The Loess/Paleosol Record and the Nature of the Younger Dryas Climate in Central China

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The use of latest Pleistocene-Holocene paleosols in defining Chinese climatic sequences is plagued by poor chronological controls caused primarily by the use of radiocarbon dates derived from bulk soil carbon. Dating of a post-glacial aeolian/paleosol sequence in the Pigeon Mountain basin of north-central China, using culturally deposited charcoal, support a wide array of other data suggesting the Younger Dryas was a period of cooler dryer conditions marked by wide-spread aeolian deposition. Periods of soil formation and higher lake levels bracket this climatic event. Climatic variability immediately before, during and immediately after the Younger Dryas interval is associated with rapid technological elaboration and innovation in the production and use of chipped stone tools, and perhaps, ground stone. © 1998 John Wiley & Sons, Inc.

INTRODUCTION

Chinese loess/paleosol sequences are frequently used as primary proxies for Late Quaternary climatic changes in the central Asian land mass (An et al., 1990, 1991; Forman, 1991; Heller and Liu, 1986; Liu, 1985; Liu et al., 1992; Maher and Thompson,

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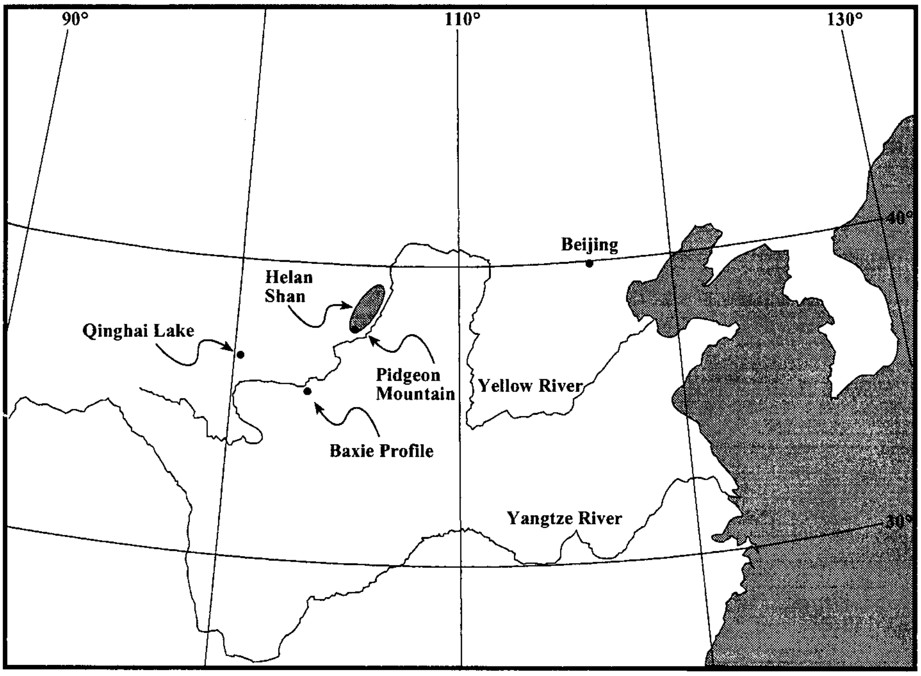

Figure 1. Location of primary loess/paleosol and lake-level sequences discussed in text.

1995; Rutter and Ding, 1993; Sun and Zhao, 1991; Winkler and Wang, 1993; Zhang et al., 1994). With very few exceptions, the paleosol horizons are interpreted as indicating warmer and more humid phases, while the less-weathered loess layers are consistently thought to represent cooler and drier phases (Maher and Thompson, 1995:383). The Baxie Paleosol (Figure 1), was once thought to have formed in northwestern China during the Younger Dryas under cool but wet climatic conditions (An et al., 1993), but a broad array of evidence now suggests Younger Dryas climates in central China were primarily cold and dry, consistent with other periods of loess deposition, but were variable, with a brief interval of warmer/wetter conditions (Zhou et al., 1996).

We are interested in the influence of this climatic variability on hunting and gathering precursors of the Neolithic in northern China. This period is poorly known, but of increasing concern to scholars (Underhill, 1997). Available evidence suggests a technological transition beginning in the Late Paleolithic, in which discoidal cores, flake points, blade tools, backed knives, and burins of the early Late Paleolithic (Shuidonggou: 26.2–17.2 ka) are supplanted by a more diverse array of flake and blade tools, developed unifacial and bifacial tools, microliths, and, perhaps, milling stones and partially ground celts in the latest Paleolithic (Xiachuan:

23.9–13.9 ka, Xueguan: 13.6 ka, and Hutouliang: 11.0 ka) (Chen, 1984; Zhou and Hu, 1985; Ningxia Museum and Ningxia Bureau of Geology, 1987; Chen and Wang, 1989; Yamanaka 1993). An early association of grinding stones, pestles, and red and gray ceramics occur in layer 6 of the Nanzhuangtou site, Hebei, dated at 10,815 ± 140 ^{14}C yr B.P. (BK87088) (Jin and Xu, 1992; Ren, 1995; Underhill, 1997: 114).

In 1990 the discovery of assemblages containing large cores, flake and blade tools, microliths, chipped and ground stone celts, and red and gray ceramics in the vicinity of (Pigeon Mountain) indicated the possibility of an important record of culture change in the Late Paleolithic (Wang and Yu, 1996). Subsequent survey in 1995/1996, conducted as part of an on-going study focused on the Helan Shan (Helan Mountains) and adjoining desert areas of western Inner Mongolia and the Ningxia Hui Autonomous Region (Bettinger et al., 1994; Madsen et al., 1996; Elston et al., 1997), recorded ten archaeological sites in the Pigeon Mountain basin (Figure 2). The Pigeon Mountain cluster includes QG3 (QG is the county designator within Ningxia), south of Pigeon Mountain, as well as QG4, QG5, QG6, and QG13. Two other sites, QG7 and QG8, are associated with a spring cluster about a kilometer to the south. Sites as yet unnumbered are associated with two spring mounds north of Pigeon Mountain, and another spring complex just south of QG7 and QG8. Most concentrations of surface artifacts in the Pigeon Mountain basin are either in, or directly associated with, springs and spring mounds, or in seasonal washes immediately downstream. Artifacts are extremely scarce in the braided channels, sand sheets, and gobi pavements away from water. Herein, we report recent dating of archaeological materials within this depositional sequence which generally supports the interpretation outlined above. Periods of warmer/wetter conditions, fostering soil formation, bracket a cool, dry Younger Dryas characterized by aeolian deposition.

