breeding system is required for *J. curcas* to occupy different habitats and build up its own populations.

The shallow, light and simple flowers, providing a platform are shown to be characteristic of fly-pollinated flowers[12]. These characters are found in *J. curcas*. The flowers provide easy access to the foragers. They produce copious amount of pollen at inflorescence level. The pollen placed in two tiers of stamens, advertises its existence. Further, nectar secreted in traces at the shallow flower base glitters against sunlight and advertises its existence to the foragers. Usually, nectar is concealed and does not advertise its existence[13].

The individual flowers are grouped together in the racemose inflorescences, an arrangement which promotes attraction and foraging rate by the foragers. The plant with monoecious sexual system essentially requires an agent for pollen transfer from male to female flowers, within or between conspecific plants. The floral rewards in both flower sexes are accessible even to short-tongued anthophilous insects. The bees, by collecting pollen and nectar and by moving between male and female flowers within and on different conspecific plants, effect pollination in a messy and soiled manner[12]. The flies also exhibit the same foraging behaviour and effect pollination. Ants and thrips remain on the same plant and effect geitonogamy only. Although all insect species effect pollination, only bees and flies effect xenogamy. The flies are represented by only one species, *Chrysomya*, and they are under-represented among the pollinators. They generally utilize many different sources of food, and usually their pollinator activity is unreliable. The proximity of a suitable breeding ground, frequent wet, decaying vegetable and dung material is important for their presence in the vicinity of *J. curcas*. In the study areas the decaying, wet material or dung or fallen decaying fruits are not found. Hence the absence of flies, especially small-bodied *Musca*, *Eristalis*, etc., which have small-distance flight range, is not surprising. The introduction of breeding material in the vicinity of *J. curcas* allows the flies to utilize the same for breeding and the flowers for food, while effecting pollination.

The natural fruit set rate indicates that the plant does not suffer seriously from under-pollination. The production of female flowers in small number, surrounded by a large number of male flowers in *J. curcas* seems to be a strategy to ensure pollination to the maximum extent. The stigma receptivity lasting three days also additionally provides opportunities for pollination, if not pollinated on the first and second day. The study indicates that pollen is deposited in sufficient amount, which is visible by its yellow colour even to the naked eye. However, the plant with predominant xenogamy requires mostly xenogamous pollen for more fruit set, after selective elimination of growing fruit. Therefore, pollen transfer between conspecifics has a great bearing on the net percentage of natural fruit set.

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Onset of climate change at Last Glacial–Holocene transition: Role of the tropical Pacific

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We study palaeoclimatic records from various sites spread around the earth, focusing on the start of the last glacial–interglacial transition. The warming, as recorded in the $d^{18}O$ record started first in the tropics, then propagated to the Antarctic and then finally to the Arctic. Our analysis of the data suggests that it took about 7.6 ka for onset of climate change to propagate globally. We propose that the tropical Pacific played a major role in initiating the warming in the tropics. We discuss mechanisms that could have transported this heat from the tropics to Antarctica and then to the Arctic during transition to the interglacial.

NUMEROUS polar and tropical ice cores, marine cores and continental records indicate that the climate has changed significantly over the past hundreds of kiloyears. Many

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previous investigations focused on climatic events like the Dansgaard-Oeschger[1–3] and the Younger Dryas[4,5]. These events, though representing violent changes in the climate, are characterized by the fact that once the events terminate, the climate returns to approximately the same state it was in prior to their onset. This is clear from an inspection of the available $d^{18}O$ records and other proxy records. In contrast, we concentrate on the transition from one distinct stable phase of the climate (the glacial phase) to another stable phase (Holocene). This transition is as important (if not more important) than the other events, since it has led to a more 'permanent' change in the climate. We analyse the leads and lags for the onset of the transition phase at various sites around the world, and propose possible mechanisms to explain them. Such studies can give us important clues to the future behaviour of the climate. In particular, it may provide answers to questions such as: Can anthropogenic effects lead to another such transition to an even warmer climatic phase? Has the next transition phase already commenced?

