A Holocene climatic record from arid northwestern China

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Abstract

The history of climatic changes during the interval 8500–3000 cal. yr B.P. has been reconstructed from stratigraphical and chronological studies and the results of Total Organic Carbon (TOC), Total Inorganic Carbon (TIC), element composition, pollen, and stable isotope analyses of a section along the Hongshui River, in the southern Tengger Desert, NW China. The record suggests that from 8450 (bottom of the studied section) to 7500 yr B.P., the climate was characterized by instability. From 7500 to 5070 yr B.P., the climatic conditions improved and can be divided into two parts: a warm–humid spell between 7290 and 6380 yr B.P., during which the average temperature was 3–4°C higher than that of today, and a warm–dry spell lasting from 5950 to 5720 yr B.P. The climate deteriorated between 6380 and 5950 yr B.P. From 5720 to 5070 yr B.P., the temperature decreased, but humidity increased. An abrupt temperature drop occurred between 5340 and 5290 yr B.P. that indicated the decline of the warmer and humid Mid-Holocene climate. From 5070 yr B.P. onward, the climate oscillated significantly and there were three large temperature decreases coinciding with high mountain glacier advances between 5070–4670 yr B.P., 4300–3740 yr B.P. and 3410–3230 yr B.P. (top of the section), respectively. The climatic fluctuations recorded in the southern Tengger Desert appear in-phase with climatic changes recognized in the Tibetan Plateau, suggesting that the period between 7290 and 6380 yr B.P. was the most warm–humid spell. One extremely dry event occurred at ca. 3000 yr B.P., and subsequently the fluvial–lacustrine depositional process terminated and wind action prevailed in the area; both of these features can be attributed to the rapid strengthening and weakening of the summer monsoon circulation, which are closely connected with global changes. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

In recent years, Holocene climate reconstruction has been the focus of attention of many geoscientists (Markgraf, 1989; Gasse et al., 1991; Shi et al., 1993; Thompson et al., 1995; O'Brien et al., 1995; Fisher et al., 1995; Crowley and Gagan, 1995; Lamb et al., 1995; Pachur and Wünnemann, 1996; Dean et al., 1996; Oppo, 1997; Bond et al., 1997; Chappellaz et al., 1997; Abbott et al., 1997; Beck et al., 1997; Gagan et al., 1998). Though more and more new results have been achieved, a general, internationally acceptable framework is still far beyond reach. The mechanism of climatic change is still unclear and is being debated (Denton and Karlen, 1973; Neje and Johannessen, 1992; O'Brien et al., 1995; Karlen et al., 1998).
One of the main reasons for this is that such studies have largely concentrated on special issues. Nevertheless, in order to understand future climate trends, NW China, as one of the main parts of the global arid belt, is one such arid region, and its palaeoclimatic importance has long been recognized (Flohn, 1968, 1981; Clemens and Prell, 1991; Ruddiman and Prell, 1997). Detailed studies in both the high western Tibetan Plateau and low Northern Xinjiang, western China have shown that the climatic changes are strongly influenced by the summer monsoonal circulation (Kelts et al., 1989; Lister et al., 1991; Gu et al., 1993; Fontes et al., 1993, 1996; Van Campo et al., 1996; Fan et al., 1996; Gasse et al., 1996; Avouac et al., 1996). Furthermore, the South Asian Monsoon is also linked to Southern Hemisphere circulation in the Indian Ocean sector by cross-equatorial wind flow, which is believed to contribute most of the latent heat released in the South Asian Monsoon (Ruddiman and Prell, 1997), and is part of the global changes (Kutzbach, 1981; Kutzbach and Otto-Bliesner, 1982). Long-lasting, wide-ranging field investigations in desert areas (Badanjiling and Tengger Deserts) also indicate that the climate has changed drastically in the past (Pachur et al., 1995; Zhang and Wünnemann, 1997; Ma et al., 1998; Wünnemann et al., 1998). These studies are continuing and will provide more information on climatic changes at both regional and global scales. In this paper, we report the results of a new study from a well-exposed section formed between 8500 and 3000 cal. yr B.P. on the southern margin of the Tengger Desert.

