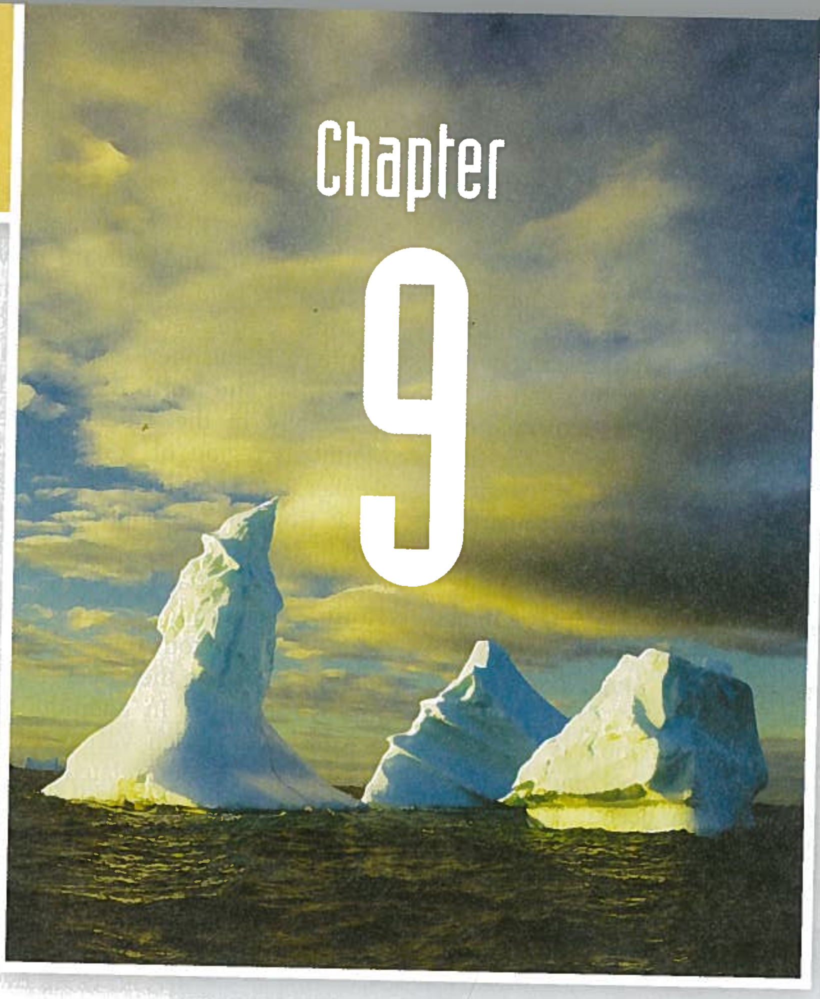


Chapter 9

Insolation Control of Monsoons

Monsoonal circulations exist on Earth today because the land responds to seasonal changes in solar radiation much more quickly than the ocean does. The reason for these differing rates of response is that the land has a far smaller capacity to store heat than the ocean. In this chapter, we will examine evidence showing that changes in insolation have been the primary driver of summer monsoons over orbital time scales. Variations in insolation produced large changes in heating of tropical landmasses and in the strength of summer monsoons at the 23,000-year cycle of precession. Just 11,000 years ago, strong summer insolation drove a vigorous summer monsoon circulation in North Africa, and the southern part of the present Sahara Desert was dotted with lakes, seasonally flowing rivers, and grassland vegetation. Similar changes have occurred across the northern tropics for millions of years. Precession-driven changes in monsoon strength have also occurred in the Southern Hemisphere, but with opposite timing. Especially large changes occurred 200 million years ago across the northern tropics of the supercontinent Pangaea.

Monsoonal Circulations

In summer, strong solar radiation causes a rapid and large warming of the land, but a slower and much less intense warming of the ocean. Rapid heating over the continents causes air to warm, expand, and rise, and the upward movement of air creates an area of low pressure at the surface (see Chapter 2). The air flowing toward this low-pressure region also warms and rises (Figure 9-1A). The air arriving from nearby oceans carries water vapor that condenses and contributes to monsoonal rainfall. Continental regions far from the ocean or protected behind intervening mountain ranges may lie beyond the reach of oceanic moisture and, therefore, may bake under the strong summer radiation.

During winter, when solar radiation is weaker, air over the land cools off rapidly, becomes denser than the air over the still-warm ocean, and sinks from higher levels in the atmosphere. This downward movement creates a region of high pressure over the land, in contrast to the lower pressure over the still-warm oceans. The overall atmospheric flow in winter is a downward-and-outward movement of cold, dry air from the land to the sea (Figure 9-1B).

Most strong summer monsoons occur in the Northern Hemisphere because landmasses there are large (Asia and North Africa) and elevations are high (especially in the Tibetan Plateau and Himalayan Mountain region of southern Asia). Monsoons are weaker in the Southern Hemisphere, where landmasses at tropical and subtropical latitudes are smaller, and high topography is more limited in extent.

Here the initial focus is on past variations of the North African monsoon for two reasons. First, North Africa lies far from the high-latitude ice sheets that might complicate the direct response of land surfaces to solar heating. In addition, the nearby seas and oceans yield a rich variety of climate records that document monsoon-related signals on North Africa.

Africa is a deceptively large landmass compared to its appearance on Mercator maps. It stretches from 37°N to 35°S, with far more of its land area north of

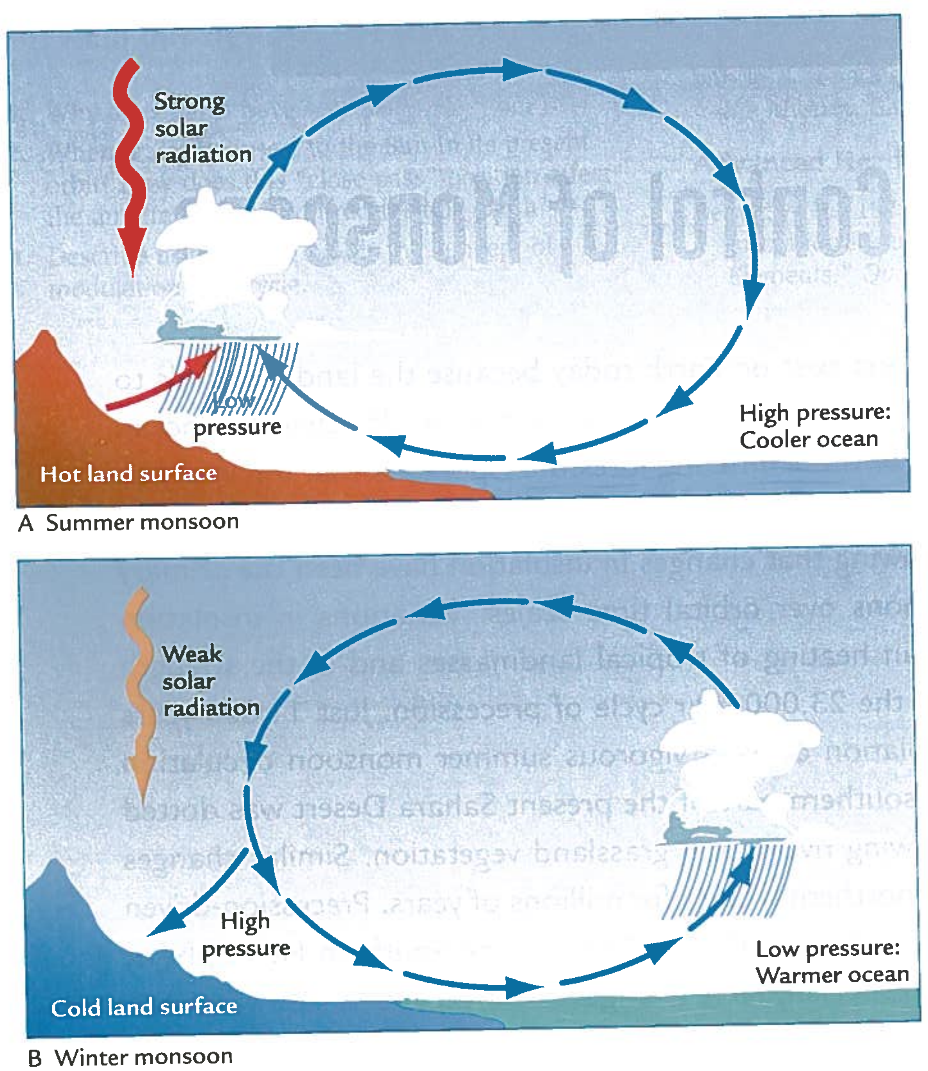


FIGURE 9-1
Seasonal monsoon circulations
Seasonal changes in the strength of solar radiation affect the surface of the land more than the ocean. In summer, intense solar heating of the land causes an in-and-up circulation of moist air from the ocean (A). In winter, weak solar radiation allows the land to cool off and creates a down-and-out circulation of cold dry air (B).

the equator (in fact an area almost twice that of the U.S. mainland). Because the huge North African land surface is situated at tropical and lower subtropical latitudes, it is strongly influenced by the overhead Sun.

As a result of strong solar heating during northern hemisphere summer, a low-pressure region develops over west-central North Africa and draws in moisture-bearing winds from the tropical Atlantic Ocean (Figure 9-2A). During typical summers, this monsoonal rainfall penetrates northward to 17°N latitude (the southern edge of the Sahara Desert) before retreating southward later in the year.

During northern hemisphere winter, the overhead Sun moves to the Southern Hemisphere, and solar radiation over North Africa is weaker. Cooling of the North African land surface by back radiation causes sinking of air from above, and a high-pressure cell develops at the surface over the northwestern Sahara Desert (Figure 9-2B). Strong and persistent trade winds associated with this high-pressure cell and with similar circulation over the adjacent North Atlantic blow southwestward from North Africa across the tropical Atlantic.

Because the trade winds of the winter monsoon carry little moisture, winter precipitation is rare in North Africa. Only two areas receive much rain during this season: (1) the northernmost Mediterranean margin, where storms occasionally form over the nearby ocean, and (2) the southwest equatorial coast (the Ivory Coast), where the moist intertropical convergence zone (ITCZ) remains over the land.

Because most of the rainfall in North Africa occurs in association with the summer monsoons, the distribution of major vegetation types reflects the monsoonal delivery of precipitation from the south (Figure 9-3). Rain forest in the year-round wet climate near the equator gives way northward to a sequence of progressively drier vegetation, first the trees and grasses of the savannas of the Sahel region and then the desert scrub of the arid Sahara.

9-1 Orbital-Scale Control of Summer Monsoons

The idea that changes in insolation control the strength of monsoons over orbital time scales was proposed by the atmospheric scientist John Kutzbach in the early 1980s, having been anticipated in part by the astronomer Rudolf Spitaler late in the nineteenth century. This concept is called the **orbital monsoon hypothesis**.