We start by defining the onset of transition from the various ice-core data. We consider $d^{18}O$ records from the following sites: Guliya ice-core [35°17′ N, 81°29′ E, Tibetan Plateau; 6200 m asl][6], Sajama, Bolivia [18°07′ S, 68°53′ W; tropical Atlantic ice-core; 6542 m asl][7], GRIP [72°34′ N, 37°37′ W; 3232 m asl][8] and GISP2 [72°6′ N, 38°5′ W; 3200 m asl][9] both in Greenland, and deuterium record from Vostok [78°28′ S, 106°48′ E, East Antarctica; 3488 m asl][10] (Figure 1). We clearly see one stable regime during the glacial phase and another stable regime during Holocene. We are interested in the onset time of the transition phase between these two climatic regimes. We define this to be the lowest point in the time series during the glacial phase from where there is a consistent increasing trend leading to the Younger Dryas event. From Figure 1 we observe that the onset of transition in the $d^{18}O$ and deuterium records varies according to the region. For example, the onset time for deglaciation is 18.8 ka BP at the Guliya ice-core in the tropics, whereas it is 16.2 ka BP at Vostok, Antarctica. However, there are uncertainties associated with age models in Antarctica and hence the above lag could be smaller. Further, the onset time at tropical Atlantic, Sajama ice-core is recorded at 15.7 ka BP. Finally, the onset times in the Arctic (Greenland ice-cores) lag behind even further at 14.7 ka BP (for GRIP) and 14.5 ka BP (for GISP2)[5].

From the above observations, we see that the warming (as seen in the $d^{18}O$ and deuterium records, which are proxy for the local temperatures) started first in the tropics, then propagated to the Antarctic and finally to the Arctic. We now postulate possible mechanisms that can account for this sequence of events. Towards this end, we look for evidence of onset times for climate change in records from representative sites with well-established age models, belonging to different geographic regions. Different proxies like greenhouse gases, grain size, pol-

len, $d^{13}C$ and $d^{18}O$ have been used. In this communication we use available CH_4 data as representative of greenhouse gases. We begin our analysis with the warming in the tropics. Tropical ice, marine and terrestrial climatological records imply that each of the three major oceans – the Pacific, the Indian and the Atlantic oceans – has different influences on the climate. But these influences are also interrelated, as we shall see below.

We first explore the role played by the tropical Pacific[11,12]. We start by looking for early signatures of climate change in the region influenced by the Pacific. In the central part of Loess plateau, China, the loess–palaeosol sequence at Luochuan [35°45′ N, 109°25′ E][13] shows that the Quartz Median Diameter (QMD) conti-
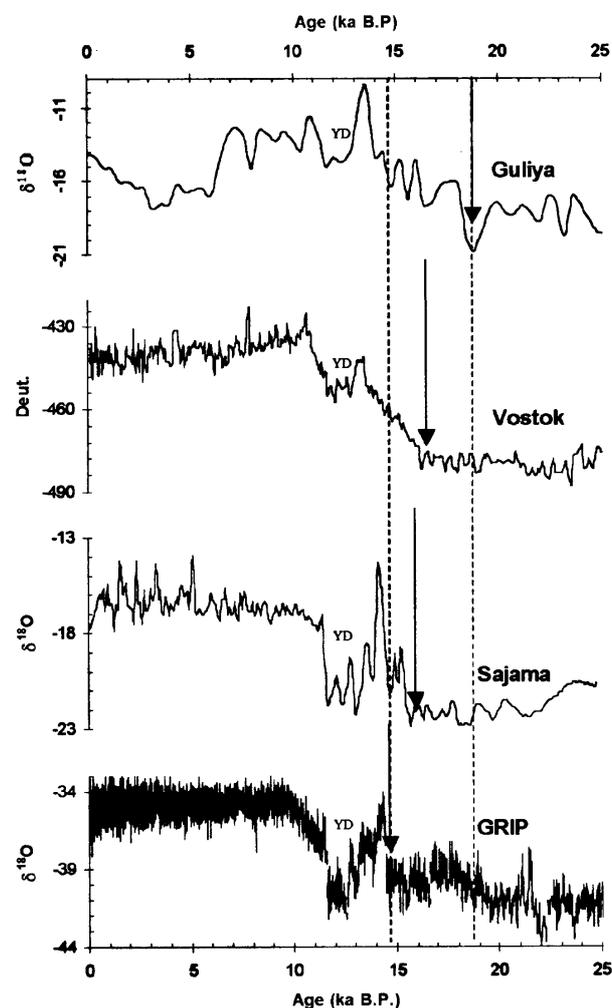