2. Study area and methods

The study area is located at the transition between the Tengger Desert and the Wuwei alluvial–fluvial fan of the Qilian Mountains, northeastern Tibetan Plateau. To the north are vast arid areas such as the Tengger Desert; to the southeast, the Loess Plateau; and in the southwest, high mountain and plateau regions. In addition to the area located to the southeast of the central Eurasian continent, the biggest continent in the world, and in contrast to the biggest ocean, the Pacific Ocean, there exist the largest topographical differences that play an important thermodynamic role in the development of the monsoon. Climatically, the area is situated at the intersection between arid–hyperarid northwest China, the arid–semiarid southeast, and the cold–high mountain–plateau region in the southwest. Coupled with the meandering of the Westerly-Jet, the strong seasonal influence of the East Asian Monsoon reaches the area during the summer, resulting in rainfall from July to September. In the wintertime, cold and dry air masses originate from the Siberian–Mongolian High, and temperatures below zero generally prevail (Zhang and Lin, 1992) (Fig. 1). Therefore, the area is very sensitive to climatic changes and is an ideal locality to study both regional and global climatic changes.

The Hongshui River section is situated at 38°10′46″N, 102°45′53″E (Fig. 2), at an elevation of 1460 m above sea level. With an annual mean temperature of 7.8 °C and an annual mean precipitation of 140 mm, it has a typical arid continental climate. The Hongshui River, flowing from south-east to northwest, is formed by springs emerging from alluvial–fluvial–lacustrine deposits at the lower part of the section. Because of the movement of dunes from the northeast towards the river, and the erosion by the springs, an elongated outcrop has been formed along the river. Regional borehole investigations have confirmed that the strata we have been studying are widely distributed in the area. Because similar sections can also be found in the southeastern Tengger Desert, which belongs to another drainage system, it is therefore a typical Holocene section for the area, and we believe that
Fig. 1. The climatic systems of the study area.
the climatic record of such a section is of regional significance.

The studied Hongshui River section is 620 cm thick and can be traced over a distance of more than 20 km. Overall, five stratigraphical units can be identified (Fig. 2; Fig. 4B). The sampling was carried out once every 10 cm in the lower part (620 to 500 cm), and once every 5 cm from 500 cm to the top. A total of 101 samples was obtained and analysed by inductively coupled plasma (ICP) for elemental composition, and by mass-spectrometer MAT252 for stable isotope composition. Standard methods were used for pollen analyses.

The chronology is based on nine $^{14}$C dates on organic matter analysed in Lanzhou University and Lanzhou Desert Institute, Academia Sinica, China. The ages between the dates are interpolated linearly (Fig. 2; Fig. 4D).

3. Study results

3.1. Chronology

Of the nine samples from the section chosen for $^{14}$C dating, two samples from the lower part
of the section are tree roots near the former surface. We found that these tree roots did not penetrate downwards too much, and stretched horizontally; therefore it likely that the dating represents the actual age of the surrounding deposits. Another two samples in the middle-lower part of the section were taken from peat layers. Because the boundaries of these two compacted peat layers are very clear, they should not have been contaminated. The other five samples from the upper part of the section were taken from layers which were easily distinguished by horizontal bedding, which eliminates the possibility of roots from different layers penetrating them. A half-life of 5568 years was used for the calculations, and the results have been calibrated by the program of Stuiver and Reimer (1993).

The age-depth relationship shows a correlation coefficient of 0.992, which indicates that the dating results are reliable.

### 3.2. Element and environmental indices

$\text{CaCO}_3$ content and its related elements, Ca, Sr and Mg, have been used as the indices of water dynamics, water depth and water properties (acid-base balance), which is consistent with the Fe/Mn ratio, one of the indices of water depth (Zhang, 1997). Total Organic Carbon (TOC) is an index of organic productivity and is supported by the Ca/C ratio, which is related to precipitation and temperature. The content of Mg, which indicates the photosynthetic activity (Gasse et al., 1991), is related to the Sr/Ba ratio. The element Fe is used as an index of Eh-pH conditions (Zhang, 1997). The content of element Al is related to the content and type of clay minerals, and hence weathering intensity (Hakanson and Jansson, 1983; Zhang, 1997). All of these indices are consistent with the stratigraphy, and are summarized in Fig. 3.

