Mid-Miocene (Barstovian) environmental and tectonic setting near Yellowstone Park, Wyoming and Montana

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ABSTRACT

Middle Miocene (ca. 10.5–16.0 Ma) deposits in the Greater Yellowstone region are confined to the newly defined Hepburn's Mesa Formation in the Yellowstone Valley, Park County, Montana, and the Colter Formation in Jackson Hole, Teton County, Wyoming. The Hepburn's Mesa Formation was deposited in and adjacent to a perennial saline lake, as shown by the mineral assemblage (clinoptilolite, smectite, calcite, halite, and gypsum) and sedimentary features (fine grain size, laminations, mud cracks, and mud flakes in pebble-sandstone lenses). Biostratigraphic dating indicates that the saline lake formed prior to 16.0 Ma and persisted until after 14.8 Ma. The presence of a saline lake and a fossil mammal fauna dominated by geomyoid rodents indicates an arid or semi-arid climate. Some of the abundant tuffaceous components of the Hepburn's Mesa Formation may have blown from coeval volcanic vents represented by the Colter Formation in Jackson Hole.

The Hepburn's Mesa and Colter Formations suggest that both the northern and southern ends of the Greater Yellowstone region were reacting similarly to tectonism during the middle Miocene. (1) Erosion dominated between ca. 18 to 17 Ma. (2) Subsidence of the basins accompanied by extensional block faulting took place from at least 16.0 to 14.8 Ma. (3) After 14.8 Ma but before 8.6 Ma, uplift occurred in northern Yellowstone Park; strata in the Yellowstone Valley and Jackson Hole tilted and eroded, then subsided faster than they had prior to 14.8 Ma. Inasmuch as the chronology of extensional deformation and volcanism in the Greater Yellowstone region is so similar to that in the Northern Rocky Mountains, the Columbia Plateau, and the central part of the Basin and Range, all of these areas appear to be related by a common tectonic process.

INTRODUCTION

The Greater Yellowstone region centers on Yellowstone Park but also encompasses adjacent areas in northwestern Wyoming and southwestern Montana (Fig. 1). Most studies of Cenozoic rocks in this region pertain to Paleogene, late Miocene, or Quaternary deposits (Chadwick, 1981, 1982; Love and others, 1973, 1976, 1978; Love and Reed, 1971; Smith and Christiansen, 1980; Sundell and others, 1984; C. W. Barnosky, 1984; Pierce, 1979; Baker, 1976; Richmond and Hamilton, 1960; Meyer and Locke, 1986). This paper focuses on a chapter of Cenozoic history that has received much less attention: the middle Miocene or, more precisely, the Barstovian Land-Mammal Age, 11.5 to 16.0 Ma (upper bound-

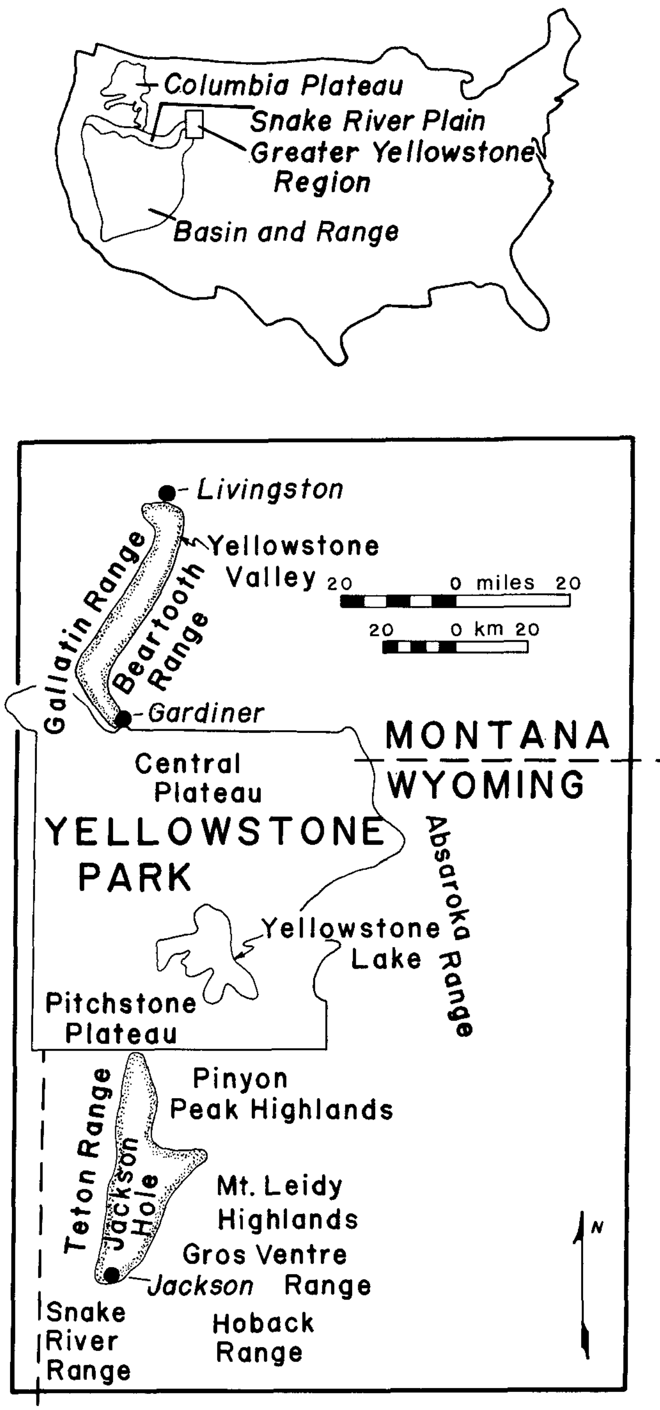

Figure 1. Location of Barstovian deposits within the Greater Yellowstone region.

