate vicinity of the uplifted Tibetan Plateau, where the winter cooling reached 10°C [*Kutzbach et al.*, this issue, Table 2], the winter cooling reflects an increased continentality of climate associated with the strengthening and northward shift of the Siberian High (Figure 3). Summers also become significantly cooler by a small amount (Figure 8).

This simulated effect of uplift agrees with the vegetation histories of northern Asia and Japan. *Wolfe* [1985] noted a change from broad-leaved deciduous forest to conifers and then to Taiga in Siberia from the late Miocene to the present. *Tanai* [1972] summarized the Cenozoic progression in northern Japan from subtropical to warm temperate to cool temperate vegetation.

Drier Winters and Summers in the Asian Interior

In the uplift sensitivity experiments a broad region of interior Asia has reduced precipitation in winter (Figure 5), and the region north and west of Tibet also has reduced precipitation in summer (Figure 8). These changes reflect several factors: "rain-shadow" effects of the higher Himalayan/Tibetan complex to the south; development of cyclonic wind flow around the north and west sides of the raised plateau brought about by the strengthened summer monsoon; and increased midlevel subsidence in summer to the north and west of the plateau [*Kutzbach et al.*, this issue, Figure 3].

These simulations agree with evidence that deciduous forest was progressively replaced by steppe in the late Tertiary both in northwest China [*Li*, 1985] and near the Black Sea [*Traverse*, 1982]. In addition, deserts expanded in the Asian interior, and this has been widely interpreted as resulting from Tibetan uplift [*Zhao and Xing*, 1984; *Whyte*, 1984; *Li*, 1985; *Liu*, 1985]. Cores in the central and western North Pacific contain increased influxes of eolian dust during the Cenozoic, with minimum influxes around 40 Ma and progressive increases subsequently [*Leinen and Heath*, 1981; *Janecek*, 1985]. This implies drying of Asian dust source areas north of 40°, consistent with the uplift simulation (Figures 5 and 8).

Wetter Summers and Winters in Southeast Asia

Uplift causes wetter conditions for both January and July in southeast Asia and along the east Asian coast south of 40°N. The small cooling simulated in January is statistically significant in some regions but small in magnitude; the very small cooling along the coast in July is not significant. The wetter conditions are caused by the stronger monsoonal flow from the south in summer and by the orographically induced deepening of the winter low-pressure system off the east coast of Asia in winter.

The relative lack of late Cenozoic vegetation change in this region indicates relative climatic stability [*Li*, 1985]. In fact, this is the region cited as the "refugium" where elements similar to the warm high-latitude Eocene flora persisted during the deterioration of global climate [*Axelrod*, 1966*b*]. Red lateritic soils that formerly occupied extensive portions of Asia to the north and west are also now confined to this region [*Zhao and Xing*, 1984]. The persistence of tropical/subtropical conditions in this region thus is in agreement with the uplift sensitivity estimate for continued warm and wet climate. The fact that both southeastern coastal Asia and the southeastern United States acted as "refugia" for warm, wet vegetation types during the late Cenozoic is thus consistent with the climates simulated by the uplift experiment.

5. REGIONAL COMPARISONS: OCEANS

The lack of an interactive ocean and sea ice in the model precludes the ocean surface from coming into adjustment with, and reacting back on, the changes in atmospheric conditions produced by uplift. For this reason, detailed comparisons between these uplift sensitivity test experiments and marine geologic observations would not be meaningful. Nevertheless, several of the atmospheric responses to uplift are sufficiently large in scale to warrant discussion of the possible implications for changes in ocean circulation, as well as comparison with marine geologic data (Table 2). Again, we focus these comparisons only on the effects of uplift and leave discussion of the importance of other factors until section 6.

Colder Winters and Cooler Summers in the Arctic

Uplift induces a winter cooling of several degrees Celsius over parts of the Arctic Ocean now covered by ice in winter (Figure 5). (Note that surface temperature changes are calculated over sea ice in this model but the sea ice is not interactive.) The only Arctic regions that did not cool in winter (and in fact warmed) were off the north coast of Alaska, as discussed previously, and near Greenland, in both cases due to southerly advection. Summers were also slightly cooler over most of the Arctic.