Figure 1. $d^{18}O$ from ice cores at Guliya, Himalayas[6]; Vostok, Antarctica[10]; Sajama, South American Andes[7]; GRIP, Greenland[8]. It can be seen that change in temperature occurred first in Guliya and last in GRIP. Arrow indicates onset time. The time duration of deglaciation (from onset time of transition to peak warming) is maximum at Guliya (~ 5.5 ka) followed by Vostok (~ 3.5 ka), Sajama (~ 1.5 ka) and GRIP (less than 0.5 ka). This difference in time duration is amplified by the fact that the peak warming occurs first at GRIP, followed by Sajama, Vostok and Guliya. We also note that the Younger Dryas starts only after the above climate change has taken place all over the globe. Further, this onset time is approximately the same for all sites.

nuously increased during the glacial phase. At 21.3 ka BP, QMD falls sharply marking the beginning of climate change (Figure 2). QMD is a proxy for wind strength and the decline in QMD suggests atmospheric change. Since this change predates even the onset of transition observed in the Guliya ice-core, it therefore forms the earliest record of the climate change from glacial to intraglacial. However magnetic susceptibility, a proxy of pedogenesis, responds with a delay of 7.6 ka and starts increasing only at 14.7 ka BP, suggesting that the early response of QMD is due to the direct influence of the atmosphere, whereas the delayed response of magnetic susceptibility is due to consequences that followed this atmospheric change. The delay suggests it took 7.6 ka for the initial climate change to influence the entire regional climatic system and get recorded in the continental sediment. A similar early influence of the atmosphere is also recorded in the largest lake in Japan located in the central part of the Honshu Island – Lake Biwa [35°15′ N, 136°5′ E][13,14], and further south at the 17940-1/2 marine core [20°07′ N, 117°23′ E] located in the South China Sea[15]. Eolian Quartz Flux (EQF) at Lake Biwa and Median Grain Size silt (MGS) at 17940-1/2 marine core record the change in wind patterns at 20.4 ka BP and 20 ka BP respectively. The EQF and MGS have their provenance in the Chinese loess plateau. This shows that the change has propagated from Chinese loess plateau to South China Sea by 20 ka BP.

The above early signature of climate change can be explained through the dynamics that originates in the Pacific Ocean. It is now recognized that several energy flows that affect the global climate come to a confluence in the western tropical Pacific[11]. For example, the orbital forcing and the consequent changes in insolation patterns[16] would have triggered the warming of tropical Pacific. This warming up of the Pacific generates a jet stream over the Chinese subcontinent. This jet stream would have then blown the sediments from the western desert and deposited them in the plains in China, leading to the change observed in Luochan at 21.3 ka BP. The jet stream would have further transported these sediments further east, as recorded in Lake Biwa and South China Sea.
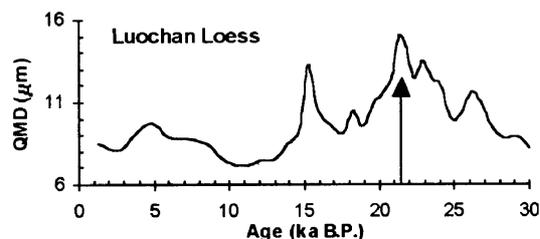
The warming of the tropical Pacific would have led to an increase in production of water vapour (a significant greenhouse gas) over the tropics[17]. As is well known[16], the water vapour feedback is the most important atmospheric feedback mechanism amplifying responses to climate forcings. This feedback would have led to continued warming of the tropical Pacific. At around 19 ka BP, a sharp increase in CH_4 concentration is observed[18,19] in polar ice cores (Figure 3). This can be explained as follows. The change in tropical Pacific temperature probably reached a critical level at this juncture, triggering the release of CH_4 from methane hydrates deposited in marine sediments during the glacial phase[20]. The near simultaneous change in CH_4 concentration observed in both Antarctic and Arctic is to be expected, since the atmospheric CH_4 content homogenizes across the entire globe on the time-scale of a few months because of atmospheric redistribution. The greenhouse gases feedback cycle proposed above has led to a warming trend recorded in the Tibetan plateau (at the Guliya ice-core) starting from 18.8 ka BP. The absence of high concentrations of water vapour in the polar regions would have prevented this feedback cycle from starting there.