The orbital monsoon hypothesis is a direct logical extension of factors at work in modern monsoonal circulations (Figure 9-4). Because seasonal monsoon circulations are driven in the modern world by changes in solar radiation, orbital-scale changes in summer and winter insolation (see Chapter 8) should have produced a similar response. When summer insolation was higher in the past than today, the summer monsoon circulation should have been stronger, with greater heating of the land, stronger rising motion, more inflow of moist ocean air, and more rainfall (Figure 9-4B). Conversely, summer insolation levels lower than those today should have driven a weaker summer monsoon in the past.

The same kind of reasoning applies to the winter monsoon. Winter insolation minima weaker than the one occurring today should have enhanced the cooling of the land surface, which should have driven a stronger down-and-out flow of dry air from land to sea (Figure 9-4C).

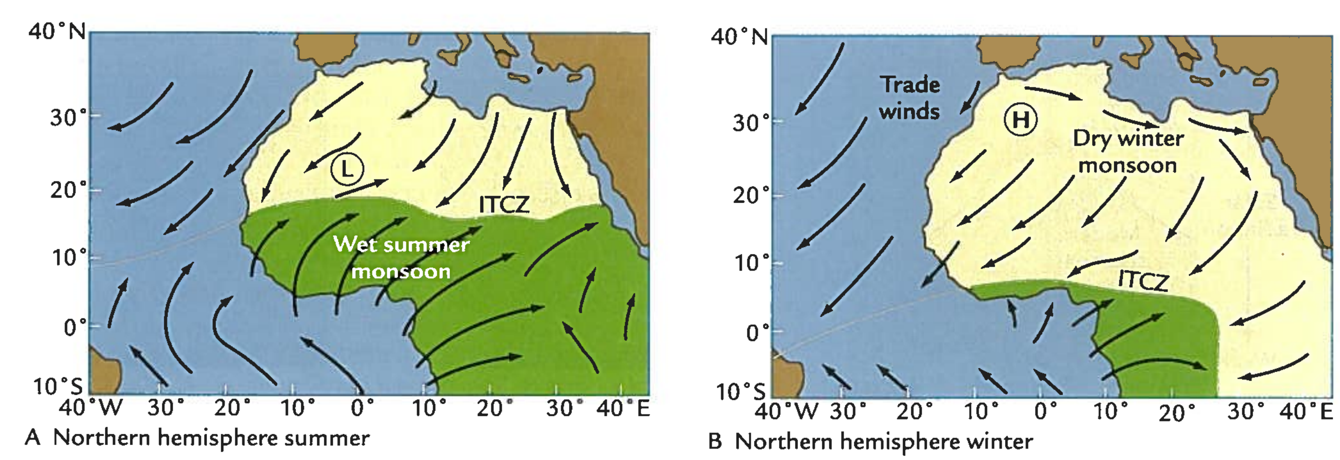


FIGURE 9-2
Monsoonal circulations over North Africa
Seasonal changes cause a moist inflow of monsoonal air toward a low-pressure center over North Africa in summer (A), and a dry monsoonal outflow from a high-pressure center over the land in winter (B). (ADAPTED FROM J. F. GRIFFITHS, *CLIMATES OF AFRICA* [AMSTERDAM: ELSEVIER, 1972].)

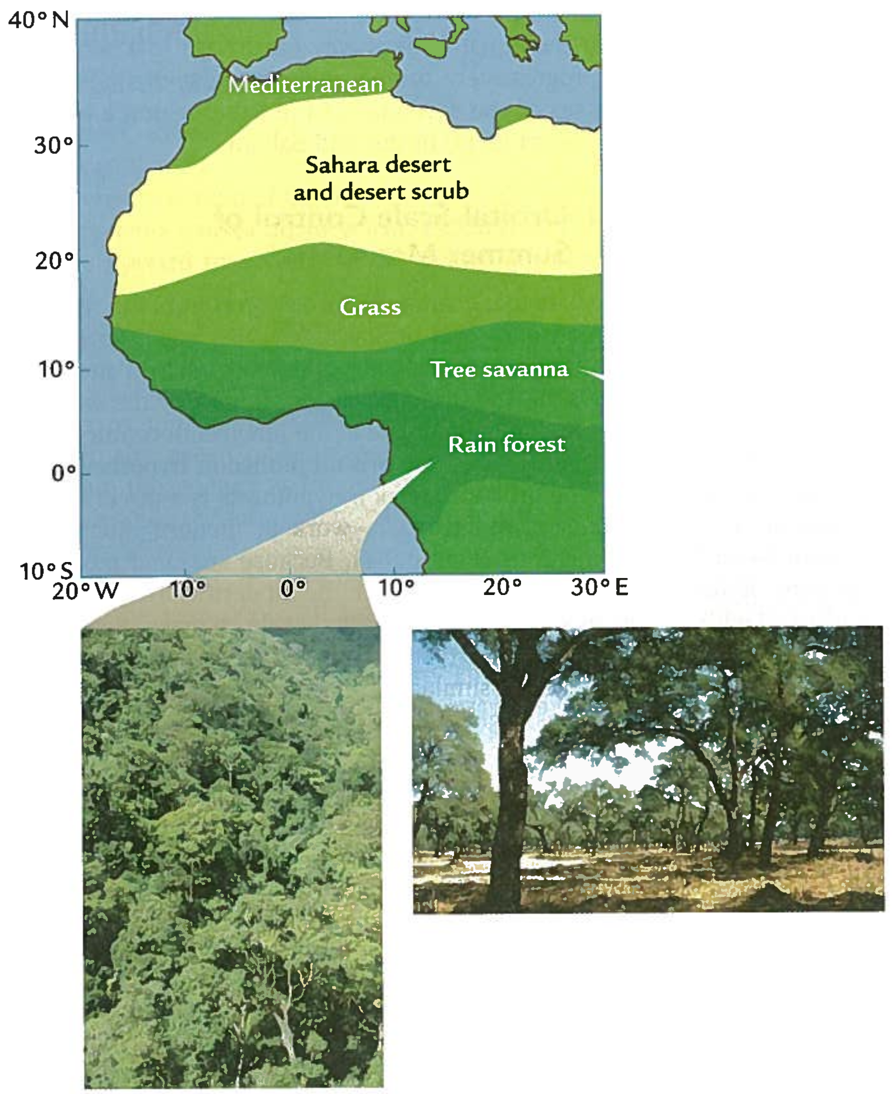


FIGURE 9-3
Vegetation in North Africa

Vegetation across the northern part of Africa ranges from rain forest near the equator to savanna and grassland in the Sahel to desert scrub vegetation in the Sahara. This pattern reflects the diminishing northward reach of summer monsoon moisture from the tropical Atlantic. (ADAPTED FROM J. F. GRIFFITHS, *CLIMATES OF AFRICA* [AMSTERDAM: ELSEVIER, 1972]. INSET PHOTOS: COURTESY OF TOM SMITH, UNIVERSITY OF VIRGINIA.)

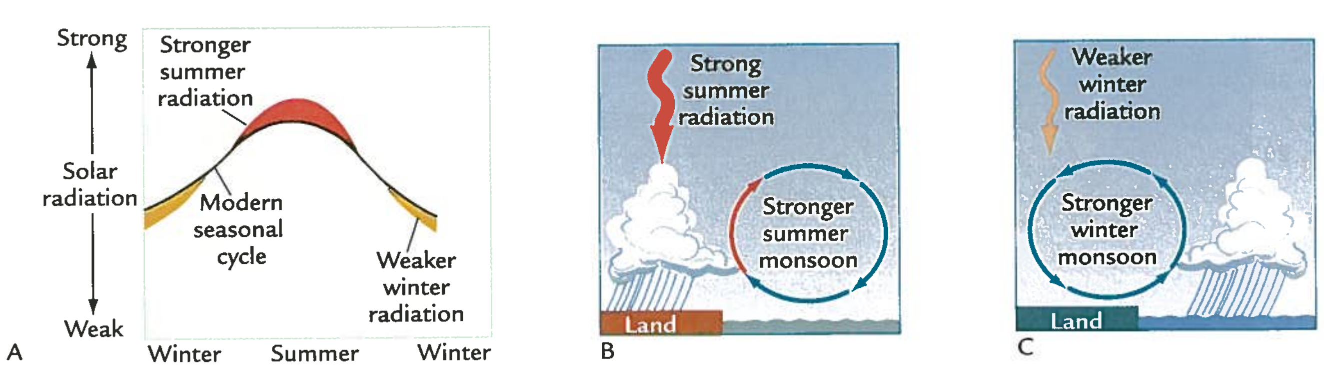


FIGURE 9-4
The orbital monsoon hypothesis

Departures from the modern seasonal cycle of solar radiation have driven stronger monsoonal circulations in the past (A). Greater summer radiation intensified the wet summer monsoon (B), while decreased winter insolation intensified the dry winter monsoon (C). (ADAPTED FROM J. E. KUTZBACH AND T. WEBB III, "LATE QUATERNARY CLIMATIC AND VEGETATIONAL CHANGE IN EASTERN NORTH AMERICA: CONCEPTS, MODELS, AND DATA," IN *QUATERNARY LANDSCAPES*, ED. L. C. K. SHANE AND E. J. CUSHING [MINNEAPOLIS: UNIVERSITY OF MINNESOTA PRESS, 1991].)

Recall from Chapter 8 that more intense summer insolation maxima and deeper winter insolation minima occur together at any one location. As a result, stronger in-and-up monsoonal flows in summer should occur at the same times in the past as stronger down-and-out monsoonal flows in winter.

At first, it might seem that the climatic effects of these opposed insolation trends in the two seasons might cancel each other, but this is not the case for annual precipitation. Monsoonal winters are always dry, regardless of the amount of insolation, because the air descending from higher in the atmosphere holds very little moisture (Figure 9-4C). As a result, orbital-scale changes in winter insolation have little or no effect on annual rainfall. In contrast, because summer monsoon winds coming in from the ocean carry abundant moisture, orbital-scale changes in summer insolation have the dominant impact on annual rainfall.

This imbalance is an example of a **nonlinear response** of the climate system to insolation: the amount of rainfall is highly sensitive to insolation change in one season (summer) but largely insensitive to changes in the other (winter). As a result, the annual response of the system has a strong summer signature even though the equal-but-opposite insolation trends in the two seasons might have been expected to cancel each other.

Orbital-Scale Changes in North African Summer Monsoons

The orbital monsoon hypothesis can be tested against a wide range of evidence, including changes in lake levels across arid regions like North Africa. At low and middle latitudes, changes in the amount of incoming solar insolation follow the 23,000-year rhythm of orbital precession (recall Chapter 8). A June insolation curve from latitude 30°N covering the last 140,000 years clearly shows this 23,000-year tempo (Figure 9-5). Note that today's June insolation level is well below the longer-term average. The orbital monsoon hypothesis predicts that stronger summer monsoons should have occurred at those times in the past when summer insolation values were significantly larger than the modern value.

The most recent instance when summer insolation values were substantially higher than today occurred near 11,000 years ago. Evidence we will examine in Part IV of this book shows that lakes across tropical and subtropical North Africa were at much higher levels during this insolation maximum, and for some time afterward. In fact, many lakes that were filled to high levels at that time are completely dry today. This evidence indicates that a strengthened summer monsoon circulation reached much farther northward into North Africa 11,000 years ago than it does today, in agreement with the orbital monsoon hypothesis.