DATING CHINESE PALEOSOLS

The dating of paleosols in the Loess Plateau region of north-central China is fraught with difficulty, particularly for the period of the latest Pleistocene and Holocene. This is related to a general problem associated with dating soils everywhere; the accurate dating of soil carbon is notoriously complex due to a variety of potential contamination vectors (Chichagova and Cherkinsky, 1993; Evans, 1985; Liu, 1985; Scharpenseel, 1971; Scharpenseel and Schiffmann, 1977). Generally, "in soil ^{14}C dating, 'soil age' is definable and applicable for absolute dating in a limited way, only in the case of well-protected paleosols, charcoal or wood relics" (Scharpenseel and Becker-Heidmann, 1992:548). Even then, "interpretation of ^{14}C dates of soil carbon in most cases requires some kind of modeling" (Becker-Heidmann and Harkness, 1995:818). The dating of shallow soils is particularly problematic since they often remain in an open system causing "a significant underestimation of their true age" (Orlova and Panychev, 1993). This is particularly true for the post-glacial period in China, where many paleosols are separated by only thin loess deposits, and where many ^{14}C dates on such paleosols are "often much younger than the true age of the sediments" (Head et al., 1989:680). As a result, despite the

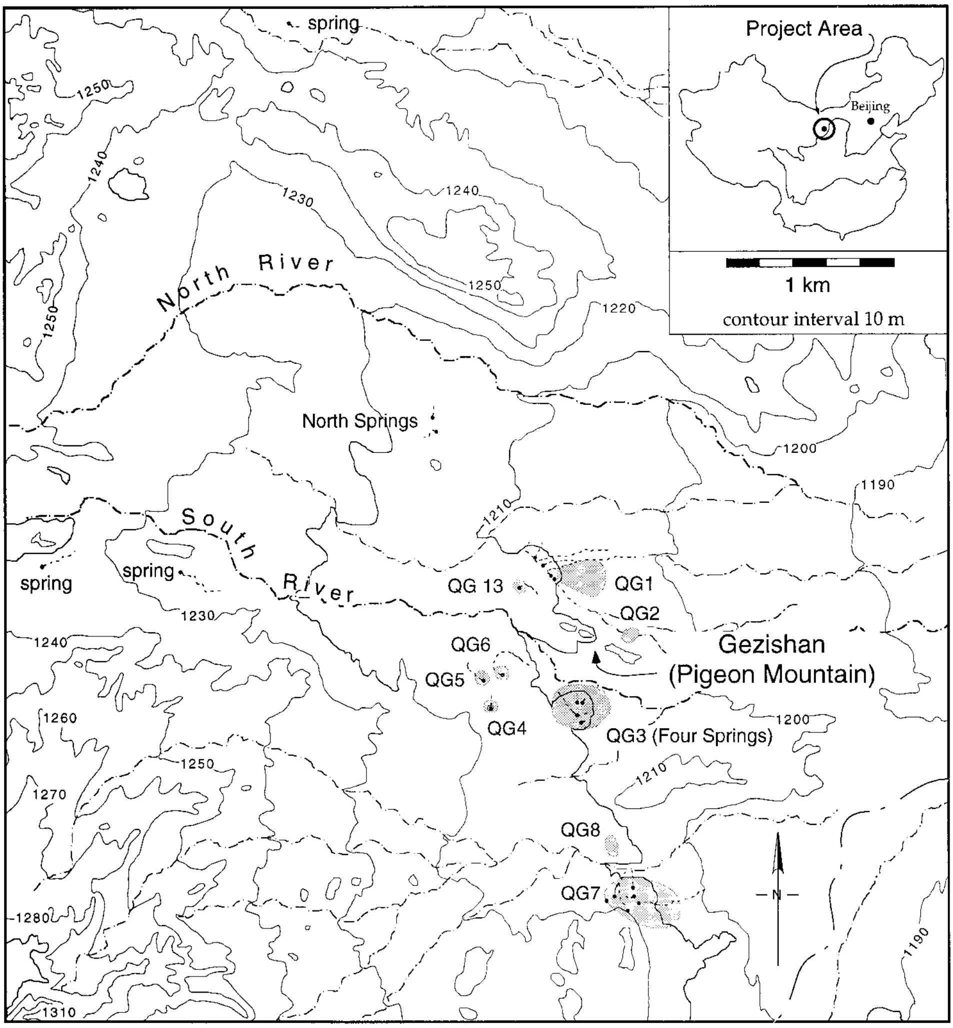

Figure 2. Pigeon Mountain archaeological sites.

presence of only five to -six cycles of deposition/soil development/erosion in any one location (Gao et al., 1993; Wu, 1992; Zhou et al., 1991), the >600 ^{14}C dates for Holocene paleosols in north-central and western China span the whole of the Holocene. While a frequency curve of these ^{14}C dates does reflect six principal periods of Holocene soil formation, many dates fall into time periods when no soils were apparently forming (Figure 3). In short, any single date or even set of dates on soil carbon and/or humates from an individual locality may easily be in error and probably should not be used in any definitive way. Attempts have been made to deal

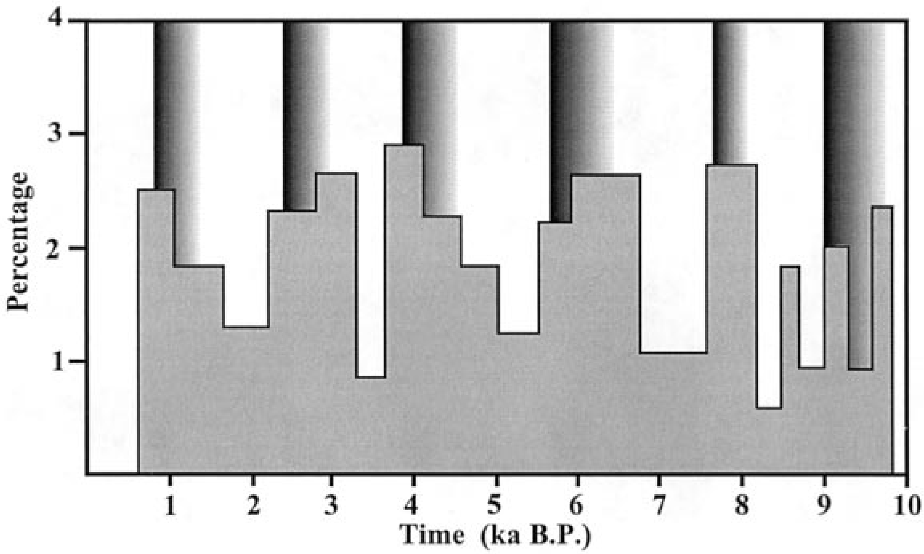

Figure 3. Frequency histogram of >600 ^{14}C dates from Holocene paleosols in north-central and western China (modified from Zhou et al., 1991). Gray-scale bars, added here, show approximate chronological position of Holocene soils identified by Feng et al. (1993), Gao et al. (1993), Wu (1992), and Zhou et al. (1991).

with these problems in a variety of ways, including the dating of cellulose from fossil wood (Head et al., 1989), pretreatment of charcoal fragments (Donahue, 1993), and direct AMS dating of pollen (Zhou et al., 1996). Here, we use the dating of culturally deposited charcoal, rather than naturally deposited organic remains, to address chronological issues associated with the Younger Dryas loess/soil sequence in central China.