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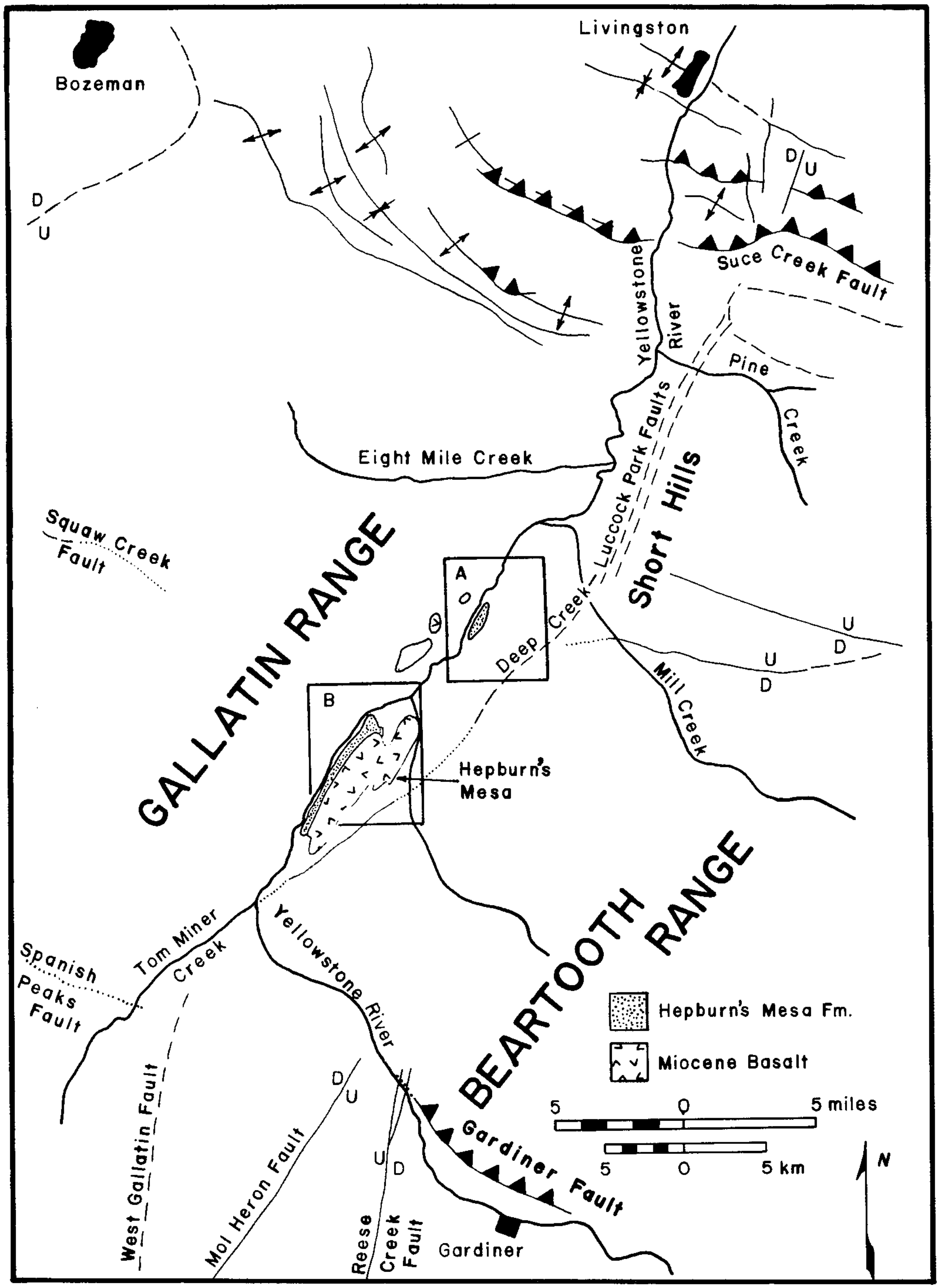

Figure 2. Simplified geologic map of Yellowstone Valley (after Montagne and Chadwick, 1982a).

ary from Tedford and others, 1987; lower boundary from MacFadden and others, 1989, and Berggren and others, 1985). Deposits of this age are scarce in the Greater Yellowstone region but are critical to interpreting middle Miocene relationships with the Columbia Plateau and Basin and Range to the west, the Northern Rocky Mountains to the north, and the Middle Rocky Mountains to the south.