But this interval was only one brief period in a long history of rainfall changes in North Africa. What about the much longer-term behavior of the summer monsoon? We can use the June 30°N insolation curve in Figure 9-5 to construct a simple conceptual model that predicts how the summer monsoon could have varied with time if the orbital monsoon hypothesis is valid. The predicted monsoon response is based on three assumptions.

First, we assume that a **threshold level** of insolation exists, below which the monsoon response will be too weak to leave any evidence in the geologic record (see Figure 9-5). A good example is the level of lakes in arid regions. The orbital monsoon hypothesis predicts that the higher the level of summer insolation, the stronger the summer monsoon rainfall, and the higher the lake levels. But if the insolation value falls below a certain level, the weakened summer monsoon may produce so little rain that the lakes dry up

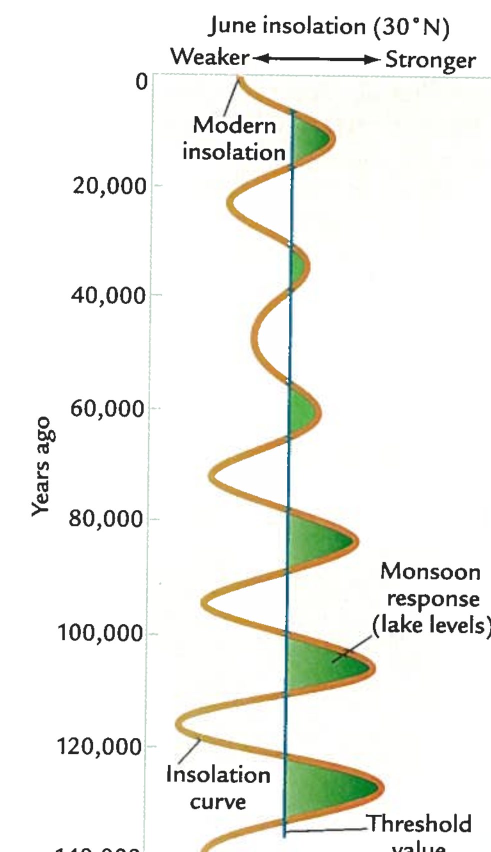


FIGURE 9-5
Conceptual model of monsoon response to summer insolation

Increases in summer insolation heating above a critical threshold value drive a strong monsoon response at the 23,000-year tempo of orbital precession. The amplitude of this strong monsoon response is related to the size of the increase in summer insolation forcing.

completely. A dry lake can no longer register changes toward even drier climates because it has reached the threshold limit of its ability to record climate.

This assumption has a good basis in fact. Many lakes in North Africa that existed 11,000 years ago during the strong monsoon interval are dry today, even though a weak summer monsoon still occurs at present in response to today's low levels of summer radiation. Apparently, it takes a threshold insolation value well above the present level to bring most North African lakes into existence.

Second, we assume that the strength of the monsoon response (such as the water level of the North African lakes) is directly proportional to the amount by which summer insolation exceeds the threshold value. This assumption has a reasonable physical basis: stronger insolation should drive stronger monsoons and fill lakes to higher levels.

Third, we assume that the strength of the monsoon in the past, as recorded in lake level records, is a composite of the average monsoon strength over many individual summers. Actually, this is more observed fact than assumption: the wet monsoonal circulations that develop every summer today inevitably cease during the following winter. The lakes fill because of the integrated effect of many wet summers. When scientists sample records of these changes, they are looking at year-by-year responses blended into a longer-term average over hundreds or even thousands of summers.

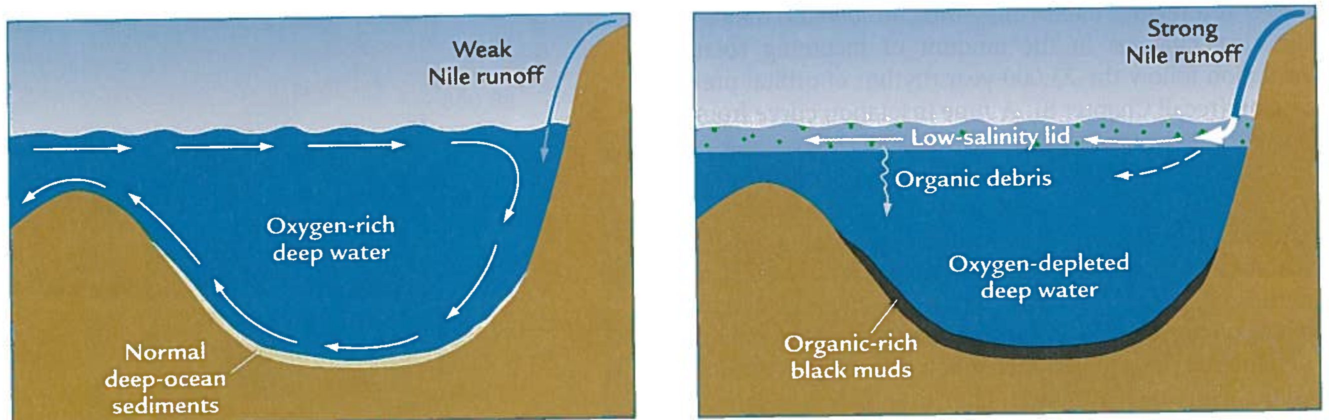
Based on these three assumptions, we arrive at the predicted monsoon response shown by the green shading in Figure 9-5. Insolation maxima above the threshold value produce a series of pulse-like monsoon (and lake) maxima at regular 23,000-year intervals. These pulses vary in strength according to the amount by which summer insolation exceeds the threshold value. Insolation levels below the threshold value leave no monsoonal evidence in the geologic record.

The strongest predicted monsoon peaks in Figure 9-5 occur between 85,000 and 130,000 years ago, when the summer insolation curve reached large maxima because of modulation of the precession signal by orbital eccentricity (see Chapter 8). In contrast, the weaker insolation maxima near 35,000 and 60,000 years ago should have produced less powerful monsoons. All long-term summer insolation minima (such as the one we are in today) fall below the critical threshold for major lake filling.

We can examine several climate records for evidence of this predicted monsoon response. Because most of North Africa is arid, and because erosion of its sediments is much more prevalent than deposition, its climate history is sparse and difficult to date. Fortunately, the nearby seas and oceans contain continuous and well-dated records.

9-2 "Stinky Muds" in the Mediterranean

The water that fills the Mediterranean Sea today has a high oxygen content. Near-surface waters are well oxygenated because they exchange oxygen-rich air with the atmosphere and because photosynthesis by marine organisms produces O_2. The high oxygen content of waters deeper in the basin results from sinking of oxygen-rich surface water during winter (Figure 9-6A).



A Weak summer monsoon

B Strong summer monsoon

FIGURE 9-6
Mediterranean circulation and monsoons
In today's Mediterranean circulation, salty surface water chilled by cold air in winter sinks and carries dissolved oxygen to deeper layers (A). At intervals in the past, strong summer monsoons in tropical Africa caused an increased discharge of Nile River fresh water into the eastern Mediterranean, creating a low-density surface-water lid that inhibited sinking of surface water and caused the deep ocean to lose its oxygen and deposit organic-rich black muds (B).

This sinking motion is a result of two factors (see Chapter 2): (1) the high salt content of the Mediterranean Sea, caused by the excess of summer evaporation over precipitation, and (2) winter chilling of salty water along the northern margins of the Mediterranean Sea during incursions of cold air from the north. These two factors make surface waters dense enough to sink to great depths in winter. The dense waters that sink into the deep Mediterranean Sea eventually exit westward into the Atlantic Ocean. As a result of this large-scale flow, the floor of the present Mediterranean Sea is covered by sediments typical of well-oxygenated ocean basins: light tan silty mud containing shells of plankton that once lived at the sea surface and benthic foraminifera that once lived on the seafloor.

Mediterranean sediments also contain distinct layers of black organic-rich muds, called **sapropels**. Their high organic carbon content indicates that they formed at times when the waters at the seafloor were **anoxic**: they lacked the oxygen needed to convert (oxidize) organic carbon to inorganic form. The lack of oxygen led to stagnation of the deep waters and deposition of iron sulfides, giving the sediments a "stinky" (rotten egg) odor. The lack of oxygen also kept benthic foraminifera and other creatures from living on the seafloor.

The paleoecologist Martine Rossignol-Strick proposed that these sinky muds mark times when the deep Mediterranean Basin was deprived of oxygen because sinking of oxygen-rich surface waters was prevented by a cap of low-density freshwater brought in by river water (see Figure 9-6B). Even though the surface waters were still chilled by cold winter air masses, the low-salinity lid kept them from becoming dense enough to sink deep into the basin. As a result, the deep Mediterranean Basin lost its supply of oxygen.

At the same time, production of planktic organisms continued at the surface, and probably even increased as the stronger river inflow delivered extra nutrients (food) to the Mediterranean. The high productivity at the surface continued to send organic-rich remains of dead plankton toward the seafloor. Sinking and oxidation of this organic carbon continually depleted the oxygen levels in the deep Mediterranean and produced the stinky muds on the seafloor.

The most recent sapropel in the eastern Mediterranean dates from 10,000 to 8,000 years ago, an interval when summer insolation levels were higher than today, the African summer monsoon was stronger, and African lakes were at higher levels. Earlier layers of organic-rich mud deeper in Mediterranean sediment cores occur at regular 23,000-year intervals during times when summer insolation was higher than it is today. The sapropels were best developed (thickest and most carbon-rich) near the time of the strongest summer insolation maxima, but poorly developed during weaker insolation maxima, and absent the rest of the time. This history of sapropel deposition matches very well the conceptual pattern predicted by the orbital monsoon hypothesis (see Figure 9-5). This close match indicates a connection of some kind to the low-latitude monsoon over North Africa.

Initially, some climate scientists questioned this explanation. The Mediterranean Sea lies at high subtropical latitudes (30°–40°N), beyond the greatest northward expansions of past summer monsoons indicated by lake-level evidence across North Africa. If climate in the confines of the Mediterranean region never became truly monsoonal, how could the stinky muds deposited in that basin be a response to the North African monsoon?