These dating problems are compounded by the nature of Chinese loess soils and by the nature of the Younger Dryas itself. In many late Pleistocene-Holocene loess/paleosol sequences, two or more paleosols are often superimposed, producing paleosol complexes of overlapping, stratigraphically compressed, paleosols with ages spanning several thousands of years (Liu, 1985). In addition, Chinese paleosols can form very quickly, often in only a matter of 1 to 2 centuries (Maher and Thompson, 1995), particularly those that are poorly developed and consist of only buried organic A-horizons. As a result, the variance of the date(s) on a soil may significantly exceed the actual time of soil formation. Where soils form during periods of very rapidly changing climatic conditions, even a statistically accurate ^{14}C date, when used to extrapolate to other data, may suggest climatic conditions dramatically different from those under which the soil formed. This is a particularly troubling problem during the period from 12,000–9000 radiocarbon years ago, when even dates with limited variance correspond to calendrical periods of 1000 years or more

(Bartlein et al., 1995). These problems are marked for the Younger Dryas, a climatic episode that is both relatively short, lasting only ~1100 calender years, sharply bounded, both beginning and ending within 10–20 year periods, and highly variable (Alley et al., 1993; GośLar et al., 1995; Johnson et al., 1992; Mayewski et al., 1993). Thus, the climatic signal generated by soil formations dated at or near the ~11,600 cal yr (~10,080 ^{14}C yr) end of the Younger Dryas are difficult to interpret.

LATE QUATERNARY ENVIRONMENTAL CHANGE IN CENTRAL CHINA

What little is known about the Late Quaternary environmental conditions along the southern flanks of the Helan Mountains (Figure 1) must be extrapolated from external sequences. To the east and south, these are primarily loess sequences (An et al., 1990, 1991, 1993; Derbyshire, 1983; Kang, 1990; Sun and Zhao, 1991; Yuan, 1978), while to the west and south these are primarily lake level sequences (Chen and Bowler, 1986; Fang, 1991; Huang and Cai, 1987; Lister et al., 1991; Geng and Chen, 1990; Pachur et al., 1995). The definition of Holocene climatic change is based primarily on paleosol sequences (Wu, 1992; Zhou et al., 1991, 1994) and historical records (Feng et al., 1993). Limited description and dating of alluvial sequences (Geng and Chan, 1992; Hofmann, 1994; Pachur et al., 1995) and dune sequences (Gao et al., 1993) are available for the Helan Mountains itself.

Along the margins of the Helan Mountains, as in most of north-central China, surfaces of Plio-Pleistocene loess, and associated fluvial and alluvial deposits, are heavily weathered and are characterized by a deep and extremely well-developed soil. This erosional surface creates a vast unconformity on which local late Quaternary geomorphological sequences are built. In many areas, such as the Ordos to the east of the Helan Mountains, the surface of this soil forms a vast peneplane to which mixed deposits of Mid-to-Late Paleolithic materials have deflated. In the Pigeon Mountain basin (Figure 2), as in the Ordos, a series of depositional and erosional sequences has been repeatedly built on, and eroded down to, this surface. As a result, most, if not all, of the mid-to-late Pleistocene sequence is missing. At present, it is only possible to reconstruct the late Quaternary sequence from the regional paleoenvironmental history.

Late Quaternary climatic conditions are complex for China as a whole (Winkler and Wang, 1993), and local paleoenvironmental change is best described through a combination of a loess sequence from south of the Tengger Desert (An et al., 1993) and lake sequences from the Tengger itself (Pachur et al., 1995), refined with locally available data. Between about 39 and 23 ka, cool summers and a higher precipitation regime created a system of large lakes and semiaquatic environments in what is now the Tengger Desert. These higher lakes are associated with marked alluviation on the desert margins and increased water supply across the alluvial fans of the Helan Mountains (Hofmann, 1994). Increasing aridity after 18 ka led to the complete desiccation of the Tengger lakes and the initiation of dune building on the lake floors. This aridity corresponds to the inception of the Malan loess,

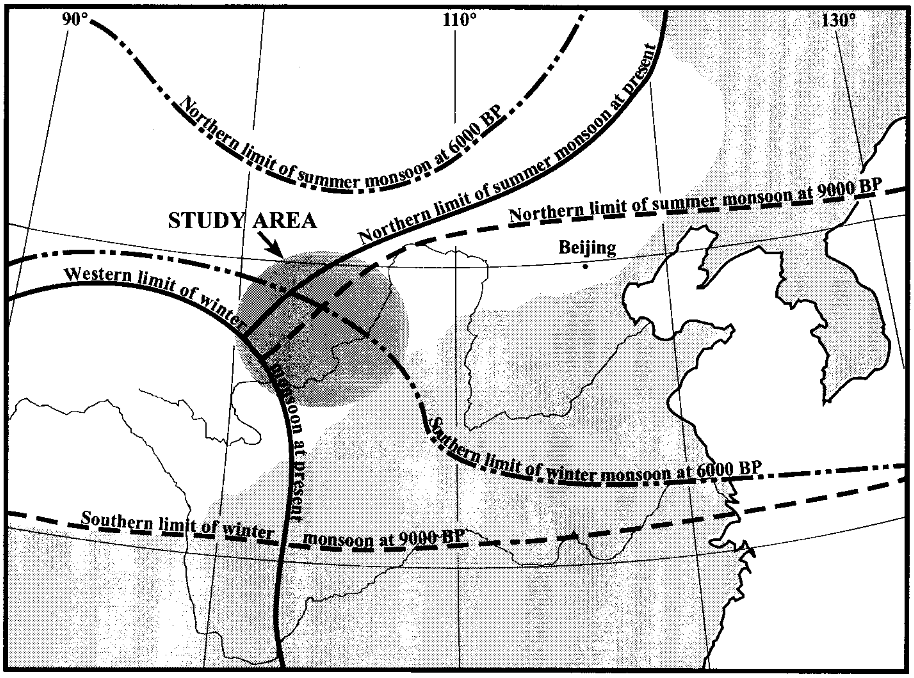

Figure 4. Variation in the limits of the winter and summer monsoon during the Holocene in China. Stippled area is the modern monsoon climate region (after Winkler and Wang, 1993).

which began to accumulate at the Baxie section south of the Tengger by 17 ka. After about 13 ka, lakes approaching in size, but not as large as those of the earlier period, reappeared in the Tengger Desert and lasted until shortly after 11 ka. Evidence of increased spring activity during this period is evident on alluvial fans on both the eastern (Hofmann, 1994) and western (Pachur et al., 1995) margins of the Helan Mountains. For the most part, this variation between periods of aridity and increased annual precipitation has been attributed to the strengthening and weakening of the summer monsoon (An et al., 1991). While this may be true for most of China, the Helan Mountains and the eastern Tengger Desert are at the limit of the summer monsoon's effectiveness (Geng and Chan, 1992), and the role the winter monsoon has played in creating the Tengger sequence remains unclear (Figure 4).