These model results agree in sign with geologic evidence for cooling of the Arctic during the Cenozoic. The earlier presence of broadleaf forests in the Eocene and Oligocene [*Wolfe*, 1985] indicates warm, moist climates incompatible with the existence of sea ice, at least along the Arctic coast. The late Pliocene onset of permafrost in Siberia [*Sher et al.*, 1979] and replacement of northern deciduous forest by tundra along the Arctic margins [*Repenning et al.*, 1987] have been attributed to a major cooling and increase in sea ice cover, although the existence of less extensive sea ice at an earlier time is not precluded. With little data available from the early middle Cenozoic, marine geologists agree only that cooling caused extensive sea ice to form prior to the late Pleistocene [*Herman and Hopkins*, 1980; *Clark*, 1982].

Although agreeing in direction, the uplift sensitivity experiments produce only a small Cenozoic cooling (several degrees Celsius) in the Arctic. Of course, our experiments may have constrained the climatic effects of uplift in the Arctic by the prescription of modern sea ice limits; models with interactive sea ice are thus needed to test more completely the effects of uplift. Moreover, other climatic forcing may well be needed to explain the large high-latitude cooling (section 6).

Changes in Circulation Over the North Atlantic

Despite the lack of an interactive ocean, several uplift-induced changes in the model atmospheric circulation point to likely changes in North Atlantic surface circulation conducive to enhanced formation of northern source deep water.

First, over the middle- and high-latitude Atlantic, surface winds changed in such a way as to increase advection of saline subtropical water poleward. In summer, the northern limb of the strengthened subtropical high moved northward from 40°N to 60°N and changed from a zonal westerly flow over the central Atlantic to a more meridional southwesterly flow along the North American coast (Figure 6), in response to the stronger summer east coast trough induced by orographic changes. Winter winds also changed from westerly to west-southwesterly across much of the North Atlantic (Figure 3).

Second, the winter cooling in eastern North America and in northwestern Europe (Fig. 5) juxtaposes increasingly cold air over land (and presumably advected out over the ocean) against salty subtropical Atlantic water being driven farther poleward by the winds. Because this configuration results in winter formation of North Atlantic Deep Water and Labrador Sea Intermediate Water today [*Worthington*, 1970], we suggest that uplift aided these processes during the late Cenozoic.

Third, changes occurred at subtropical latitudes in a direction favorable to increased salinity of North Atlantic subtropical waters. Stronger rising motion in summer over the uplifted plateaus was balanced by increased subsidence over the North Atlantic subtropical gyre (see Figure 3 of *Kutzbach et al.* [this issue]; also note the rise in sea level pressure in Figure 6 of this paper). Increased subsidence led to decreased precipitation in summer (Figure 8) and to more negative summer and estimated annual P/E balances (not shown). This should cause increased salinity in the subtropical gyre. Similarly, increased summer drying of the Mediterranean (Figure 8; section 4) presumably raised the surface salinity and increased the outflow of saline Mediterranean waters at intermediate depths [*Thunell et al.*, 1977]. Finally, independent of our model, uplift and emergence of the Panamanian Isthmus in the Pliocene [*Keigwin*, 1982] may also have altered the salinity balance of the low-latitude North Atlantic Ocean by eliminating the oceanic export of salt and altering the atmospheric advection of water vapor to the Pacific.

This model-based inference of the effects of uplift on deep-water formation is broadly consistent with stable isotopic evidence [*Woodruff and Savin*, 1989]. Northern source deep-water formation in the Atlantic was basically nonexistent in the Eocene and Oligocene, may have occurred sporadically during the early and middle Miocene [*Miller and Fairbanks*, 1985], and then intensified in the late Miocene [*Woodruff and Savin*, 1989]. In addition, the mid-Miocene switch of biosiliceous sedimentation from the North Atlantic to North Pacific is also consistent with the initiation of significant deep-water formation in the North Atlantic [*Woodruff and Savin*, 1989].