The critical link turned out to be the Nile River (Figure 9-7), which gathers most of its water from the highlands of eastern North Africa at tropical latitudes much farther south. Even today these highlands

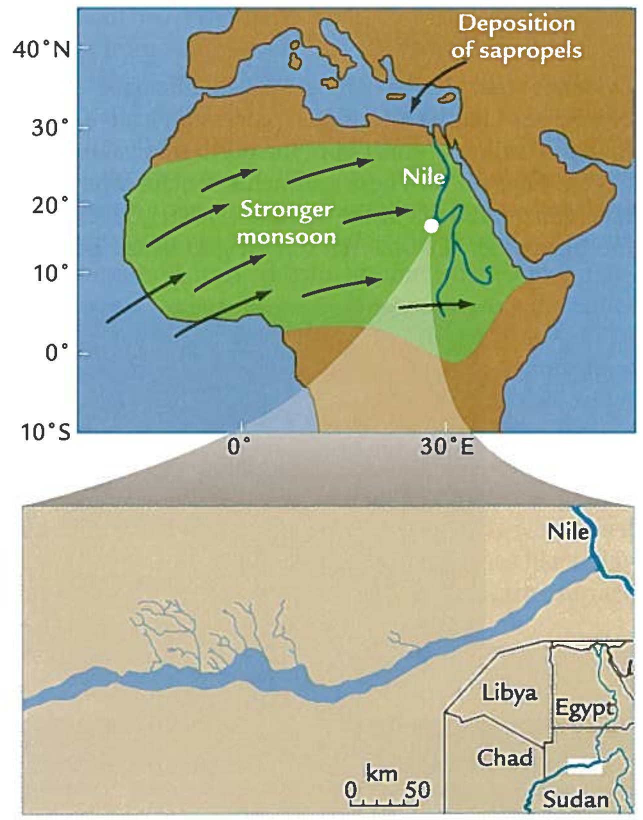


FIGURE 9-7
Monsoons and the Nile River
Strong summer monsoons in tropical North Africa periodically produced large discharges of Nile freshwater into the Mediterranean Sea. Satellite sensors have detected riverbed sediments deposited during strong monsoons but now buried beneath sheets of sand in the hyperarid eastern Sahara Desert (inset). (INSET: ADAPTED FROM H.-J. PACHUR AND S. KROPLEIN, "WADI HOWAR: PALEOCLIMATIC EVIDENCE FROM AN EXTINCT RIVER SYSTEM IN THE SOUTHEASTERN SAHARA," *SCIENCE* 237 [1987]: 298-300.)

receive summer rains during the relatively weak tropical monsoon, and the Nile delivers the water to the Mediterranean Sea far to the north. At times when summer insolation was much stronger than it is today, the strengthened summer monsoon expanded northward and eastward, bringing much heavier rainfall to these high-elevation regions. In effect, rainfall in the North African tropics exerts a remote control on the salinity of the subtropical Mediterranean Sea via the Nile River connection.

Satellite sensors have detected the buried remnants of streams and rivers that once flowed across Sudan (see Figure 9-7) but are now covered by sheets of sand blowing across the hyperarid southeastern Sahara Desert. The fact that these streams once flowed eastward and joined the Nile River indicates that lower-elevation regions also contributed to the Nile's stronger flow during major monsoons.

9-3 Freshwater Diatoms in the Tropical Atlantic

Evidence that North African lakes fluctuate at the 23,000-year tempo of orbital precession can also be found in sediment cores from the north tropical Atlantic Ocean (Figure 9-8). These sediments contain layers with high concentrations of the opaline ($SiO_2 \cdot H_2O$) shells of the freshwater diatom *Aulacoseira granulata*. Because these diatoms could not have lived in the ocean, and because these ocean cores are hundreds to thousands of kilometers away from the land, the only explanation for the presence of the diatoms is wind transport. In arid and semiarid regions, winds scoop out ("deflate") sediment from the beds of dry lakes and blow the fine debris far away, some of it to the nearby oceans.

The only plausible source of these lake diatoms in these tropical Atlantic sediment cores is the one lying directly upwind in the prevailing northeasterly flow of winter trade winds—arid and semiarid North Africa. The intervals in the Atlantic cores containing freshwater diatoms mark times in the past when North African lakes were drying out and their muddy lakebeds were becoming exposed to, and eroded by, strong winter trade winds.

Records from the Atlantic sediment cores show that lake diatoms were delivered in distinct pulses separated by 23,000 years. As was the case for the Mediterranean sapropels, this 23,000-year tempo in diatom influxes is a direct indication of a connection to the tropical monsoon fluctuations in North Africa. In this case, however, each diatom pulse occurs later than the summer insolation maxima by as much as 5,000 to 6,000 years (Figure 9-9).

This delay makes sense if seen as a part of the sequence of likely events during a typical monsoon cycle. Lakes in North Africa filled to maximum size during the summer insolation maxima that drove the strong monsoons. These high lake levels deposited lakebeds rich in diatoms. Then, as summer insolation began to decrease toward the next insolation minimum, the monsoon weakened, summers became drier, and the lake levels began to drop. The fall in lake levels exposed the diatom-bearing silts and clays to winter winds, which scooped them up and blew them out to the ocean. Once the lakes had dried out completely and most of the diatom-bearing sediments had been blown away, transport of diatoms to the ocean slowed or stopped, even though the monsoon continued to weaken as summer insolation continued to fall toward the next minimum. As a result, the diatom pulses sent to the ocean lagged well behind the summer monsoon maxima because of the time needed to dry the lakes, but they preceded the subsequent summer insolation minima because many of the lakebeds had already been fully exposed just partway into the drying trend.

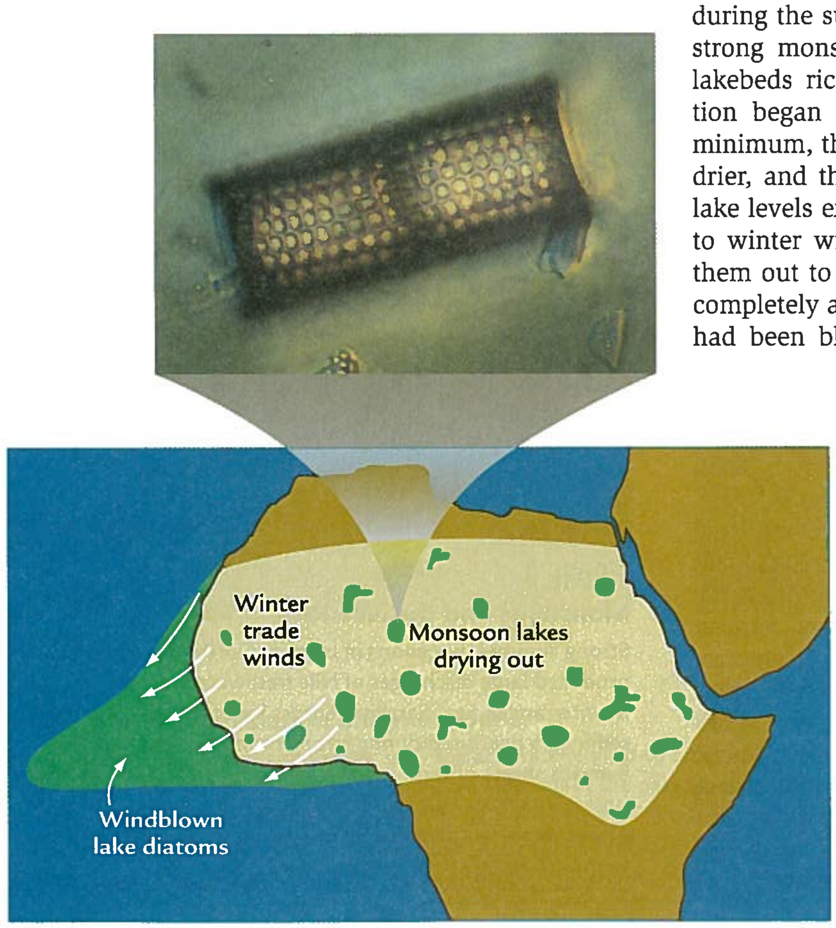


FIGURE 9-8
Drying of monsoonal lakes
North African lakes filled by strong monsoonal rains later dried out and were exposed to erosion by winds. Lake muds containing the freshwater diatom *Aulacoseira granulata* (inset) were carried by winds to the tropical Atlantic. (INSET: COURTESY OF BJORG STABELL, UNIVERSITY OF OSLO.)

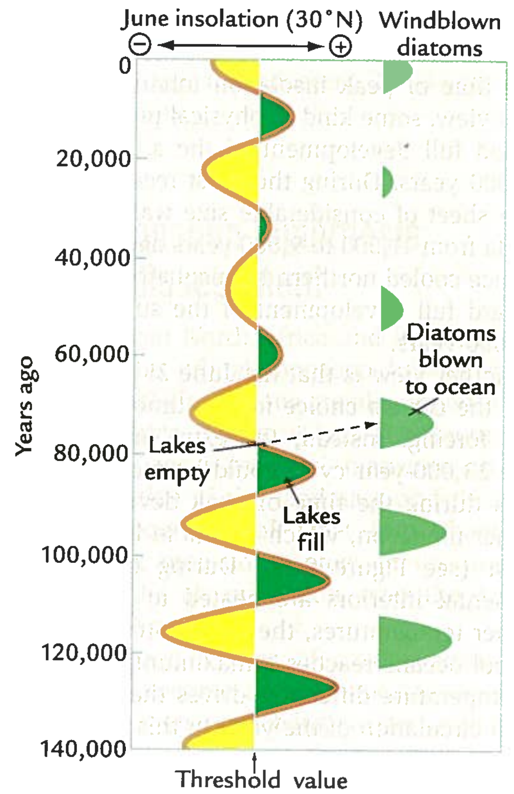


FIGURE 9-9
Delayed diatom deposition in the Atlantic
Diatoms from North African lakes were deposited in the tropical Atlantic Ocean several thousand years after the intervals of strongest monsoons, as the lakes dried out. (ADAPTED FROM W. F. RUDDIMAN, "TROPICAL ATLANTIC TERRIGENOUS FLUXES SINCE 25,000 YEARS B. P.," *MARINE GEOLOGY* 136 [1997]: 189–207; BASED ON E. M. POKRAS AND A. C. MIX, "EARTH'S PRECESSION CYCLE AND QUATERNARY CLIMATIC CHANGES IN TROPICAL AFRICA," *NATURE* 326 [1987]: 486–7.)

Another indication of a link to the North African summer monsoon comes from the amplitude of the diatom peaks. Each 23,000-year diatom pulse has the same relative strength as the immediately preceding summer insolation maximum (see Figure 9-9). This pattern is consistent with a scenario in which stronger insolation maxima drove stronger summer monsoon maxima, which created bigger lakes, which provided larger sources of diatom-bearing sediments for subsequent transport to the ocean.

9-4 Upwelling in the Equatorial Atlantic

Atlantic Ocean sediments contain additional evidence consistent with the hypothesis that the North African monsoon fluctuates at the 23,000-year tempo of orbital precession. Cores in the eastern Atlantic just south of the equator show that the structure of the upper water layers has varied with a prominent 23,000-year rhythm. Part of the reason for this response is that the North African summer monsoon imposes an atmospheric circulation pattern that overrides the local circulation.