The moist interval between 13 and 11 ka was followed by a brief, but relatively intense period of desiccation lasting from about 10.8 ka to after 10 ka during which regional lakes again dried up and the Taohe Loess accumulated rapidly at the Baxie section. The remainder of the Holocene is characterized by a similar pattern of alternating periods of lake desiccation/aeolian activity and lake recharge/incipient soil development (Gao et al., 1993). However, the overall trend is one of increasing aridity throughout the Holocene, with sequent lake high stands reaching lower elevations at each iteration (Geng and Chen, 1990; Pachur et al., 1995). Lake high stands are evident between about 8.7–7.3 ka, 6–4.6 ka, and 1.9–1.4 ka, and are interleaved with episodes of increased aeolian activity and dune building. The periods of increased available moisture suggested by these lake high stands appear to be contemporaneous with periods of soil formation and to periods of warmer and wetter climatic conditions (Liu, 1985; Zhou et al., 1994; Maher and Thompson, 1995). Feng et al. (1993) recognize eight cool/dry periods separated by warm/wet intervals during the Holocene. Several of these are minor events, however, and the differential extent of summer monsoons into northwestern China (Winkler and Wang, 1993; Zhou, et al. 1996) may account for some of the confusion in the paleosol record.

YOUNGER DRYAS AGE SOILS IN CENTRAL CHINA

The loess-paleosol sequence for the last 130,000 years in central China suggests a strong correlation between the formation of paleosols, the extension of the summer monsoon north and west into a more continental position, and changes to warmer and wetter conditions (An et al., 1991; Banerjee, 1995; Heller and Liu, 1986; Kemp et al., 1995; Maher and Thompson, 1992, 1995; Winkler and Wang, 1993). This pattern is clearly documented for the well-developed Late Pleistocene soils, but also is evident for less well-defined transitional and Early Holocene soils as well (An et al., 1990; Liu, 1985; Zhou et al., 1991, 1996). This pattern has also been described for the extended Chinese historical period (Feng et al., 1993; Zhang, 1984; Zhou et al., 1994).

In north-central and western China, most data suggest the Younger Dryas is associated with an aeolian deposition bracketed by two soils. In sections near Qinghai Lake on the extreme western margin of the Loess Plateau, the dates on two soils above and below dune sands are considered to be 12,300 $\pm$ 100 and 9910 $\pm$ 80 ^{14}C yr B.P. (Pan and Xu, 1989). This interval correlates well with nearby pollen and lake-level data growth (Van Campo and Gasse, 1993:310) suggesting ". . . increasing moisture and temperature after ca. 12,500 yr B.P. with a maximum at ca. 11,500–11,000 yr B.P." This is followed by ". . . a reversal event toward dry, cold conditions centered around 10,500 yr B.P. and end[ing] at ca. 9900 yr B.P.," and ". . . a major environmental change, as brief as a few centuries, occurr[ing] . . . ca. 10,000–9,900 yr B.P. (ca. 11,000 cal yr), with the establishment of moist conditions favorable to vegetation." This, in turn, matches a record from Qinghai Lake itself, suggesting Younger Dryas aridity between 10,800 and 10,000 yr B.P. (Lister et al., 1991), as well as other paleoenvironmental data from the region (Pachur et al., 1995) and from China generally (Li, 1988; Winkler and Wang, 1993). An initial interpretation of the Baxie section near Lanzhou (An et al., 1990) is in accord with these records, suggesting two paleosols bracket a Younger Dryas-age loess deposition. The Baxie Soil was subsequently interpreted to indicate it formed throughout the Younger Dryas under cool but wet climatic conditions (An et al., 1993). However, more recent work at loess sections in the Ordos region within the great bend of the Yellow River (Zhou et al., 1996) suggest soil formation is limited to a brief period of increased precipitation within a Younger Dryas dominated by cool/dry conditions and aeolian deposition. A soil sequence we defined and dated in 1995 and 1996 matches this interpretation.

PIGEON MOUNTAIN BASIN GEOMORPHOLOGY

The surface geomorphology of the Pigeon Mountain basin is a product of this late Quaternary paleoenvironmental sequence (Figure 4). The basin itself is a shallow drainage basin heading in the low foothills of the southern Helan Mountains where elevations reach less than 1600 m. Limited annual precipitation of less than 150 mm/yr drains eastwards towards the Yellow River across a relatively shallow gradient averaging $<1°$ (in contrast to alluvial fan slopes along the central Helan Mountains, which, as the result of intensive late Quaternary tectonic activity, range from 2° to 4°). Surface sediments consist, for the most part, of course, alluvium on the resistant, and very well-developed, Plio-Pleistocene soil found throughout the region. Isolated outcrops of bedrock materials form occasional low hills within this alluvial plain. The largest of these is Pigeon Mountain itself, a low 100 m by 1 km ridge composed of a friable and rapidly weathering conglomerate, located approximately in the middle of the basin. The surface of this flat-topped ridge is less than 30 m above the surrounding terrain.

Vegetative cover is extremely sparse across the basin and the surface is heavily deflated down to gravels and cobbles in braided stream channels, and to the resistant Plio-Pleistocene soil on the relatively flat surfaces between active channels. These stream channels are marginally incised into the basal alluvium, but the intermittent exposures are only about 1 m in depth. The Plio-Pleistocene soil is itself formed on older alluvium composed primarily of coarse gravels and cobbles. These are cemented by the fine-grained clays and silts of an extremely well-developed soil. The dark red B-horizon of this soil exceeds the 2 m of exposure available in channel profiles. Repeated deflation to this surface has exposed gravel and cobbles (to 20 cm diameter) which form a gobi-like stony surface across the floor of the Pigeon Mountain basin. Within the soil itself, however, many of the cobbles are almost completely decomposed. The Plio-Pleistocene soil surface constitutes a landscape feature evident throughout the Pigeon Mountain basin and in cuts along the Yellow River terrace some 20 km to the east. Artifacts on this deflation surface span at least the last 30,000 years and most are highly polished by sand-blasting. Many are so eroded as to be almost unrecognizable as artifacts, and it is very possible that any cultural materials older than the Late Paleolithic cannot be identified as such due to intense wind modification.