Next, we study the onset of deglaciation through the $d^{18}O$ record and other related proxies in the palaeoclimatic marine records that were influenced by the Indian Ocean. The deglaciation observed in the Guliya core has been shown[21] to affect the Arabian Sea marine core 74KL [14°19′16″ N; 57°20′49″ E; 3212 m asl][22]. The increase in temperature in the Tibetan plateau starting at 18.8 ka BP led to large-scale melting of ice sheets after a time lag and the consequent large-scale influx of water into the Arabian Sea affected the marine core
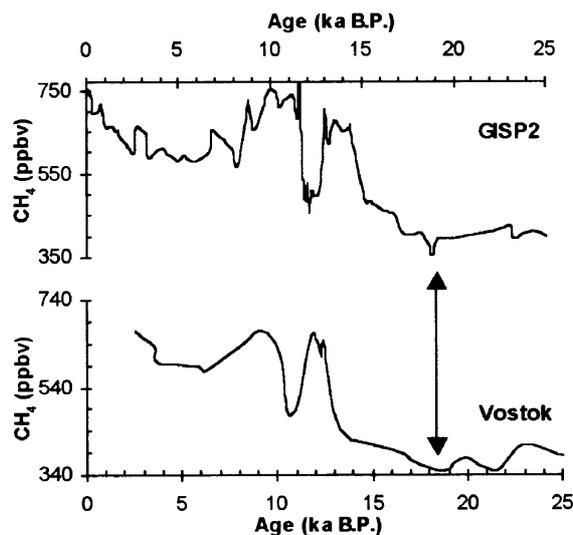

Figure 2. Quartz mean diameter (QMD) from Loess–Palaeosol sequence at Luochuan, Central China[13]. Arrow indicates onset time. QMD is a proxy to the wind strength. Change in QMD at 21.3 ka BP suggests atmospheric changes, forming the earliest record of climate change.

Figure 3. CH_4 records from Northern and Southern Hemisphere (GISP2, Greenland[18] and Vostok, Antarctica[19] ice cores). Arrow indicates onset time. Change in CH_4 concentration occurs simultaneously in both Antarctic and Arctic, since the atmospheric CH_4 content rapidly homogenizes across the entire globe.

74KL. Therefore, the onset of transition in the temperature proxy of ocean water, $d^{18}O$, and the proxy for melt water influx, Total Rare Earth elements (TREE), are seen only at 17.5 ka BP. We see a similar delayed response in the $d^{18}O$ records from other marine cores. At RC-27-23 [18° N, 57°59′ E; 815 m], the change is recorded at 14.15 ± 160 ka BP[23]; at GC-5 [10°23′ N, 75°34′ E; 280 m asl] NE of Cochin on the continental shelf, it is at 15.2 ka BP[24].

We now reconstruct the onset of climate change on the neighbouring continent using various proxy records and field observations. Various continental records from higher elevations such as Kashmir loess show climate change at 18 ka BP based on development of palaeosol, pollen, C/N ratio, heavier values of $d^{13}C$, contribution of C3 plant and well-developed Upper Palaeolithic archaeological sites in Kashmir[25]. Records from the northern Sahara desert and Kalahari desert suggest that dunes got stabilized at around 17.7 ka BP and 17.3 ka BP respectively[26]. Vegetation and pollen response to the climate change at the southern tip of the Indian subcontinent, Nilgiri [11°–11°30′ N, 76°20′ E; 1800–2500 m asl][27] and Madagascar [Crater lake, Tritrivakely: 19°47′ S; 46°55′ E; 1778 m][28], are observed at 16.5 ka BP and 17 ka BP respectively. In Burundi, Central Africa [2°28′ to 3°42′ S; 1850–2240 m asl] eight pollen sequences, recorded as differences between reconstructed absolute values and modern precipitation values and calculated at each site by the interpolation method based on meteorological data, show climate change beginning at 13.5 ka BP[29]. We conjecture that the onset times depend on the processes influenced by the SW Indian monsoon that affects this region, and hence local effects play an important role. However, we can make a general observation that the onset at higher elevations lags behind the onset observed in Guliya ice-core and the lag appears to increase as we get closer to the equator. Records from the plains of western India show even larger delays in response to climate change. In western Indian plains of Rajasthan – Didwana Lake [27°20′ N, 74°35′ E], the response was delayed by 5.8 ka. At Didwana, the influx of clastic sediments ends the hypersaline condition in the lake, and the rise in wetland pollen taxa at 13 ka BP[30] marks the response of the plains to deglaciation that triggered climate change at 18.8 ka BP in Guliya.