When the North African summer monsoon is relatively weak (as it is today), trade winds along the equator have a strong east-to-west flow (Figure 9-10A). The strongest trade winds occur in southern hemisphere winter (July and August) and blow from the South Atlantic toward the equator. Part of this flow crosses the equator, turns to the northwest, and enters North Africa in the summer monsoon flow, but this part of the flow is not strong when the monsoon is weak, as it is today. Instead, strong trade winds at times like today blow mainly toward the west and drive warm surface

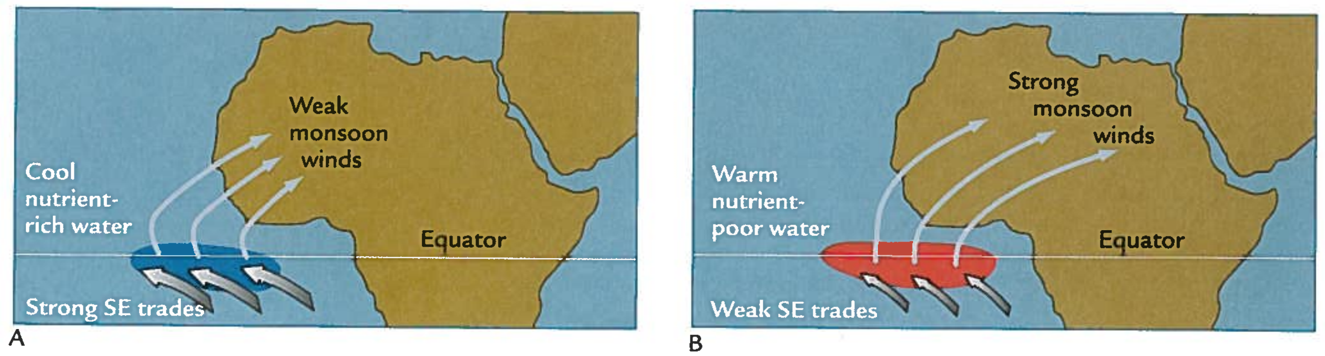


FIGURE 9-10
Effect of monsoons on southeast trade winds
When monsoonal circulation over North Africa is weak, strong southeasterly trade winds in the eastern tropical Atlantic cause cool nutrient-rich waters to rise close to the surface (A). When a strong monsoonal circulation over North Africa weakens the trade winds, tropical waters become warm and depleted in nutrients (B).

waters southward away from the equator (recall Chapter 2). This upper-ocean flow causes a shallowing of the seasonal thermocline, a subsurface region of steep temperature gradients between the warm surface waters and much cooler temperatures below. As the thermocline shallows, cooler waters rich in nutrients rise toward the sea surface just south of the equator.

In contrast, at times when summer insolation was higher than it is today, such as 11,000 years ago, the stronger summer monsoon flow overrode this circulation pattern (Figure 9-10B). A much larger portion of the southern trade-wind flow crossed the equator, turned to the northeast, and was drawn into North Africa in the monsoonal circulation. This strengthening of the monsoon flow into North Africa weakened the westward trade-wind flow along the equator, and the weaker trade winds reduced the upwelling of cold waters, leaving the surface waters poorer in nutrients from below.

Changes between these two circulation patterns over time can be measured by examining variations in the relative amounts of planktic organisms that inhabit near-surface waters and leave shells in the sediments below. In equatorial Atlantic sediments, planktic foraminifera and coccoliths are the most common shelled organisms (see Chapter 3). Different species of these two kinds of plankton prefer different environmental conditions near the sea surface—either warmer waters with fewer nutrients or cooler waters rich in nutrients. Sediment cores from the Atlantic Ocean just south of the equator show 23,000-year cycles of alternating abundances in these two types of plankton, still another indication of the effects of the North African summer monsoon.

9-5 The Phasing of Summer Monsoons

The idealized monsoon model presented so far in this chapter has suggested that peak development of past summer monsoons at the 23,000-year cycle occurred as a direct response to strong insolation forcing at the timing of the June 21 summer solstice. In fact, however, the strongest monsoons have occurred about 2,000 years later. For example, the most recent low-latitude insolation maximum was about 11,000 years ago, but the most recent Mediterranean sapropel didn't occur until 9,000 years ago. This offset has been interpreted in two ways.

Recall that Earth's precessional motion produces a family of monthly insolation curves, each offset from the preceding month by one-twelfth of a 23,000-year cycle, or slightly less than 2,000 years (see Chapter 8, Figure 8-18). This entire family of 23,000-year insolation curves is available to boost the strength of the summer monsoon, but the problem is to provide a specific physical justification for the one chosen.

One interpretation starts with the assumption that June 21 is the best choice because the summer solstice is the time of peak insolation forcing (Figure 9-11). In this view, some kind of physical process must have retarded full development of the summer monsoon by 2,000 years. During the most recent deglaciation, an ice sheet of considerable size was still present in Canada from 11,000 to 9,000 years ago, so perhaps its presence cooled northern hemisphere climate enough to retard full development of the summer monsoon for 2,000 years.

Another view is that the June 21 summer solstice is not the correct choice for the time of critical insolation forcing. Instead, the extra insolation forcing at the 23,000-year cycle could be more effective if it occurs during the time of peak development of the summer monsoon, which occurs in late July to early August (see Figure 9-11). During this month, the continental interiors are heated to their maximum summer temperatures, the temperature contrast with the cool oceans reaches a maximum, and the resulting temperature difference drives the strongest monsoonal circulation of the year. In this view, insolation changes at the 23,000-year precession cycle that are aligned with this most intense midsummer heating should have the greatest impact in boosting continental temperatures still further and driving even stronger

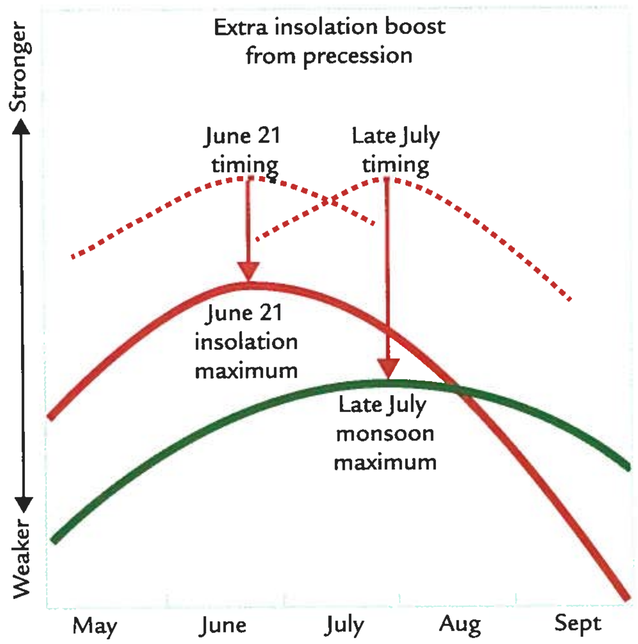


FIGURE 9-11
Alignment of insolation and summer monsoon
Peak summer monsoon development in the Northern Hemisphere occurs in late July to early August. If the increment of extra insolation caused by Earth's precession falls in late July, the summer monsoon will be stronger.

monsoons. Choosing late July or early August as the time of critical insolation forcing (rather than June 21) would eliminate the apparent 2,000-year lag in monsoon responses.

Orbital Monsoon Hypothesis: Regional Assessment

The evidence from North Africa and its surrounding seas supports the orbital monsoon hypothesis, but strong monsoonal circulations are also found on Asia and South America (Figure 9-12). The South Asian monsoon is the most powerful monsoon on Earth because of the size of the landmass and the extent of the high topography of Tibet (see Chapter 7). The only strong monsoon system in the Southern Hemisphere is the air mass flow over the Amazon Basin in South America.

In recent years, studies of cave deposits (speleothems) have become a very important source of information on monsoon changes over orbital time scales. Groundwater dripping through soils and bedrock and into caves deposits stalactites and stalagmites constructed of calcite ($CaCO_3$). These deposits build up layer by layer over thousands to tens of thousands of years, and they can be very accurately dated by radiometric analysis of small amounts of thorium and uranium (see Chapter 2). The relative amount of the two isotopes of oxygen (^{16}O and ^{18}O) in the calcite layers varies though time because of several factors, the most important of which is changes in the amount of surface precipitation that feeds the flow of groundwater into the caves. As a result, changes in overlying monsoonal air masses can be accurately reconstructed from variations in the $\delta^{18}O$ signal of cave calcite.

Records from caves in southern China and in South America both show $\delta^{18}O$ variations so large that they can only be explained by air mass changes linked to the changes in monsoon strength in those regions (Figure 9-13). Layers with highly negative $\delta^{18}O$ values indicate a very strong monsoon flow from the ocean and greater fractionation of the oxygen isotopes, and layers with less negative values indicate weaker monsoon flow from the ocean and lesser fractionation.

In China, these $\delta^{18}O$ changes correlate closely with changes in midsummer (late July) insolation at 25°N, especially during the interval prior to 100,000 years ago when both the insolation and $\delta^{18}O$ changes were largest. This record provides unambiguous evidence in support of the orbital monsoon hypothesis:

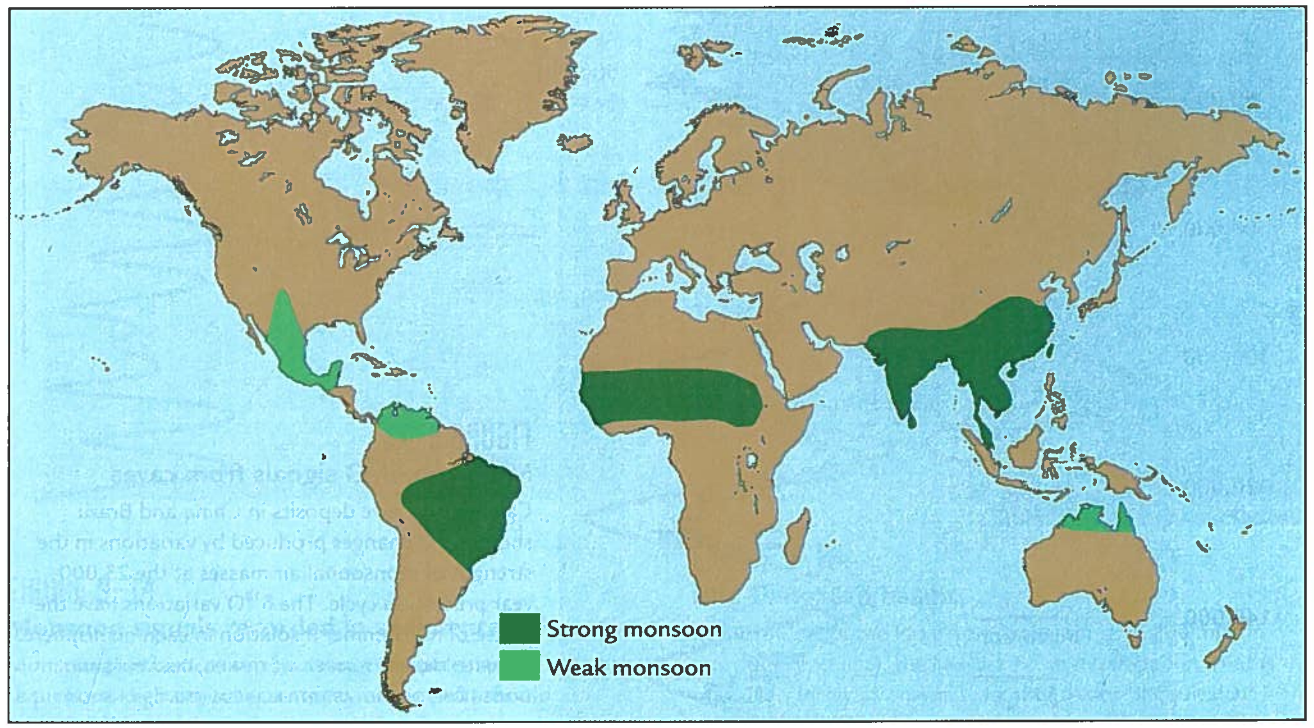


FIGURE 9-12
Major monsoon systems
Strong summer monsoons occur in North Africa, Asia, and South America.

the Asian monsoon has varied at the 23,000-year cycle and the changes had a midsummer (late July) phase. Other changes are also evident in the $\delta^{18}O$ signal over intervals shorter than the orbital-scale oscillations.