### 3.3. Stable isotope composition and temperature

Carbonate in autochthonous lacustrine deposits is formed under isotopic equilibrium and reflects the isotopic composition of the environmental water (Geyh, 1983). Therefore, the $\delta^{18}O$ in authigenic lake carbonate is a proxy of continental climate and depends on the isotopic composition of lake water and water temperature, as well as environmental temperature. Friedman and O’Neil (1997) indicated that the isotopic fractionation between calcite and water varies by $-0.24 \pm 1/\circ C$ of temperature. Isotopic studies on North American lakes show that there exists a strong positive link between $\delta^{18}O$ in atmospheric precipitation and mean annual surface temperature (Yu and Eicher, 1998). These authors took the average coefficient of $0.6\% /\circ C$ (Rozanski et al., 1993) and assumed that the water temperature closely tracks air temperature. Based on this assumption, they estimated a coefficient of $0.36\% /\circ C$ between $\delta^{18}O$ in carbonate and air temperature, which means that, per mille, the $\delta^{18}O$ variation equates to $2.8 \circ C$ air temperature change.

Detailed stable isotopic studies on the Hongshui River section proved that there is no significant correlation between $\delta^{13}C$ and $\delta^{18}O$ that was proposed to indicate hydrological open lakes (Talbot, 1990). On the other hand, $\delta^{13}C$ seem more closely related to the content of $\text{CaCO}_3$. Taking into account the carbonate content in the strata, since there is no carbonate content in the coarse materials, exogenic detrital carbonate can be ruled out. The stable isotopic equilibrium reactions during the formation of the authigenic carbonate are the crucial processes for exploring the relationship between $\delta^{18}O$ and the environmental temperature. In arid areas, $^18O$ precipitates with carbonate preferably under higher temperatures, and results in a more negative value of $\delta^{18}O$. Temperature decrease results in a more positive value of $\delta^{18}O$, and therefore the $\delta^{18}O$ of carbonate in the clay component can be used as a proxy for temperature. However, it should be noted that the relationship between $\delta^{18}O$ and air temperature is a negative one (Asian Monsoonal type), as shown by Yu and Eicher (1998). This result agrees with the results of Gasse et al. (1996) and Fontes et al. (1996). If this temperature-driven $\delta^{18}O$ hypothesis proves to be true, and in a certain temperature range the isotopic changes were entirely the result of temperature differences, we could use it, with caution, to estimate the temperature. It is unclear whether this relationship is applicable to other
Fig. 3. Stratigraphy and geochemical indices for the Hongshui River section.
areas or not, because under different conditions the main factors influencing the isotopic equilibrium reaction may change or become more complex. For example, the conditions in the study area with elevation 1500 m above sea level are very different from those 5000 m above sea level (see also the discussions by Fontes et al., 1996; Gasse et al., 1996). In this study, we estimate that, per mille, the $\delta^{18}O$ change is equivalent to 1.5°C or the average coefficient of $-0.67$‰/°C between $\delta^{18}O$ in carbonate and air temperature. This value is close to the 0.65‰/°C estimate by Geyh, 1983. If we take the equilibrium coefficient between water and precipitated carbonate under 7.8°C (the average annual temperature in the study area) to be 1.0328, based on the average $\delta^{18}O$ values of rainwater (that is $-5.79$‰), then the $\delta^{18}O$ in precipitated carbonate should be $-6.0$‰. This figure is near the average value for the whole section ($-6.07$‰, change between $-3.47$ and $-10.30$‰). Therefore, we can obtain $[-6.0 - (-10.30)] \times 1.5 = 4.30 \times 1.5 = 6.45$ (°C warmer than present, maximum) and $[-6.0 - (-3.47)] \times 1.5 = 2.53 \times 1.5 = 3.8$ (°C cooler than present, minimum). Based on the above considerations, the temperature change history during the Mid-Holocene can be reconstructed (Fig. 4C).

3.4. Pollen analytical results

In addition to the pollen assemblages of samples from the section which have been analysed, the atmospheric pollen samples (1 to 3 m above the ground surface) of one annual cycle from June to September in the following year, and the surface soil samples in the study area, have also been systematically collected for analysis. The pollen-analytical results will be reported in more detail in a separate paper. Here we report only the principal results.

Based on the analyses of palynomorphs and from variations in pollen concentration, the following sporo-pollen assemblage zones can be recognized (Fig. 4A).