The relatively flat and deflated gobi surface morphology is interrupted in a number of locations by spring mounds composed of stratified dunes surrounding one or more small spring throats. The springs occur along a line oriented 15° west of north and may be fault related. No evidence of spring deposits older than ~15 ka

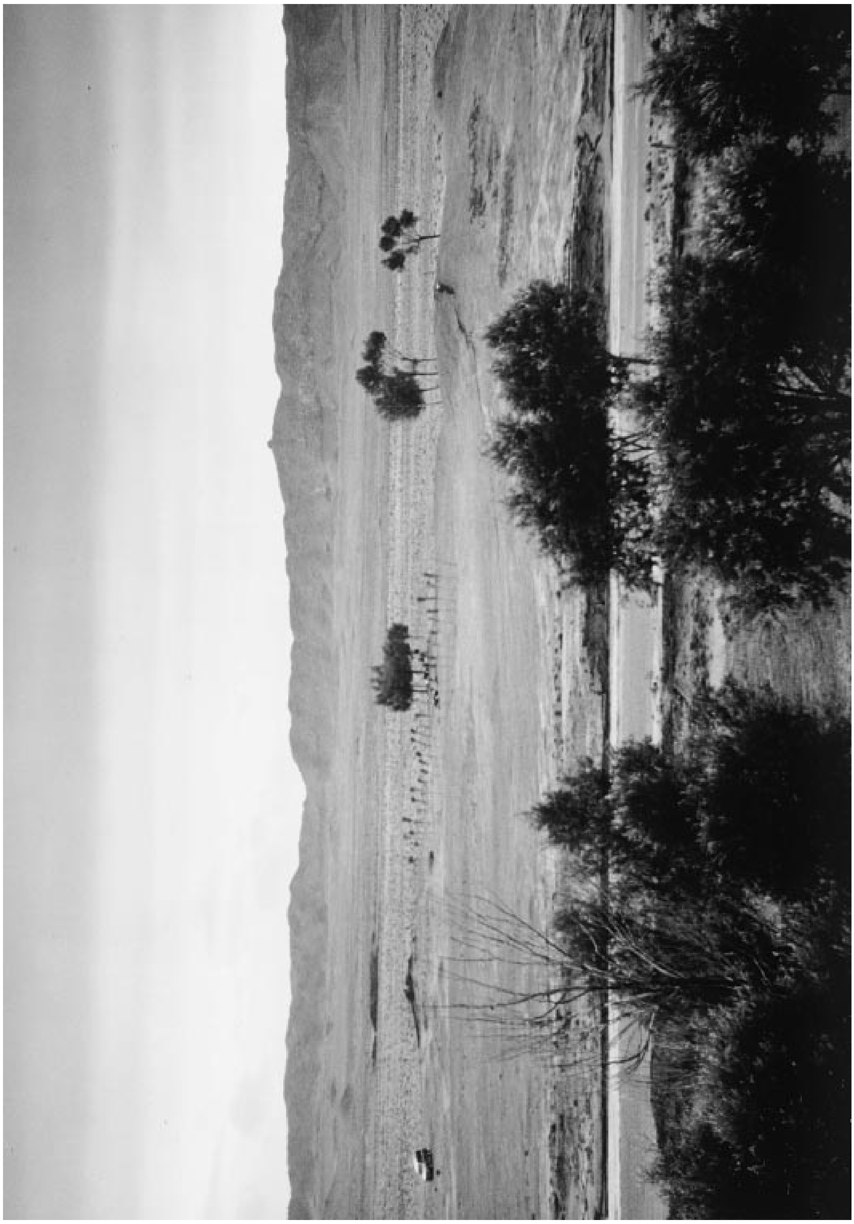

Figure 5. Spring mound/dune complex at the QG3 locality in the Pigeon Mountain basin.

have yet been identified and the springs may be associated with a fault dating to that period. Eight such dune/spring mounds were identified within the Pigeon Mountain basin, but the basin was incompletely surveyed during this preliminary investigation phase, and there may be others. These spring mounds appear to vary in size according to the number of spring throats and the amount of spring discharge at each location. The largest is more than 100 m in diameter and 10 m high, while the smallest is less than 10 m in diameter and 2 m high. The number of spring throats at any one spring mound varies from one to four. Discharge was not measured directly, but appears to vary from as little as 0.001 m^3/s to as much as 1 m^3/s or more. The largest of these are currently used for irrigation and spring flow is diverted through underground piping systems to nearby fields. In some localities aeolian deposits associated with these mounds at times extended across the basin floor, damming the low-gradient spring-fed streams, creating a series of shallow ponds. Without exception, prehistoric cultural materials within the Pigeon Mountain basin are associated with spring mounds and they were the focus of our investigations. One spring was tested extensively, small exploratory exposures were excavated at two more, and shovel scrapes were used to evaluate dune stratigraphy at a number of others. These exposures suggest a relatively uniform depositional sequence at all spring mound locations.

The spring mounds are composed of aeolian sands which appear to have been part of a dune complex that covered much of the Pigeon Mountain basin a number of times during the Holocene (Figure 5). This basin-wide cover of aeolian sands appears to have built and deflated in concert with the Holocene climatic sequence defined above. Calcareous tufas are also present at virtually every spring locality, but it is not yet clear how these tufas relate chronologically to the depositional sequence at the spring mounds and to the overall climatic sequence. They may be similar in age to calcareous crusts on the east and west slopes of the Helan Mountains dating to 13–12 ka (Hofmann, 1994), but they may also be composites of tufas laid down during multiple periods of increased spring flow.

To some extent, the spring mound dunes are self-generating, with aeolian materials deposited in and around the spring-supported vegetation during periods of aggradation. During periods of deflation, areas around the spring complexes were protected and stabilized by this same spring-fed vegetative cover, while away from the springs the aeolian sands were eroded completely down to the more resistant Plio-Pleistocene soil. Dune profiles within the spring mounds exhibit repetitive cycles of aeolian deposition and stabilization, development of an incipient soil and/or organic spring mat, and creation of a deflation surface and erosional unconformity. The sequence is then repeated in overlying strata. Four to five such cycles are evident at the QG3 mound and may be present in the other large spring mounds as well. In some of the small mounds, supported by only small springs, several of the most recent cycles appear to be missing, probably because the upper most materials have been deflated during the current erosional cycle.