The $\delta^{18}O$ record in cave calcite from southeastern Brazil also shows a close link to summer insolation forcing, with a strong 23,000-year cycle (Figure 9-13). Note, however, that this 23,000-year cycle correlates to insolation during February rather than July. This result is an elegant confirmation of the orbital-monsoon hypothesis. Because late February is midsummer in the Southern Hemisphere, monsoon variations on southern continents should have this phase at the precession cycle, and they do.

In Summary, evidence from several continents fully supports John Kutzbach's orbital monsoon hypothesis. At this point, the hypothesis has passed so many tests that it merits the higher status of a theory (see Chapter 1).

Monsoon Forcing Earlier in Earth's History

The concept of insolation forcing of summer monsoons can be used to investigate the more distant geologic past. Orbital-scale changes in summer insolation at the precession cycle have driven changes

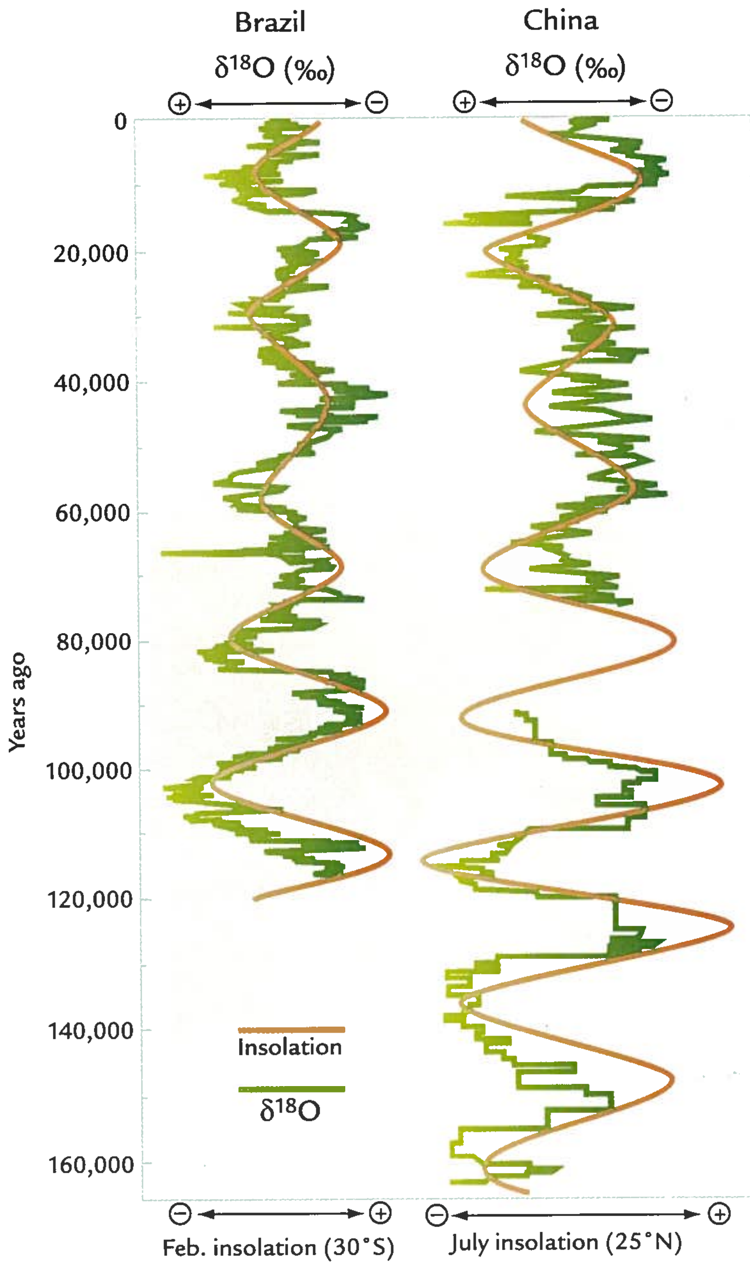


FIGURE 9-13
Monsoon $\delta^{18}O$ signals from caves
Calcite from cave deposits in China and Brazil shows $\delta^{18}O$ changes produced by variations in the strength of monsoonal air masses at the 23,000-year precession cycle. The $\delta^{18}O$ variations have the phase of midsummer insolation in each hemisphere. (ADAPTED FROM D. YUAN ET AL., "TIMING, DURATION, AND TRANSITIONS OF THE LAST INTERGLACIAL ASIAN MONSOON," *SCIENCE* 304 [2004]: 575-578; AND FROM F. W. CRUZ ET AL., "INSOLATION-DRIVEN CHANGES IN ATMOSPHERIC CIRCULATION OVER THE PAST 116,000 YEARS IN SUBTROPICAL BRAZIL," *NATURE* 434 [2005]: 63-66.)

in monsoonal precipitation, and processes linked to precipitation have left evidence in ancient climate records, such as pulses of sediment-laden runoff and changes in lake depth. As a result, many ancient sedimentary rocks contain valuable information about varying monsoon strength.

In cases where high-quality time control is available for ancient deposits, we can look for evidence of the kind of monsoon signature shown in Figure 9-5. Because many of these deposits extend over millions of years, we can expect to see records that look like those in Figure 9-14. A wide range of sediment indicators linked to precipitation, erosion, runoff, transport, and deposition may have this appearance. In some cases, this relationship can even be used to refine ("tune") time scales initially set by radiometric dating (Box 9-1).

As in the case of North African lakes, we expect the monsoon signature to show clusters of two or three strong maxima separated by clusters of two or three weaker maxima, with these clusters repeating in the record at intervals of about 100,000 years because of control of the amplitude of precession by orbital eccentricity (see Chapter 8). In this case, however, because we are looking at much longer records, we should also see clusters of monsoon-driven maxima at the longer eccentricity period near 400,000 years (see Figure 9-14). The truncation of the summer monsoon response pattern at a critical threshold value is called a **clipped response**. As a result of this truncation, many monsoon responses register only one side of each 23,000-year precession cycle, with modulation of the amplitude of this one-sided response at 100,000 and 400,000 years.

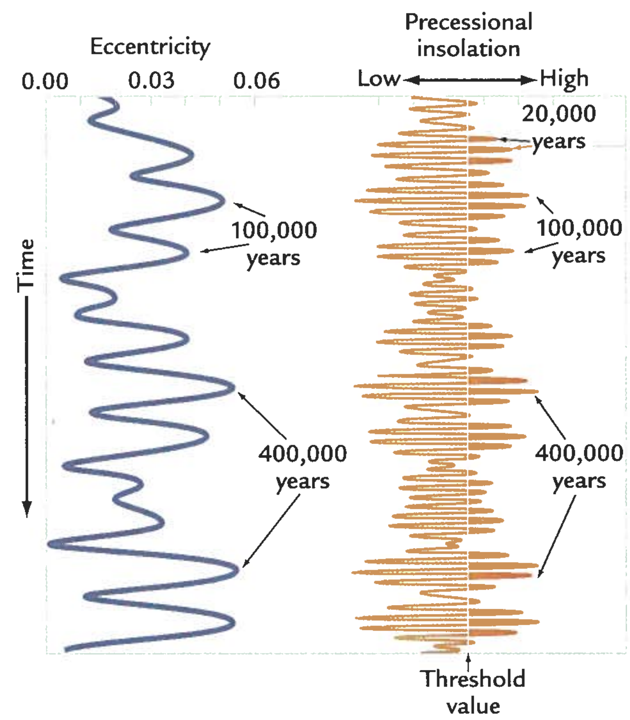


FIGURE 9-14
Monsoon signals recorded in sediments
Monsoonal influences can be detected in older sediment sequences. High orbital eccentricity values (left) should amplify individual 23,000-year precession cycles approximately every 100,000 and 400,000 years (right). The monsoon signal in the sediments could resemble the red-shaded area to the right of the threshold insolation value.

9-6 Monsoons on Pangaea 200 Million Years Ago

Just before 200 million years ago, a chain of basins formed in a region that is now the eastern United States but at that time was deep in the interior of the giant supercontinent Pangaea (Figure 9-15). These deep depressions in a region of generally high terrain

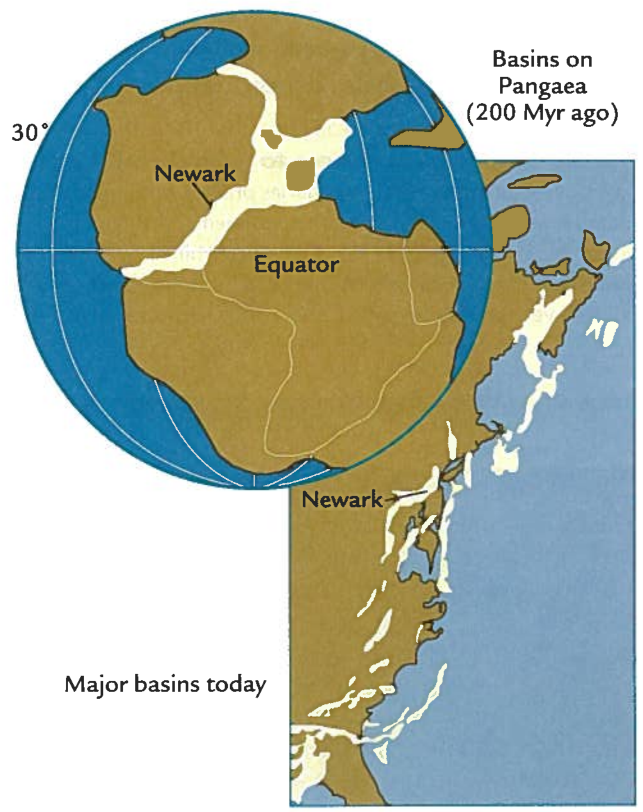


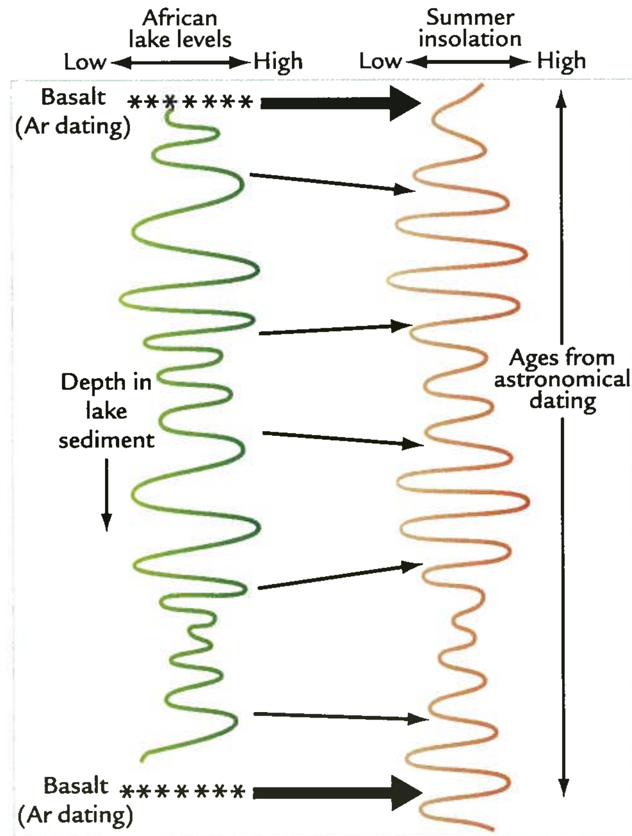
FIGURE 9-15
Mid-Pangaean basins
In the middle of the Pangaean supercontinent 200 million years ago (top left), the Newark Basin developed in what is now New Jersey as one of a chain of basins of equivalent age (bottom right). (ADAPTED FROM P. E. OLSEN AND D. V. KENT, "MILANKOVITCH CLIMATE FORCING IN THE TROPICS OF PANGAEA DURING THE LATE TRIASSIC," *PALAEOGEOGRAPHY, PALAEOCLIMATOLOGY, PALAEOECOLOGY* 122 [1996]: 1-26.)