Zone 1: 8450–7950 yr B.P. (620–500 cm). The pollen concentration is generally low. The sporo-pollen assemblage is dominated by steppe taxa, notably Artemisia (mean 62.52%; range 27.78–83.33%) and Chenopodaceae (mean 6.92%, range 0–14%), indicating an arid climatic condition.

Zone 2: 7950–7500 yr B.P. (500–390 cm). Pollen concentration is higher than in Zone 1. The percentages of steppe taxa are relatively decreased, but those of wetland taxa, e.g., Typha, increased. Low tree and shrub pollen percentages, together with high percentages of wetland taxa, reflect grassland vegetation, and hence increased moisture and mild warm-humid climatic conditions with clear oscillations.

Zone 3: 7500–6490 yr B.P. (390–280 cm). During this period, pollen concentration is at its highest value and various species are the most abundant in the section (the mean values are Pinus, 8.36%; Ephedra, 5.73%; Picea, 19.18%; Gramineae, 15.27%; Chenopodiaceae, 15.27%). Decreases in the steppe taxa coincide with a slight expansion of the needleleaf trees (Artemisia decreased from 34.10% in Zone 2 to 13.96%; needleleaf trees increased from 18.00% in Zone 2 to 31.09%). Broadleaf trees and shrub taxa indicate a forest landscape. This zone is also characterized by the abundance of wetland and aquatic indicators. These characteristics reflect a forest-grassland under warm-humid climatic conditions, and can be regarded as the Climatic Optimum stage recorded in the Hongshui River section.

Zone 4: 6490–6290 yr B.P. (280–260 cm). Pollen concentration is also relatively high. The content of needleleaf trees, such as Picea, increased considerably (needleleaf trees increased from 31.09% in Zone 3 to 63.50%; Picea increased from 19.18% in Zone 3 to 49.30%), implying a strong temperature decrease which represents an abrupt climatic cooling event in the Mid-Holocene.

Zone 5: 6290–5670 yr B.P. (260–195 cm). Pollen concentration is low. The predominant taxa in this zone are Artemisia (53.16%), Chenopodaceae (10.77%) and traces of wetland taxa. These sporo-pollen assemblages indicate arid steppe vegetation and a warm-dry climate. Here it should be noted that the assemblage of Artemisia and Chenopodaceae is usually used to indicate the humidity/aridity conditions (Gasse et al., 1991, 1996). Based on newly published studies (Song et al., 1996) and our field observations, Artemisia grows in at least two types of climatic conditions.
Fig. 4. Pollen diagram and pollen-climate zones. (A) Pollen zones and their ages. (B) Stratigraphic units. (C) Temperature differences from today based on $\delta^{18}$O. (D) Ages (yr B.P.).
One is a cold–humid condition, where the average surface air temperature in July is 10–17 °C and annual mean precipitation is 300–900 mm. In this situation the percentage changes of Artemisia are very sensitive to temperature and a temperature decrease will cause the representation to increase very quickly. Another type grows in warm semiarid areas where the average temperature in July is 24–28 °C and the annual mean precipitation is 200–350 mm. Here the content of Artemisia is more sensitive to precipitation, and the drier the climate, the higher the content. Where the annual precipitation is 100–800 mm, Chenopodiaceae are more or less sensitive to humidity; the higher the aridity, the higher the content. Thus considerable caution is needed in interpreting pollen analytical data, especially in arid–semiarid areas.

Zone 6: 5670–5010 yr B.P. (195–135 mm). Pollen concentration is low. The percentage of needleleaf trees, which are mainly Pinus (35.17%, in Zone 5 5.85%) and Picea (17.33%, in Zone 5 9.08%), is greatly increased. Generally, the vegetation consisted of needleleaf trees, indicating a drop in temperature compared with Zone 5. The climate was cold and humid.

Zone 7: 5010–4470 yr B.P. (135–100 cm). Pollen concentration is low. Taxa indicative of dry grassland (Artemisia 25.55%; Chenopodiaceae 20.14%) and needleleaf tree taxa (Pinus 29.14%; Picea 14.86%) dominated alternatively, indicating large dry–cold variations in climatic conditions.

Zone 8: 4470–3510 yr B.P. (100–30 cm). Pollen concentration is low. The pollen assemblages indicate that needleleaf trees (68.07%) dominated the vegetation, indicating a cold–humid climate.