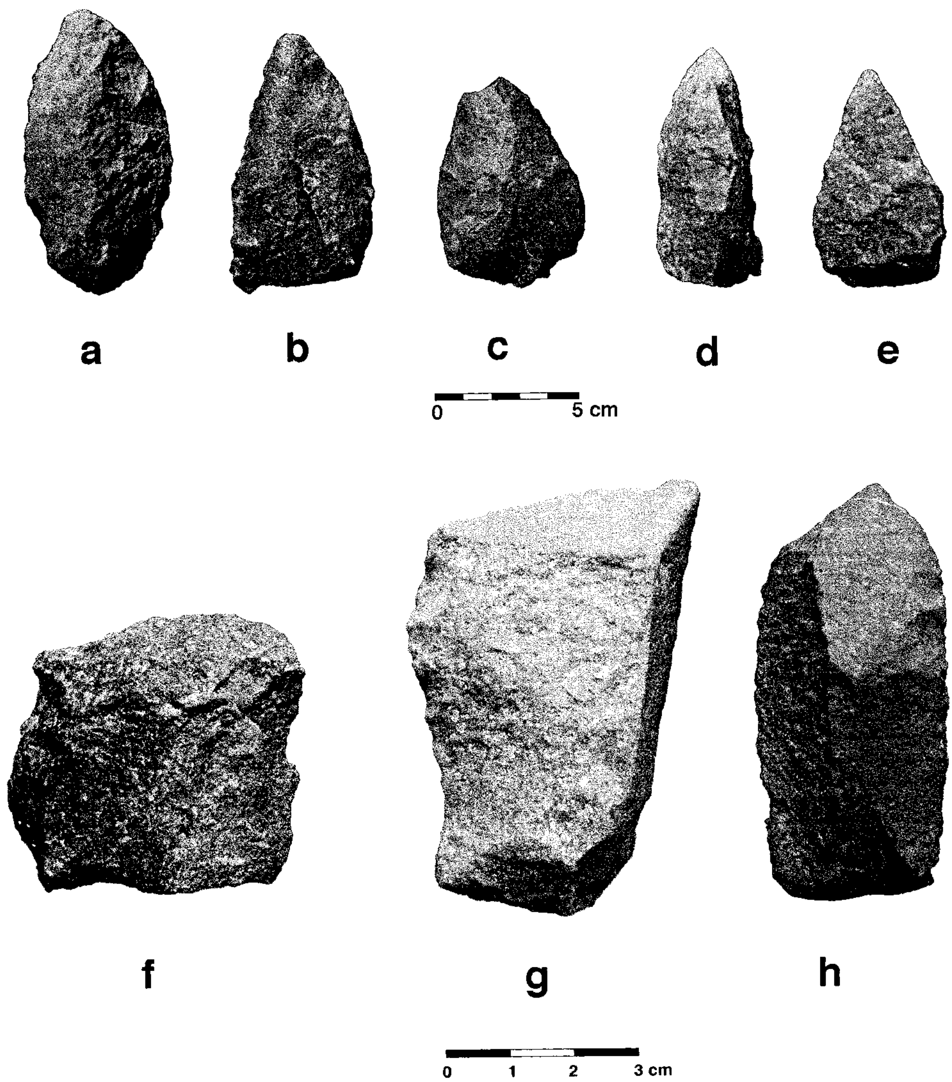

Figure 6. Macrolithic Tools from Pigeon Mountain Basin. a–e, quartzite Helan Points; f, quartzite utilized flake; g, quartzite unifacial scraper; h, greedstone convergent scraper.

LITHIC ARTIFACTS OF PIGEON MOUNTAIN BASIN

Pigeon Mountain Basin lithic artifacts suggest a lithic technology incorporating both macrolithic (Figure 6) and microlithic (Figure 7) elements. Macrolithics, made of quartzite, metavolcanics (greenstone), and fine grained sandstone, include Helan Points (Figure 6[a–e]), a distinctive series of bifaces and unifaces similar to those

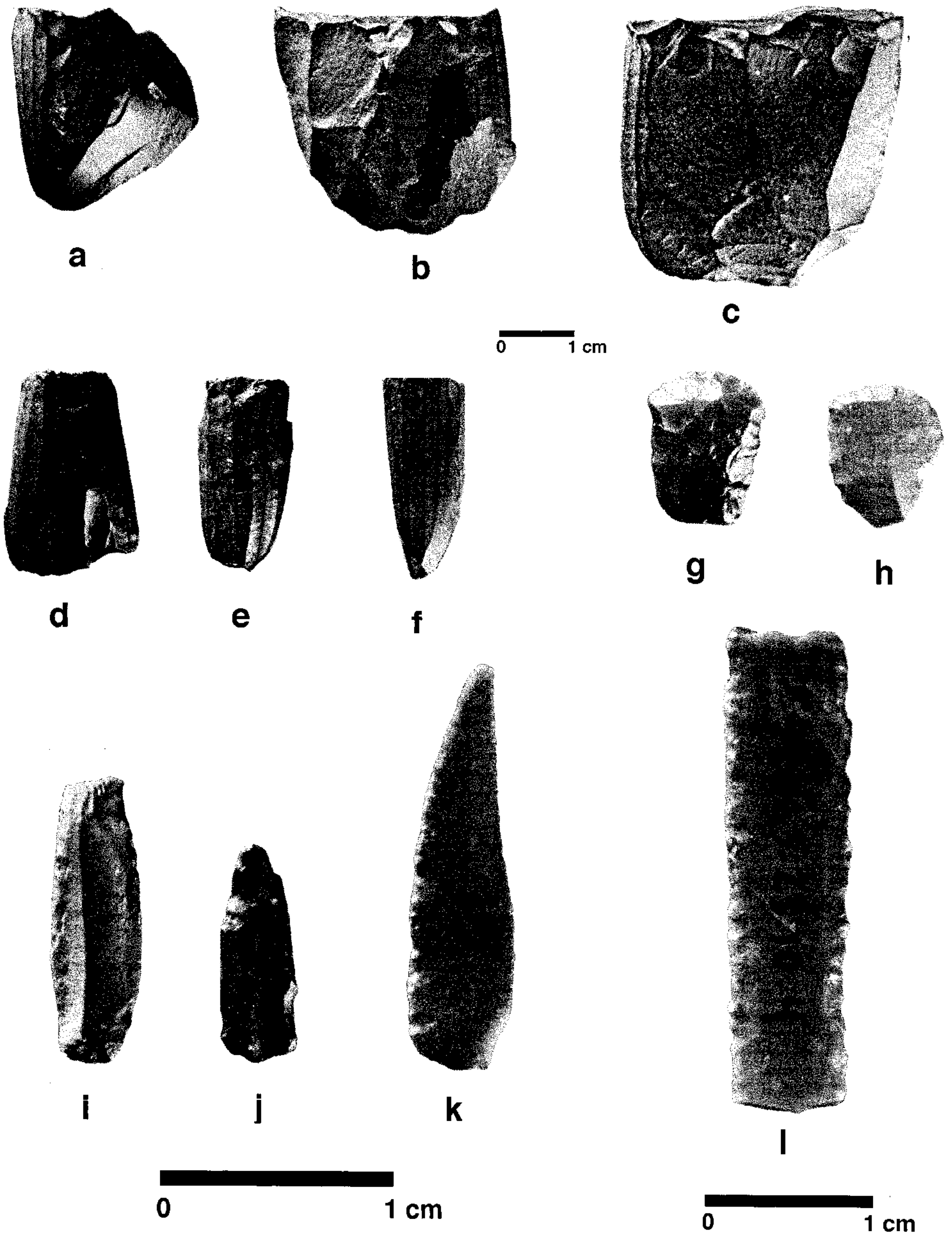

Figure 7. Microlithic Tools from Pigeon Mountain Basin. a,d, chert split pebble microblade cores; b,c,e,f, chert wedge-shaped microblade cores; g–h, chert thumbnail end scrapers; i–l, retouched microblades.

from Hutouliang, Xiachuan, and Xueguan (Chen 1984; Chen and Wang 1989), cores, unifacial scrapers (Figure 6[g]) and gouges made on cobbles and flakes, flake (Figure 6[f]), and (rarely) blade (Figure 6[h]) blade tools, battered subangular chunks (spheroids), and coarse percussion debitage. The microlithic component is mostly chert and chalcedony and includes split pebble (Figure 7[a,d]) and wedge-shaped (Figure 7[b,c,e,f]) microblade cores, thumbnail end scrapers (Figure 7[g,h]), retouched microblades (Figure 7[i–l]), and flakes, fine pressure retouched bifaces, small triangular arrow points and microlithic debitage. Many microblades are medial segments with distal and proximal ends snapped off (Figure 7[l]), probably intended to be inserted horizontally in slotted bone armatures. Less frequent are end scrapers (Figure 7[i]), drills (Figure 7[j]), and knifelike tools (Figure 7[k]) hafted on one end. Thin, lightweight grinding stones, partially ground celts made on flaked blanks and elongate stream cobbles, and plain and cordmarked pottery complete the assemblage. This combination of the macrolithic and microlithic suggests a technological transition between the wholly macrolithic Late Paleolithic technologies and technologies of the early Neolithic based on microlithics and ground stone tools.