Looking Deeper into Climate Science

Insolation-Driven Monsoon Responses: A Chronometer for Tuning

The clearly demonstrated link between summer insolation forcing and monsoon responses at low latitudes has become part of the basis for a new way of dating sedimentary records on land and in the oceans. This method, **orbital tuning**, can provide even better time resolution than radiometric methods.

The tuning method is based on the relationship between the insolation signal (the forcing) and the summer monsoon changes (the response). The timing of orbital insolation changes is known with great accuracy from astronomical calculations, and a range of monsoon responses can be measured in sediments. By making the simple assumption that the insolation driver and the monsoon responses have kept the same relationship in the past, the monsoon responses in the sediments can be dated with nearly the same accuracy as the orbital forcing.



Tuning sediment sequences to orbital variations If lavas or magnetic reversal boundaries provide radiometrically dated levels in terrestrial or marine sediments, the age of intervening sediment intervals can be determined by tuning monsoon-driven sediment responses to insolation changes at the 23,000-year precession cycle.

Box 9-1

This method is most easily applied in ocean sediments because deposition in the ocean tends to be continuous. Assume that a sediment core contains two magnetic reversal boundaries that have been dated by correlation to the global magnetic stratigraphy established by dating basalt layers on land (see Chapter 5). The ages of these reversals constrain the intervening sequence of sediments to a particular interval of time. Also assume that this sediment sequence contains a record that is directly tied to the strength of the tropical monsoons, such as a sequence of sapropel layers.

In many cases, the monsoon-related response measured in the sediments will show an obvious correlation to the summer insolation forcing. Both the insolation changes and the monsoon responses will show cycles near 23,000 years and obvious modulation of these cycles at eccentricity periods near 100,000 and 400,000 years. The tuning process (matching maxima and minima in the monsoon response to correlative features in the insolation signal) allows the ages in the sedimentary sequence to be assigned to specific precession cycles in the past at a finer resolution than the length of each cycle.

This method has also been applied to sediments deposited in long-lived lakes on continents. In regions like East Africa, volcanic eruptions deposit beds of basalt (lava) or volcanic ash that can be radiometrically dated by K/Ar methods. The volcanic deposits provide the initial time framework for the tuning process, analogous to the use of magnetic reversal boundaries in marine sediments. The monsoon-driven variations in lake size during the sequence lying between the basalt layers can then be tuned to the astronomically dated record of summer insolation. Because records on land are generally much more difficult to date than ocean sediments, tuning of these lake sequences provides an enormous improvement over other dating methods.

For sedimentary records that contain climatic responses at the cycles of both precession and obliquity, the tuning method can be tested even more rigorously. In this case, the "tuned" time scale must match not just the amplitude-modulated precession cycle at 23,000 years, but also the tilt cycle at 41,000 years. This added requirement makes the tuning process an even more demanding exercise, but also one that is even more likely to yield a uniquely accurate time scale.

were formed by precursors of the forces that would eventually pull the Pangaean continent apart and create the Atlantic Ocean. But this part of Pangaea would not break up until tens of millions of years later (see Chapter 7).

Sediments deposited in one of these depressions, the Newark Basin in modern New Jersey, have been extensively investigated. The fossil compasses provided by magnetic evidence from volcanic rocks indicate that the Newark Basin was located in the tropics 200 million years ago, about 10° of latitude north of the equator (see Figure 9-15). Because of its tropical location, the Newark Basin was situated in a regime dominated by precessional insolation changes, similar to those in modern North Africa and southern Asia. Because the basin was far from the ocean, its climate was relatively arid, but enough moisture arrived to create a lake that varied greatly in size over time.

Evidence preserved in a thick (greater than 7,000 m) sequence of lake sediments shows that the size of this lake fluctuated at a tempo near 20,000 years. Several layers of molten magma that intruded into the lakebed sequence and quickly cooled have been dated by radiometric methods. These dates show that the lakebed sequence was deposited over an interval of at least 20 million years centered near 200 million years ago.

This estimate is confirmed by the presence of fine laminations (varves) in parts of the sequence. The varves are tiny (0.2–0.3 mm) couplets of alternating light and dark layers, with one light/dark pair deposited each year. Darker organic-rich layers were deposited in summer and lighter mineral-rich layers in winter. Dissolved oxygen concentrations must have been low or zero in the deeper levels of the lake when the organic-rich layers accumulated to prevent destruction of the delicate varves by animals moving across and in the sediments. Use of these varves as an internal chronometer to count elapsed time (see Chapter 3) confirms that the total time of lake-sediment deposition was about 20 million years.

The types of sediment deposited in the Newark Basin varied widely in response to changes in lake depth. When the lake was deep (100 m or more), the sediments tended to be gray or black muds with large amounts of organic carbon. These sediments contain finely laminated varves and are rich in well-preserved remains of fish. Sediments deposited when the lake was shallower or entirely dried out tend to be red or purple because they were oxidized (rusted) by sporadic contact with air, and they often contain mud cracks due to exposure to dry air. Dinosaur footprints and the remains of plant roots are also common in sediments from the dried-out, vegetated parts of the lakebeds (Figure 9-16).

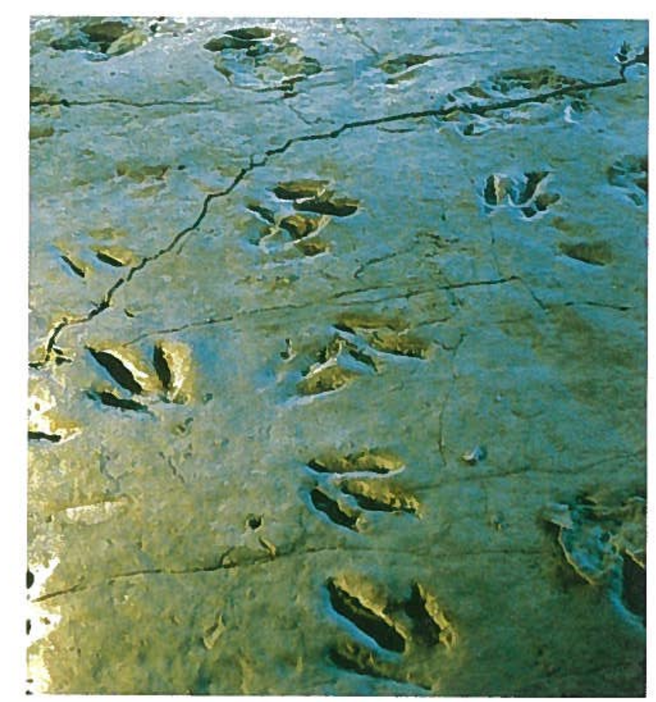


FIGURE 9-16
Evidence of changing lake levels
Dinosaur footprints in lake muds that have since hardened into rock show that the Pangaean lakes occasionally dried out completely. These footprints are from a basin in Connecticut formed at the same time as the Newark Basin in New Jersey. (DINOSAUR STATE PARK, ROCKY HILL, CT.)

The thick sequences preserved in the Newark Basin repeatedly fluctuate between sediments typical of deep lakes and those that indicate a shallowing or complete drying up of the lakes. Individual layers in these sequences are continuous over large areas, indicating that the wet-dry variations in climate affected the entire basin.

Extensive investigations show that these fluctuations in lake depth over millions of years were cyclic (Figure 9-17). The shortest cycles occur over rock thicknesses averaging 4 to 5 meters, equivalent to about 20,000 years in time based on the average thickness of each annual varve (0.2–0.3 mm). These cycles were driven by precession. Monsoons filled and emptied these Pangaean lakes (see Chapter 5) in response to orbital precession in the same way that North African lakes have filled and emptied during much more recent times. Because we are looking much farther back in time, the periods of the orbital cycles were slightly shorter than they are today (see Chapter 8).

Two larger-scale groupings of cycle peaks are also evident. The amplitude of individual 20,000-year peaks in lake depth rises and falls roughly every five or six cycles separated by 20 to 25 meters of sediment, or a little less than 100,000 years. An even larger-scale change in amplitude of the monsoon-cycle peaks occurs between approximately 530 and 620 meters depth in the core (Figure 9-17), equivalent to a time interval of about 400,000 years.

These two longer-term patterns match the expected monsoon signature shown in Figure 9-14 remarkably well. They reflect a modulation of the strength of the 20,000-year precession cycles by eccentricity changes at intervals of about 100,000 and 400,000 years. The full imprint of ancient monsoons is amazingly clear in the sediments of this basin despite the passage of 200 million years.