Zone 9: 3510–3230 yr B.P. (30–0 cm). In this zone, wetland taxa, such as Typha, increased greatly (the percentage of Typha and Potamogeton increased from 3.55% in Zone 8 to 67.89%). This indicates a rise in the precipitation/evaporation ratio, and hence a relatively humid climatic stage.

It is also worth mentioning that discrepancies exist between climate reconstruction deduced from pollen and the temperature reconstruction based on stable isotope studies, in which the pollen components are influenced by the geographical background and are susceptible to both temperature variations and the moisture available locally.

The vegetation evolution is more sensitive to changes in the precipitation in the area.

4. Holocene palaeoclimate reconstruction in the southern Tengger Desert

The reconstructed history of climatic change (Fig. 4) shows that from 8450 (age of the bottom of the studied section) to 7500 yr B.P. the climate oscillated around an unfavourable background and can be divided into two intervals. Before 7950 yr B.P., the lithology of the section consists of alluvial–fluvial siliciclastic material, and the grain size is coarse. The contents of CaCO$_3$, TOC and the elements Ca, Sr and Mg are very low; Al and Fe are only 10 and 15 mg/g, respectively, with very little variation between samples. This interval coincides with pollen Zone 1, which is dominated by steppe taxa, notably Artemisia and Chenopodiaceae, indicating arid climatic conditions. From 7950 to 7500 yr B.P., the lithology is composed of alluvial to shallow water deposits; the CaCO$_3$ and Ca and Sr contents of several samples increased, but generally are near zero. This interval corresponds to pollen Zone 2, in which the content of steppe taxa is decreased, but the wetland taxa, e.g. Typha, increased, together with the mixed needleleaf trees, broadleaf trees and shrubs, indicating a climate condition characterized by increased humidity and strong temperature fluctuations.

From 7500 to 5070 yr B.P., the climate changed to warmer and more humid than before. The strata include the upper part of units 2, 3 and part of unit 4, composed of two peat–lacustrine layers topped by silt to wind transported sand, changing into typical lake deposits again. Fig. 3 shows that from the sand layer downwards, the contents of CaCO$_3$, elements Ca, Sr and Mg are all very low, except in a few samples. However, the contents of Al, Fe and TOC reach their maximum in the section, at 48, 58 and 15.4 mg/g, respectively. The value of $\delta^{18}$O also reaches its lowest, −10.3‰. This period corresponds to pollen Zones 3, 4, 5 and 6, which show a clear trend of climatic and environmental change. It can be divided into the following stages: from 7500 to 7290 yr B.P., the
temperature increased considerably until the first high temperature period occurred. A climate optimum occurred during 7290 to 6380 yr B.P. Based on $\delta^{18}O$ values, it can be estimated that the average temperature was 3–4°C higher and the maximum temperature was 4–6.5°C higher than present (Fig. 4C). This is 1–2°C higher than the previous estimate (Shi et al., 1993). From 6380 to 5950 yr B.P., the lithology is composed of clay to clay-silt, indicating a shallow lake or swamp environment. The content of Al, Fe and TOC decreased and the $\delta^{18}O$ value increased with some variations, indicating that the temperature during this time decreased with three cold stages centred at 6260, 6170 and 5980 yr B.P. At the same time, it became drier than before. Of these three cold stages, the earliest one corresponds to pollen Zone 4, which is characterized by an increase in needleleaf trees, especially *Picea*, representing an abrupt fall in temperature which may correlate with a worldwide Mid-Holocene temperature decrease. From 5950 to 5720 yr B.P. was the second warmest stage. The stratum is composed of fine wind transported sand deposited in shallow water, which indicates a southward expansion of mobile dunes. This represents the geological evidence of climatic deterioration. Low $\delta^{18}O$ values show it was an interval with high temperatures. The pollen analyses indicate that the vegetation was mainly composed of *Artemisia* and *Chenopodiaceae*, which can also grow under a warm–dry climatic condition. From 5720 to 5070 yr B.P., the palaeolake developed fully. The content of TOC increased and the $\delta^{18}O$ values were higher than those of the second warmest stage, implying that the temperature had decreased. The pollen assemblage of mainly *Pinus* and *Picea* indicated a cold–humid climate. During this period, there occurred a strong temperature decrease from 5340 to 5290 yr B.P., which is clearly demonstrated by a layer of silt–sand that can easily be distinguished in the section. The stable isotope composition changed sharply. It was estimated that the temperature dropped by almost 4°C in about 20 years preceding the deterioration of the Mid-Holocene warm–humid climate.