The entirely macrolithic Late Paleolithic assemblage at Shuidonggou Locality 1, across the Yellow River 60 km to the east, for example, displays different raw materials and flaking techniques, large prismatic blades, fewer tool types overall, and no Helan points (Teilhard and Licent, 1924; Zhou and Hu, 1985; Ningxia Museum and Ningxia Bureau of Geology, 1987; Yamanaka, 1993; Wang and Yu, 1996). The Pigeon Mountain materials more closely resemble those from Shuidongou Locality 2, which although not yet formally studied, is known to contain microblades and Helan points.

Nonsystematic surface collections were made at QG1, QG2, QG3, QG7, and QG8, and systematic collections from 15 × 15 m quadrats at QG1, QG4, QG5, and QG6. In addition, test excavations were conducted at the QG3 site. Except for QG3, all these assemblages are dominated by macrolithic tools and debitage (>96%). Macrolithics are also abundant at QG3 but microlithics comprise 48% of the total assemblage. Artifact returns are summarized in Table I.

Raw material preferences for macrolithic and microlithic technologies are clearly reflected in debitage (Table II). Quartzite, metamorphics, and fine sandstone are available locally, as are some of the lower quality cherts and chalcedony. However, the sources of greenstone and finer cherts and chalcedony remain to be discovered; the cherts may have been imported, but the greenstone source must be relatively near since it frequently arrived at Pigeon Mountain basin sites in the form of large, unreduced cobbles.

CHRONOLOGY AT QG3

Our 1995/1996 exploratory investigations focused on QG3, one of the larger spring mounds within the Pigeon Mountain basin. Chronological controls for the earliest portion of the sequence at QG3 are provided by five $^{13}C/^{12}C$ corrected ac-

Table I. Collected and excavated artifacts from Pigeon Mountain Basin sites.

	QG3*	QG3**	QG1[H]	QG4[HH]	QG5[HH]	QG6[HH]	Total
Macrotechnology							
Helan Point	16		6	4	2		28
Gouge	3		8	6	1	2	20
Scraper	6	5	8	10	1		30
Flake Tool	26		7	5			38
Blade Tool	2		3			2	7
Spheroid	6		19	20	8	2	55
Core			5				5
Chopper				1			1
Miscellany				5	1	1	7
Debitage	58	384	231	492	110	204	1479
Microtechnology							
End Scraper	6	8	2	1			17
Flake tool	14	7		3	1		25
Retouched microblade	3	9					12
Arrow Point		1	1	1			3
Biface	2		2				7
Anvil		2					2
Microblade core	25	15	9				49
Microblade	50	166		8			224
Debitage	69	212	16	42	5		344
Ground Stone							
Celt	14		7	5	2	2	30
Milling slabs	5	1	1				7
Hand stone	1						1
Ceramic							
Historic		2	4	41		10	57
Cord-marked; plain		4		1			3
Totals	309	816	329	645	131	223	2453

* Purposive surface collection.
** Excavated collection.
[H] Collected from four 15 × 15 m quadrants.
[HH] Collected from one 15 × 15 m quadrat.

celerator dates (Figure 8) on charcoal deposited by Late Paleolithic occupants of the locality, and, thus, differ from the great majority of radiocarbon dates on bulk soil carbon which complicate interpretation of the post-glacial loess/paleosol sequence in north-central China. All dated samples consist of large charcoal fragments treated to remove calcareous material and younger humic matter. We consider them to be reliable estimates for the age of the deposits which contain them because these samples (1) are not derived from soil organics, (2) consist of completely reduced charcoal, (3) are composed of large fragments rather than dispersed fine-grained charcoal more subject to migration, and (4) have an obvious linear relationship with depth.

The basal post-glacial deposit is a massive, well-sorted, fine-grained aeolian sand

Table II. Frequency and proportion of raw materials in Pigeon Mountain Basin debitage.

Raw Material	Macrodebitage		Microdebitage		Microblades	
	n	%	n	%	n	%
Quartzite	493	58%			12	5%
Greenstone	292	34%			2	1%
Dark Metamorphic	43	5%	18	6%	23	9%
Chert/Chalcedony	18	2%	259	91%	163	67%
Other	12	1%	7	3%	45	18%
Total	858		284		245	

whose overall age is uncertain, but, given its stratigraphic position, was most probably deposited at the end of the last glacial cycle (Figure 9). It may, in part, be related to the deposition of the Malan Loess (An et al., 1993). The top 20–25 cm of Sand I represents the B-horizon of a paleosol, but the upper surface is an erosional unconformity and there is no evidence of the paleosol A-horizon. Macrolithic debitage and a few flake scrapers are found throughout this zone, and a flat sandstone grinding slab was observed eroding out of Sand I near its upper surface. These are most heavily concentrated on a thin gravel stringer derived from a brief fluvial event and/or period of increased spring flow, which occurs intermittently across the site within this zone. A radiocarbon assay on culturally deposited charcoal collected from this gravel stringer dates to 12,710 ± 70 yr B.P. (Beta 97242). A charcoal fragment collected from Sand I at the deflationary surface dates to 11,620 ± 70 yr B.P. (Beta 86731). It is not yet possible to be certain whether the younger charcoal fragment was associated with the deposition of Sand I or postdates the deflationary event which removed the upper portion of the soil, and, hence, after the formation of the soil itself. Regardless, it provides a limiting age for the deposition of the overlying deposits.

The deflationary event at the Sand I surface (Figure 9[B]) is overlain by a second aeolian sand deposit with an incipient paleosol developed on its surface as well. Sand II consists of well-bedded, well-sorted, fine-grained sand layers containing thin lenses of coarser-grained sand. In some areas, several of these bedding units are lightly stained by organics. Individual beds are ~1 mm thick and alternate between inorganic and organically stained units. The organic stained sands increase in frequency and thickness from bottom to top. Macrolithic tools and debitage are present, but microlithic tools and tool production detritus, including microblades, microblade cores, and end scrapers occur with increasing frequency throughout the deposit, although there is little culturally deposited charcoal or ash staining within the sand. A single large piece of charcoal from within Sand II (below the paleosol on its surface) dates to 10,230 ± 50 yr B.P. (Beta 97241). This places the deposition of Sand II within the Younger Dryas, but the sample was collected from a test unit excavated in arbitrary levels and the position of the date within the Sand II itself cannot yet be determined.