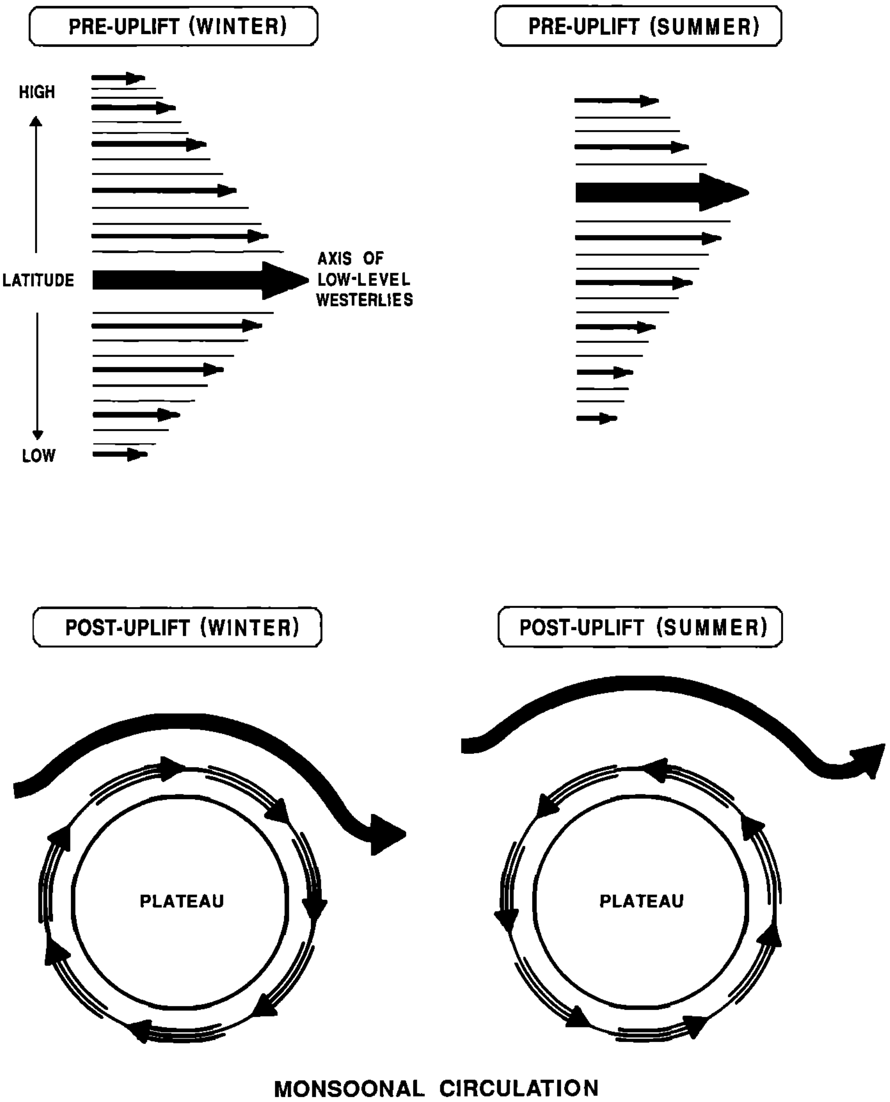

Fig. 9. Schematic map view summary of changes in wind directions due to late Cenozoic uplift of plateau/mountain regions in southern Asia and the American west. Thick arrows show changes in zonal westerly winds due to diversion around orographic barrier. Thin arrows in lower panel show thermal (monsoonal) effects on the circulation: low-level cyclonic flow in summer around low-pressure heat source over plateau; low-level anticyclonic flow in winter around high-pressure heat sink over plateau.

Held [1983] previously noted that the meridional orientation of the zero curl of the wind stress over the modern North Atlantic is linked to orographic control by the Rockies. *Warren* [1983] viewed this meridional wind field, which strongly influences both temperature and salinity over the subpolar ocean, as a major reason why deep water forms today in the North Atlantic but not in the North Pacific, where winds are more zonal.

Northeast Pacific

The uplift-induced shift in summer wind direction from westerly to northerly along the Pacific coast of North America should enhance coastal upwelling of cool waters due to increased Ekman drift of water toward the southwest and west. (Note also that increased upwelling of cool waters along the American west coast should in turn strengthen the low-level temperature inversion and reduce rainfall, thus further enhancing the uplift-induced trend toward summer aridity in California.)

Several studies have detected long-term Cenozoic surface water cooling on the Pacific coast [*Durham*, 1950; *Addicot*, 1968]. In addition, deposition of diatom-bearing sediments attributed to upwelling began in the middle Miocene and has continued to the present, although diluted in Plio-Pleistocene times by a much greater influx of terrigenous clastics attributed to coastal tectonism [*Ingle*, 1973].