Having investigated the onset of the transition phase in the regions influenced by tropical Pacific and Indian oceans, we now explore possible teleconnections between these regions, and the Antarctic and Arctic. As observed earlier, the onset of transition phase from glacial to inter glacial started in Antarctica, followed by tropical Atlantic and finally the Arctic (Figure 1). This is also supported by the marine core data from northern Atlantic-MD952042 [37°47′99″ N, 10°09′99″ W; 3146 m asl][31] which show the onset of transition at 15 ka BP. The above sequence of warming can be explained as follows.

Heat from the tropical Pacific is transported to both hemispheres but with a significant bias towards the south, reflecting the need to supply the South Atlantic through the Antarctic circumpolar current (ACC)[11,12]. This mechanism causes the observed initial change in Antarctica at around 16 ka BP. Subsequently, the well-established great 'conveyor belt' circulation of the Atlantic Ocean, which carries warm surface-water north and cold, dense, deep-water south[32], would have led to a warming of the Andes ice-cores followed by the Greenland ice cores. According to us, during the last glacial–Holocene transition it is the initial warming of North Atlantic Ocean that influenced the higher latitude polar region and not the melt water from the Arctic. The above mechanism establishes an intimate link between tropical Pacific, ACC, Antarctica and the Arctic.

We now highlight an important difference between the climate change represented by the last glacial–Holocene transition and the climatic event represented by the Younger Dryas. A characteristic feature of the transition is that the change is first observed in CH_4 records at the ice cores, whereas the change in $d^{18}O$ follows later with a lag. This is in contrast to what happens during events like Younger Dryas, where the signature is sharp and climate rebounds back to the original state. During such events, the changes in CH_4 and $d^{18}O$ records are synchronous in all the ice cores (Figures 1 and 3). Therefore, during Younger Dryas the melt-water mechanism would explain the proxy records. This can be seen as follows. During the last glacial–Holocene transition, the climate change was driven by changes in orbital forcing and the consequent large increases in concentration of greenhouse gases (CH_4, CO_2 and water vapour). Even though the CH_4 concentration equalizes across the globe rapidly, the rise in temperature would be locally controlled by various factors (like water vapour and CO_2 concentration), and therefore shows a lag that may range from a few years to a few kiloyears. In the case of Younger Dryas, it was large-scale deglaciation which drove the climate change. This deglaciation brought large volumes of cold water to the oceans, whereby the temperature dropped suddenly. We observe that temperature and CH_4 are directly proportional, and so the sudden drop in temperature resulted in a sudden drop in CH_4.

To summarize our results, we observe that the tropical Pacific had played an initial and vital role in triggering a climate change at around 21.5 ka BP. Its first fingerprints were recorded by the QMD, a proxy of wind strength over the loess plateau of China. The above climate change then initiated a warming in the tropical Pacific region. Further, a tropical Pacific current was established, which transported the heat to the ACC and thus influenced the Antarctic region at around 16–17 ka BP. The ACC then affected the Atlantic Ocean which, through the thermohaline currents, warmed the tropical Atlantic and then finally the North Atlantic polar regions.

We conclude with possible implications of the above observations for future climate change. A characteristic feature of the transition is that the change as first observed in CH_4 (one of the greenhouse gases) records at the ice cores. The change in $d^{18}O$ follows later with a lag that depends on the region. This is in contrast to what happens during events like Younger Dryas, where the climate rebounds back to the original state. During such events, changes in CH_4 and $d^{18}O$ records are synchronous in all the ice cores. If we now study the palaeoclimatic records during the late Holocene period, we observe a situation similar to that prevailing prior to the glacial–Holocene transition. The CH_4 records from polar ice cores show a significant increase starting at around 3 ka BP, whereas there is no corresponding increase (yet) in the $d^{18}O$ records from these ice cores. The only exception is the $d^{18}O$ record at Guliya, which indicates the commencement of a warming trend starting at around 3 ka BP[6]. This is again similar to the situation during the transition, where the Guliya core records the change first. Therefore, are we seeing the beginning of another transition to an even warmer climate? Obviously, we need more data before we can answer such questions with confidence.

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