From 5070 to 3230 yr B.P. (top of the section), the climate deteriorated further. This period coincides with pollen Zones 7, 8 and 9, and experienced dry–cold, cold–humid and humid climatic shifts. These can be demonstrated by three long-lasting strong temperature decreases occurring between 5070 and 4670 yr B.P., 4300 and 3740 yr B.P., and 3410 and 3230 yr B.P., and can be correlated with the mountain glacier advances on the Tibetan Plateau and in the Qilian Mountains (Chen, 1987; Zheng, 1990). At ca. 3000 yr B.P., which is as an important datum, the climate in the studied area deteriorated severely, desertification processes prevailed, sand dunes intruded and erosion occurred. During the last ca. 3000 years, the climate may have improved at times, for example, around 2000 yr B.P., but it never returned to the previous favourable conditions.

5. Discussion and conclusions

One of the most significant findings of this study is the Holocene warm climate that prevailed between 7500 and 5070 yr B.P. The most intense warm–humid climate occurred at 7290 to 6380 yr B.P. These records are in agreement with results from lacustrine deposits (Gasse et al., 1991, 1996) and both the Dunde Ice Cap (Thompson et al., 1989) and the Guliya Ice Cap (Yao et al., 1997) on the Tibetan Plateau. Between 6380 and 5950 yr B.P., the climate became drier and the temperature decreased, the lowest temperatures occurring at 5660, 5560 and 5380 yr B.P. These three low temperature events can be correlated with the Dunde ice core records (Thompson et al., 1989). The earliest event, which coincides with pollen Zone 4, can be correlated with a dramatic temperature decrease in north and northwest China (Shi et al., 1993). It can also be correlated with the glacier advances during 5700 and 5350 yr B.P. in the Tianshan Mountains in west China (Zheng, 1990), and coincides with the second New glacial event at ca. 5800 yr B.P. described by Denton and Karlen (1973). Overall, this was a period of climatic deterioration punctuated by several large-amplitude climatic events. Between 5950 and 5720 yr B.P., pollen analyses indicate that *Artemisia* and *Chenopodiaceae* were the most abundant taxa, which have been proved to represent a warm–dry climate condition based on the present
Stratigraphic and geochemical evidence also supports this conclusion. From 5720 to 5070 yr B.P., although the temperature was lower than before, the humidity increased. The short but strong temperature decrease from 5340 to 5290 yr B.P. indicates the decline of the warm Mid-Holocene. Afterwards, the temperature and humidity increased from 4670 to 4500 yr B.P. and formed a mild warm-humid spell. Later, the temperature decreased again between 4300 and 3740 yr B.P., which might be correlated to the interval of glacier advance during 4100–3950 yr B.P. on nearby high mountains, e.g. the enlargement of the Guliya Ice Cap at 3980 ± 120 yr B.P. (Shi et al., 1993). Although the temperature increased somewhat after 3740 yr B.P., it decreased considerably at 3410 yr B.P., and formed another cold spell that might be correlated to the mountain glacier advances between 3100 and 2500 yr B.P. (Zheng, 1990; Shi et al., 1993).

The history of Mid-Holocene climatic change recorded in this arid continental part of NW China generally agrees with the model simulation results by COHMAP members (1988), but differs from the results of the Greenland ice core, which indicate the occurrence of high-frequency, low-magnitude temperature oscillations. In the Greenland ice core, only the physical structure shows a 'brittle zone' in the depth interval between 800 and 1300 m, which corresponds to about 8000 to 4000 yr B.P. and indicates a remarkably stable climate during the past 10 ky (Dansgaard et al., 1993). In this respect, the Holocene climate, at least in the high-latitude North Atlantic region, has been regarded as anomalously stable. Thus it seems that the climatic change processes are very different between mid–low latitudes and high latitudes during the Holocene period. Nevertheless, a mechanism linkage is necessary to explain the phenomena.

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