6. DISCUSSION: FORCING OF LATE CENOZOIC CLIMATE

There is wide agreement that the late Cenozoic has been a time of progressive climatic deterioration, culminating in the northern hemisphere glaciations of the late Pliocene and the Pleistocene. The term "late Cenozoic climatic deterioration" usually refers to the general climatic cooling at middle and high latitudes but may also encompass increased drying and continentality (seasonal extremes of temperature and precipitation).

Our comparison of the results of the uplift sensitivity experiments with geologic data indicates that progressive uplift of plateaus and mountains in southern Asia and the American west provides a physically and geographically coherent explanation for some of the fundamental patterns of regional climatic change in the northern hemisphere during the last 10 or 15 m.y. (Tables 1 and 2). This includes circulation changes leading to colder winters (and to a lesser extent summers) over North America, northwest Europe, and northern Asia; drier summers along the American Pacific coast, in the interior of Eurasia, and in the Mediterranean; winter drying of the American northern plains and the interior of Asia; and maintenance of warm/wet conditions along the southeast coasts of Asia and the United States. The patterns involving greater east-west differentiation toward wetter or drier climates are particularly suggestive of the effects of uplift.

In addition, the uplift experiment produces several changes in atmospheric circulation over and around the North Atlantic that are, by inference, clearly favorable to increased formation of deep and intermediate waters. It also simulates a small cooling over most of the Arctic Ocean, as well as changes in the eastern North Pacific favorable to increased upwelling of cool water in summer.

The direction of climatic change simulated by the uplift sensitivity experiments appears to be incorrect in Alaska and in parts of the southern plains and Rockies. In both regions the lack of realistic treatment of narrow mountain ranges and consequent barrier effects to circulation appears to account at least in part for the discrepancy.

Collectively, the individual uplift-induced climatic changes in Table 1 encompass most of the northern hemisphere continental trends usually cited as evidence of "late Cenozoic climatic deterioration." The success of the progressive uplift sensitivity experiments in explaining the direction of these trends, including especially the increasing east-west differentiation of the hydrological responses, points to uplift as a major factor in the pattern of late Cenozoic climatic change at Northern Hemisphere middle latitudes. We do not claim that these uplift sensitivity experiments account for the full amplitude of the observed changes; we emphasize here the model's success in capturing the basic direction (and thus pattern) of climatic change.

In compiling the observed trends for Tables 1 and 2, we ignored climatic oscillations spanning intervals of a few million years [e.g., *Shackleton and Kennett*, 1975; *Savin et al.*, 1975; *Wolfe*, 1979; *Miller et al.*, 1987]. These fluctuations may reflect complex couplings and feedbacks in the climatic system or may indicate the effects of climatic forcing other than uplift. We also do not explicitly address the origin of steplike changes observed in many Cenozoic records [*Berger et al.*, 1981]; these may involve threshold responses to climate forcing, including uplift. Our focus here is the basic long-term direction of climatic change over the middle and especially the late Cenozoic.

Because differences exist in the response of different general circulation models to the same forcing, there is always room for doubt as to what results from any model are really credible. We offer two arguments that our results are robust.

First, as summarized in Figure 9, the changes in atmospheric circulation and surface climate predicted by uplift from the NM to the M experiments are geographically coherent consequences of three basic physical processes: (1) increased orographic diversion of the westerly circulation (planetary waves) and upslope/downslope vertical motions, especially in winter; (2) increased low-level cyclonic/anticyclonic circulation around the elevated plateaus, which form heat sources/sinks in summer/winter; and (3) intensified vertical motion in large-scale east-west atmospheric cells due to the fundamental shifts of atmospheric mass between the ocean and the summer-heated and winter-cooled plateaus. (Note that changes in vertical motion are not shown in Figure 9.)

Second, as noted here and by *Kutzbach et al.* [this issue], the model predictions in large part match those from other orography experiments with GCMs [*Manabe and Terpstra*, 1974; *Hahn and Manabe*, 1975] and linear barotropic models [*Held*, 1983]. There is also a need to consider that simulated responses to uplift (or any other climate forcing) are in part a function of grid box resolution [*Rind*, 1988].