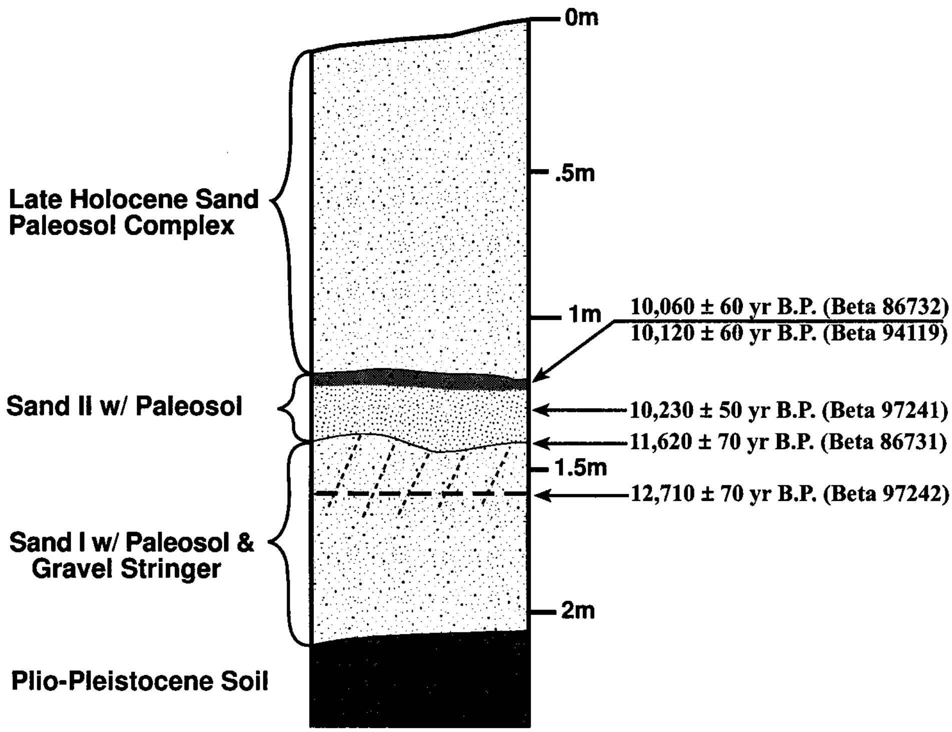

Figure 8. Lithology and associated radiocarbon dates of the QG3 aeolian/paleosol sequence, Pigeon Mountain basin, Ningxia, China. Dashed line indicates position of gravel stringer.

A highly organic A-horizon was formed on the surface of Sand II (Figure 9[A]), but there is little evidence of additional soil development below this humus zone, and it appears to be primarily an organic spring mat formed on top of aeolian sands, a feature common to many quickly forming Holocene paleosols (Liu, 1985). Two radiocarbon samples composed of culturally deposited charcoal recovered from this organic mat date to 10,060 ± 60 yr B.P. (Beta 86732), and 10,120 ± 60 yr B.P. (Beta 94119). The aeolian deposition of Sand II is thus bracketed to a period of ~1600 radiocarbon years between 11.6 and 10 ka, a period which corresponds almost exactly with age determinations for the Younger Dryas derived from ice cores (Mayewski et al., 1993).

CULTURAL SEQUENCE

Table III summarizes the stratigraphic distribution of artifacts recovered from QG3 grouped by major stratigraphic unit, Sand I and Sand II. Artifacts from the unsystematic general surface collections are also enumerated for comparison. Given the nearly 3000-year depositional span represented by the two sands, differ-

Figure 9. View of exposed profile at QG3. (A) Trowel points to the top of the Sand II sequence. (B) Arrow points to erosional surface between Sand I and Sand II.

Table III. Stratigraphic distribution of artifacts from QG3.

	General Surface	Excavated Surface	Sand II	Sand II/I	Sand I	Site Total
Microtechnology						
End Scraper	5	1	6	1	1	14
Flake Tool	14		3	1	3	21
Retouched Microblade	3	8		1		12
Arrow Point			1			1
Biface	5					5
Anvil			2			2
Microblade Core	19	6	9	2	4	40
Microblade	4	46	150	2	10	212
Debitage	50	19	130	3	28	230
Macrotechnology						
Helan Point	16					16
Gouge	3					3
Scraper	6			2	3	11
Flake Tool	26					26
Blade Tool	2					2
Spheroid	6					
Debitage	58	58	136	91	157	500
Ground Stone						
Celt	14					14
Grinding Stone	3				1	4
Hand Stone	2					2
Ceramic						
Historic	2					2
Plain	2					2
Cord-marked	2					2
Fauna						
Cut & Polished Bone			1			1
Bones and Teeth		103	781	3	34	921
Total	242	241	1221	107	238	2043

ences between the two units in relative amounts of macrolithic and microlithic artifacts may address cultural adaptation and development in this critical transitional period. Although the sample of excavated macrolithic tools is small, all five are from Sand I or the Sand1/II interface. Macrolithic debitage is more abundant in Sand I, and relatively less so in Sand II (df 1, Chi Square = 126.87; p = 0.0001), while numbers of microlithic tools and debitage drastically increase in Sand II. This may indicate an increasing displacement of macrolithic technology by microlithic technology after ~11.6 ka. The slab grinding stone from Sand I apparently dates to between ~12,710 and ~11,620 radiocarbon years ago, making it one of the earliest known in northern China (Underhill, 1997).

SUMMARY

Data from the Pigeon Mountain basin, based on culturally derived charcoal dates rather than on soil humus, suggest a Younger Dryas aeolian deposition is bracketed by two soils dating to 12–11 ka and to ~10 ka. The Pigeon Mountain basin sequence corresponds to other loess/paleosol sequences in the region (Pan and Xu, 1989), which, together, suggest the Younger Dryas in north-central and western China was characterized primarily by aeolian deposition. These loess/paleosol data, in turn, correspond to lake level sequences (Lister et al., 1991; Pachur et al.,1995) and to vegetational data (Li, 1988; Van Campo and Gasse, 1993; Winkler and Wang, 1993), which also suggest the Younger Dryas in central and western China was generally a period of colder and drier climatic conditions, with a reduced summer monsoon influence very much like other cold periods in the late Quaternary paleoclimatic sequence of China.

The QG3 paleosol sequence is not sufficiently complete to address the possible occurrence of a brief soil-forming interval within the Younger Dryas, as suggested by Zhou et al. (1996), although the nature of the Sand II deposits does suggest a high degree of climatic variability during the Younger Dryas (Mayewski et al., 1993). This climatic variability immediately before, during, and after the Younger Dryas, appears to be associated with a transition to broad-spectrum foraging and seed processing by hunter-gatherers in western and central China, and, ultimately to the advent of agriculture.

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