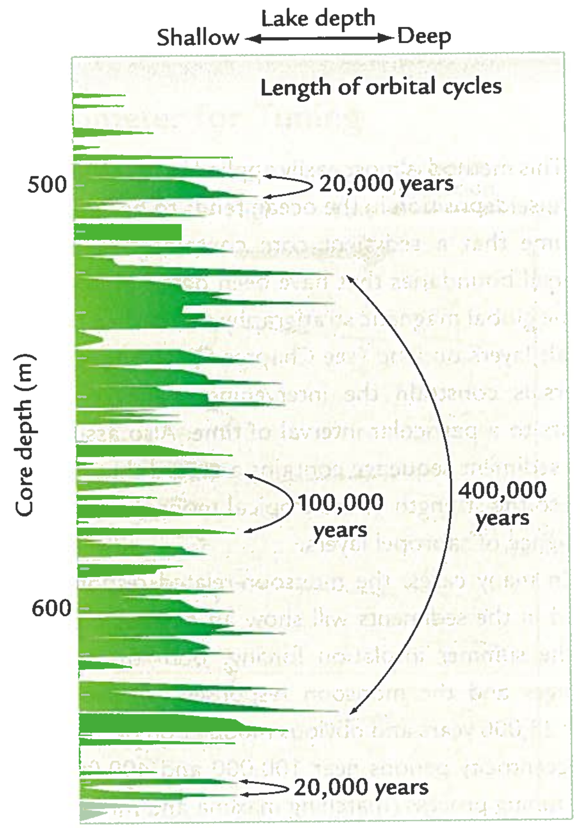


FIGURE 9-17
Fluctuations of Pangaean lakes
Newark Basin lake sediments varied in depth from very shallow to over 100 m deep at three tempos. Individual cycles in lake depth every 4–5 m occur at a period of 20,000 years, clusters of larger deep-lake maxima every 20–25 m occur at intervals near 100,000 years, and unusually large deep-lake clusters at intervals of 90–100 m occur every 400,000 years. (ADAPTED FROM P. E. OLSEN AND D. V. KENT, "MILANKOVITCH CLIMATE FORCING IN THE TROPICS OF PANGAEA DURING THE LATE TRIASSIC," *PALAEOGEOGRAPHY, PALAEOCLIMATOLOGY, PALAEOECOLOGY* 122 [1996]: 1–26.)

9-7 Joint Tectonic and Orbital Control of Monsoons

We saw in Chapter 5 that tectonic changes affect the intensity of monsoonal circulations. Large landmasses such as Pangaea intensify monsoons by offering a larger area for the Sun to heat. Positioning of landmasses at lower latitudes is important because solar radiation is more direct and albedos are much lower than at higher, snow-covered latitudes. Topography is a key control over monsoon strength at tectonic time scales because high-elevation regions focus strong monsoonal rains on their margins.

The processes that control monsoon intensity over tectonic time scales interact with those at orbital scales. Tectonic-scale processes alter the average strength of the monsoon over millions of years, while the orbital-scale insolation changes drive shorter-term monsoon strength at a cycle near 20,000 years. One way the tectonic and orbital factors might interact is suggested in Figure 9-18. On the left is a schematic version of a low-latitude summer insolation curve, with individual maxima and minima at the 20,000-year precessional cycle and modulation of this cycle over intervals of 100,000 and 400,000 years. The smooth curve in the center represents gradually changing tectonic-scale processes, such as the slow uplift that gradually intensifies the average strength of the monsoon over millions of years. This slow tectonic-scale increase in monsoon strength combines with the orbital-scale monsoon cycles to produce the response shown on the right—a *slow increase in the amplitude of the orbital-scale cycles* caused by tectonic amplification.

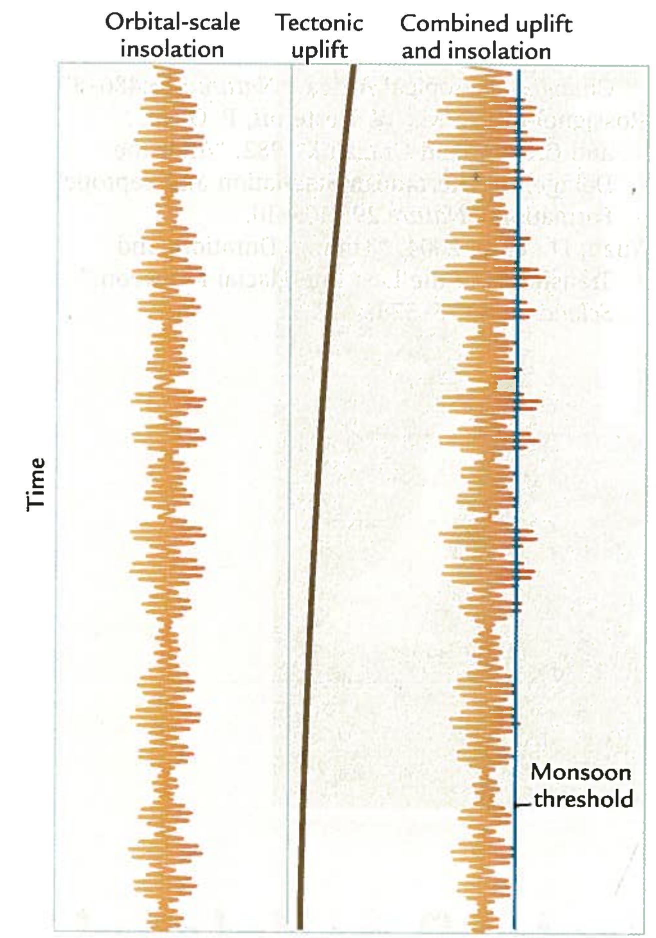


FIGURE 9-18
Combined tectonic and orbital forcing of monsoons
Monsoons are driven by orbital-scale variations in insolation (left) and by slower-acting tectonic factors such as plateau uplift (center). The combined tectonic and orbital forcing causes the amplitude of orbital-scale monsoon responses to increase and gradually exceed critical thresholds (right).

We can hypothesize the existence of a threshold value above which key climatic indices record monsoon responses but below which no response is registered (as in Figure 9-5). In the changes shown on the right in Figure 9-18, the tectonic influence may have been weak enough during the earlier intervals that the orbital-scale monsoon cycles never exceeded this threshold. Later, as tectonic processes created conditions more favorable to monsoons, peaks in summer insolation would have driven monsoons that began to exceed the threshold by small amounts, and then later by steadily increasing amounts.

Something like this kind of evolving climatic response is thought to have occurred in Southeast Asia over the last 30 or 40 million years. A long-term tectonic increase in monsoon intensity due to uplift progressively intensified the amplitude of orbitally driven monsoon cycles in this region. Simulations run with general circulation models indicate that the combined effects of orbital-scale insolation and uplift are not additive in a simple linear way. Instead, it appears that plateau uplift sensitizes the monsoon system to insolation forcing in such a way that the combined monsoon response to uplift and insolation is stronger than a simple linear combination of the two effects.

Key Terms

orbital monsoon hypothesis (p. 179)
nonlinear response (p. 181)
threshold level (p. 181)
sapropels (p. 183)
anoxic (p. 183)
clipped response (p. 189)
orbital tuning (p. 190)

Review Questions

1. In what way is the orbital monsoon hypothesis an extension of processes driving modern monsoons?
2. Why does the intensity of 23,000-year monsoon peaks vary at intervals of 100,000 and 413,000 years?
3. How did the Mediterranean Sea acquire a freshwater lid during times when very little precipitation was falling in that region?
4. Explain how the opposed July/February timing of past monsoon changes in China and Brazil lends strong support to the orbital monsoon hypothesis.
5. Does peak monsoon strength lag behind summer insolation forcing?
6. What similarities exist between monsoon changes in Pangaea 200 million years ago and those in North Africa during the last several hundred thousand years?
7. How do tectonic uplift and orbital variations combine to affect the long-term intensity of monsoons?

Additional Resources

Basic Reading

Kutzbach, J. E. 1981. "Monsoon Climate of the Early Holocene: Climate Experiment with Earth's Orbital Parameters for 9000 Years Ago." *Science* 214:59–61.

Ruddiman, W. F. 2006. "What Is the Timing of the Orbital-Scale Monsoon Changes?" *Quaternary Science Reviews* 25:657–58.

Advanced Reading

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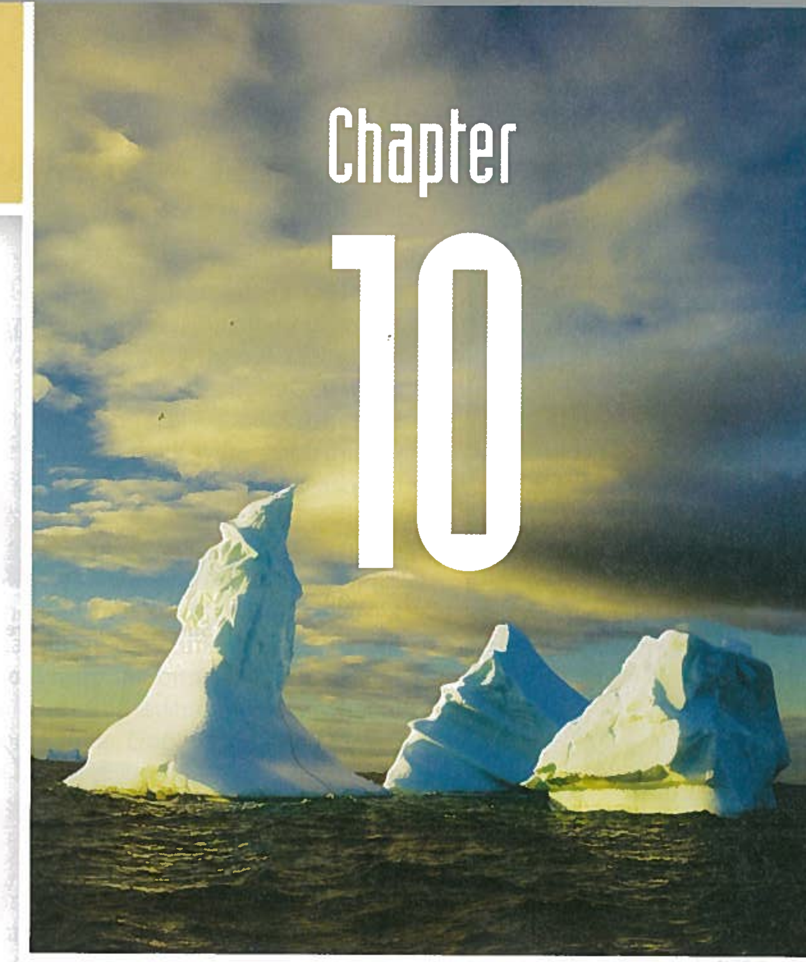
Rossignol-Strick, M., W. Nesteroff, P. Olive, and C. Vergnaud-Grazzini. 1982. "After the Deluge: Mediterranean Stagnation and Sapropel Formation." *Nature* 295:105–10.

Yuan, D., et al. 2004. "Timing, Duration, and Transitions of the Last Interglacial Monsoon." *Science* 304:575–578.



Insolation Control of Ice Sheets

Ice sheets covered northern North America and Europe 20,000 years ago. The present locations of Toronto, New York, Chicago, Seattle, and London were buried under hundreds of meters of ice. Later, the ice melted, and the last remnants disappeared by 6,000 years ago, near the time human civilizations came into existence. The fact that ice sheets first appeared in the Northern Hemisphere in the last 3 million years can be explained by very slow tectonic-scale cooling (see Part II), but the evidence that ice sheets grew and melted over much shorter intervals of time requires a different explanation.

Orbital changes in insolation are the initial driver of these shorter-term variations in the amount of ice. In this chapter, we investigate how changes in summer insolation control the size of ice sheets by determining the rate of ice melting or accumulation. We explore two lags that are important to understanding the ice response: the lag of slow-responding ice sheets behind the insolation changes, and the delayed depression of bedrock beneath the weight of the overlying ice. Both of these lags are thousands of years in length. Then we examine past changes in