Several of the climatic trends potentially explained by the results of the uplift experiment first developed or markedly accelerated during the late Miocene and early Pliocene (10–3 Ma), just prior to the onset of northern hemisphere glaciation (Tables 1 and 2). A number of these climatic trends also continued to intensify through the late Pliocene and Pleistocene, based on evidence from interglaciations free of overprints from the periodic, orbitally driven glaciations [*Milankovitch*, 1941; *Hays et al.*, 1976].

This observed intensification toward the present of climatic trends matching the model-simulated climatic effects of uplift is consistent with geologic evidence that uplift in southern Asia and the American west accelerated in the latest Cenozoic, such that net uplift over the last 5–10 m.y. was probably comparable to that over the preceding 30–35 m.y. [*Ruddiman et al.*, this issue]. The match between the accelerating tempo of tectonic forcing and that of the regional climatic changes indicates that uplift became an important factor in Northern Hemisphere climate evolution by the late Miocene and early Pliocene. Because several of the climatic changes simulated by the progressive uplift experiments are favorable to glaciation in the locations where ice sheets eventually formed, these results also suggest that uplift could have been an important factor in the late Pliocene initiation and the mid-Pleistocene intensification of northern hemisphere glaciation.

Other Factors in Long-Term Climate Change

Other factors are also important in long-term climatic changes, although it is not yet clear that these provide strong forcing in the later part of the Cenozoic. Plate motion continental movements have had a large impact on global climate over the last 140 m.y. [*Donn and Shaw*, 1977; *Barron*, 1981, 1985], but these geographic changes have been rather small during the large climatic changes of the late Cenozoic (e.g., 200 km of opening of the Atlantic Ocean in the last 10 m.y. [*Barron et al.*, 1981]). Studies with energy balance models suggest that plate motion may have played a role in Cenozoic glaciation in the northern hemisphere but only by invoking very abrupt threshold climatic responses to small amounts of plate motion [*North and Crowley*, 1985; *Crowley et al.*, 1986].

True polar wander during the last 5 m.y. has not exceeded 1° of latitude and is not significantly different from zero [*Schneider and Kent*, 1988]. Estimates for preceding Cenozoic intervals are subject to increasing errors that render conclusive statements impossible. Very abrupt thresholds would also have to be invoked to call on true polar wander as a major factor in climate change over the last 10 m.y.

Withdrawal of shallow seas from the continents has played a role in long-term climate changes [*Hays and Pitman*, 1973]. *Barron* [1985] inferred from GCM sensitivity tests that the decreased fraction of surface area covered by ocean may be a factor in the late Cenozoic cooling and the onset of northern hemisphere glaciation. There appears, however, to be a basic mismatch between the tempo of northern hemisphere climate change and the land/sea ratio. The fastest decreases in flooded continental area in the last 100 m.y. occurred in the late Cretaceous and Paleocene without resulting in overwhelming climatic cooling or drying, whereas the much slower decreases during the last 20 m.y. coincide with very rapid cooling and regional drying.

Antarctic glaciation is another important factor that may have forced long-term climate change in the northern hemisphere, and *Barron* [1985] included changes in polar ice in general circulation model experiments for 60, 40, and 20 Ma. Increased cooling and glaciation of Antarctica occurred mainly near 36 Ma and 14 Ma [*Shackleton and Kennett*, 1975; *Savin et al.*, 1975; *Miller et al.*, 1987]. These changes significantly cooled oceanic deep waters and may have contributed to surface cooling and drying in the southern hemisphere and perhaps as far north as the African tropics [*van Zinderen Bakker and Mercer*, 1986]. The extent (and direction) of climatic linkages between the two polar regions is, however, still in doubt, both on orbital time scales [*Broecker*, 1984] and over the longer term. Even the cause of Antarctic glaciation remains unclear, in part because Antarctica has been in a polar position continuously since the Cretaceous. Forcing of long-term Antarctic climate from the northern hemisphere [*Schnitker*, 1980] is not ruled out.

Decreasing CO_2 is a possible factor in late Cenozoic climatic cooling. Coupled atmosphere-ocean climate models show strong sensitivity to changes in atmospheric CO_2 levels [*Manabe and Bryan*, 1985]. Atmospheric CO_2 levels an order of magnitude higher than today is a possible explanation of polar warmth during the Cretaceous [*Rubey*, 1951; *Barron and Washington*, 1984; *Manabe and Bryan*, 1985; *Schneider et al.*, 1985], with a subsequent CO_2 decrease and polar cooling. Modeling of the likely CO_2 input from ridgecrest and volcanic processes, however, suggests that most of the CO_2 decrease from this source occurred prior to the Cenozoic, with almost no change in the last 30 m.y. [*Berner et al.*, 1983; *Lasaga et al.*, 1985]. This model, however, omits other possible sources of long-term CO_2 change. *Raymo et al.* [1988] proposed that increased chemical erosion in rapidly uplifting areas could reduce atmospheric CO_2; this would provide additional climatic forcing in the late Cenozoic. At this point, however, no firm geologic evidence constrains the late Cenozoic trend of atmospheric CO_2.

In summary, many of the above factors play a role in long-term climate change, but their effects during the late Cenozoic generally appear to be either modest in amplitude or poorly constrained. In contrast, the results of the sensitivity experiments for uplift in southern Asia and the American west explicitly match the direction of a large number of observed late Cenozoic climatic changes in the northern hemisphere. Based on this, we rank uplift as an important factor in the evolution of northern hemisphere climate during the middle and late Cenozoic.

Prior Hypotheses Linking Uplift and Climate

In previous work, uplift has often been acknowledged as a possible contributing factor in Cenozoic climate change but usually at local spatial scales [e.g., *Leopold*, 1967; *Axelrod*, 1977; *Crowell and Frakes*, 1970; *Hay*, 1984; *Barron*, 1985]. The most commonly cited view is that mountain uplift interrupted the more zonal atmospheric circulation of the earlier Cenozoic and increased the "continentality" of climate, causing colder winters (via lapse rate effects) and more uneven seasonal precipitation (via rain-shadow effects). In most such papers, these cooling/drying effects are specified only in the region of mountain uplift, but a few studies have speculated about larger-scale regional cooling [*Ramsay*, 1924; *MacGinitie*, 1958; *Hamilton*, 1968].

Barron [1985] inferred from GCM simulations that uplift had little effect on the global cooling trend in the Tertiary but that it might be important in regional-scale responses. His experiment simulated annual-average conditions and therefore could not explore the large changes in winter planetary wave patterns and summer monsoon circulations that are the dominant features of our results. Changes in geography due to plate motion produced the largest cooling trend in his simulations, but the cooling was only a fraction of his estimate of the required cooling. He thus called on a sizeable CO_2 decrease as a possible explanation of the Tertiary global cooling trend.

Several "topographic" hypotheses have proposed uplift as an explanation of the onset of the northern hemisphere ice age [e.g., *Flint*, 1957; *Emiliani and Geiss*, 1959]. These efforts note that late Cenozoic uplift has occurred in many regions but focus mainly on uplift in mountain and highland regions of northeastern Canada where ice sheets are thought to have nucleated. The implied mechanism for initiating glaciation is lapse rate cooling of high terrain in the areas where ice sheets form.

More recently, *Ruddiman et al.* [1986] and *Ruddiman and Raymo* [1988] used previous uplift experiments of *Manabe and Terpstra* [1974] as a basis for proposing that plateau and mountain uplift would induce southward meanders in the mean positions of the planetary waves over North America and Europe, resulting in large coolings in regions where ice sheets formed. This provided a mechanism for projecting a large-scale cooling into lower-lying areas well beyond the regions of strongest uplift. Also, *Prell* [1984] noted that Tibetan uplift might intensify the southeast Asian monsoon, consistent with recently increased attention to this subject by Asian scientists [*Whyte*, 1984; *Liu*, 1985].

The results of our uplift sensitivity experiments support most of the earlier and more recent ideas about the importance of uplift in producing both local and regional climatic effects. Our results place greater emphasis on the effects of broad plateaus rather than high mountain terrain (Figure 9). And they expand substantially the geographic scope of likely effects to include much of the northern hemisphere north of 30°N (Tables 1 and 2).

Although plateau and mountain uplift can explain much of the direction (and thus pattern) of climate changes in the northern hemisphere, we do not claim that it can fully account for the large magnitude of climatic cooling at high northern latitudes during the late Cenozoic. Similarly, as noted above, *Barron* [1985] found that the combined effects of geographic and topographic changes also failed to explain the full amplitude of late Cenozoic cooling and that other factors appear to be required.

Two of the potential additional explanations for global cooling are also related to uplift: (1) decreasing atmospheric CO_2 levels due to increased chemical weathering of uplifted terrain [*Raymo et al.*, 1988] and (2) increased aerosol loading due to uplift-induced aridification, particularly in the Gobi desert region of central Asia and the Sahara desert in northern Africa. Preliminary modeling experiments suggest that both global and regional climates may be sensitive to large changes in atmospheric dust loading [*Coakley and Cess*, 1985; *Harvey*, 1988], although the projected amplitudes of these climatic effects are not yet tightly constrained.

Our results also suggest that the critical long-term link between climate and tectonics may involve not only orogenic regions (such as Tibet and the Himalayas) but also nonorogenic or postorogenic regions effected by epirogenic uplift (most of the American west). Our results further imply that the effects of broad-scale uplift on climate will depend critically on the specific geographic siting of the uplifted regions with respect to latitude, longitude, and land-sea distributions. The basic principles of orographic and monsoonal changes in atmospheric circulation shown in Figure 9 provide a conceptual framework for predicting uplift-induced climatic changes for specific distributions of land, ocean, and uplifted regions during earlier intervals of geologic history.

7. FUTURE TESTS OF UPLIFT FORCING OF CENOZOIC CLIMATE

There are several critical limitations of the model used here. Future experiments with more complex models will remove some of these limitations and no doubt alter to some extent these estimates of the climatic effects of uplift. In several cases, these more comprehensive models may increase the amplitude of uplift-induced climatic changes simulated by the model used here.

Particularly interesting in this respect should be the strong winter but much weaker summer coolings in the regions where ice sheets began to form in the Pliocene and Pleistocene. These coolings, favorable to glaciation in sign but only to a very modest degree in amplitude, may increase substantially due to inclusion of additional feedbacks noted below. This should permit a more conclusive test of the idea that uplift of these plateaus caused Northern Hemisphere glaciation [*Ruddiman et al.*, 1986; *Ruddiman and Raymo*, 1988].

Over land, snow cover is held constant at modern limits for all experiments (42°N in winter, 69°N in summer). For the experiments with lower mountain/plateau elevation, this constraint keeps surface air temperatures unrealistically cool near the southern snow cover limits. Had the snow cover been interactive with the atmosphere, the more southerly wind directions over several northern hemisphere regions in the NM case might have melted back the snow and warmed the lower atmosphere. A warmer no-mountain case would then translate into an increased cooling due to uplift in regions of North America, Europe, and Asia that already show significant winter coolings (Figures 5 and 8). Similar changes might also be expected over land as a consequence of removing the fixed constraint of cool surface water temperatures and sea ice in interior bays and coastal oceans used in these experiments.

Soil moisture was also prescribed in these experiments. It is likely that an interactive soil moisture parameterization will amplify uplift-induced wetting and drying effects obtained from the current model [e.g., *Gallimore and Kutzbach*, 1989]. Finally, integrations with full seasonal cycles will provide assessments of the full seasonal range of climates rather than those of "perpetual" January and July.

To test further all of these effects, we have begun an experiment with a model in which the land snow cover, the ocean mixed layer, the soil moisture, and sea ice are all interactive with the atmosphere. Dynamical feedbacks connected with heating changes resulting from these interactive components may also alter flow patterns, and this could in turn affect the longitudinal structure of climatic responses.

Finally, we reemphasize the fundamental success of this sensitivity test of tectonic uplift in capturing the direction of late Cenozoic climatic changes in many areas. Uplift is thus a potent explanation of the regional differentiation of temperature and precipitation responses in the northern hemisphere during the late Cenozoic. At the same time, we also caution that results from these progressive uplift experiments fall far short of explaining the full amplitude of Cenozoic cooling, especially at high northern latitudes, as did earlier modeling studies involving a broader range of geographic changes [*Barron*, 1985]. Future model experiments will better assess effects of uplift on the amplitude of Cenozoic climate changes.

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