The only Barstovian deposits known in the Greater Yellowstone region are located in the Yellowstone Valley, Montana, and Jackson Hole, Wyoming (Fig. 1). Our purpose here is to present new information from the Yellowstone Valley, integrate it with what is already known from Jackson Hole (Barnosky, 1984, 1986), and discuss the implications of these data for regional paleoenvironments and tectonic patterns.

THE YELLOWSTONE VALLEY

Geologic Setting

Thorough discussions of the geologic setting of the Yellowstone Valley can be found in Montagne and Chadwick (1982a). Only a brief review

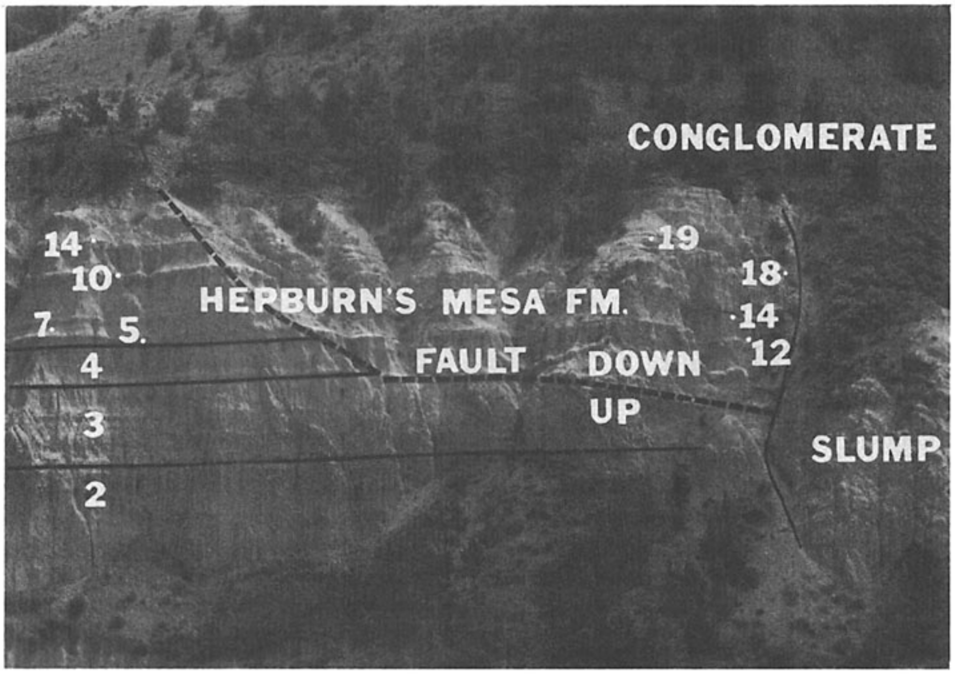

Figure 3. Type section of Hepburn's Mesa Formation at Hepburn's Mesa. Numbered strata correspond to those labeled in the measured section in Figure 5.

of salient points is presented here. The Yellowstone Valley (Figs. 1, 2) is a northeast-southwest–trending half-graben bordered by the Gallatin Range to the west and by the Beartooth Range to the east (Montagne, 1982a; Montagne and Chadwick, 1982b). Horberg (1940) recognized it as "the easternmost of the Tertiary basins so characteristic of the Northern Rockies," but he noted a structural affinity to the Middle Rockies. Fields and others (1985), following Montagne (1982b) and Eaton (1979), regarded the Yellowstone Valley as among the northeasternmost basins in the Basin and Range Tectonic Province, a conclusion supported by the valley's orientation (see Stewart, 1978, p. 4), its structural relationship to surrounding highlands, and its fill of exclusively Cenozoic deposits. Faults bound the valley on three sides: the normal Deep Creek and Luccock Park faults to the east, the Gardiner high-angle reverse fault to the south, and the Suce Creek thrust fault to the north (Fig. 2). Late Cenozoic extensional movement along the Deep Creek–Luccock Park system, and possibly also the Gardiner fault, has lowered the Yellowstone Valley relative to the Beartooth Range (Montagne, 1982a, p. 15).

The Barstovian sediments dip gently northeastward (10° to 20°) and locally may be cut by the Deep Creek–Luccock Park faults, notably at Short Hills (Fig. 2), where remnants of probable Miocene age are uplifted relative to the valley floor (Montagne, 1982b). Paralleling the Deep Creek–Luccock Park fault system, there are minor faults and major fractures. At Hepburn's Mesa (Figs. 2, 3), for example, ~12 m of offset is evident along a fracture plane that dips steeply to the east. The contact of Barstovian deposits with underlying strata is not exposed, but Eocene volcaniclastics crop out in other parts of the valley (Chadwick, 1982). Unconformably overlying the Barstovian rocks, there is a 40-m-thick, coarse conglomerate with imbricated clasts, indicating a source in the Yellowstone Park area to the south. The conglomerate is capped by two basalt flows. The lower flow has been dated at 8.6 Ma, and the upper one, at 5.5 Ma (Chadwick, 1982; dates corrected according to Dalrymple, 1979).

Hepburn's Mesa Formation

Exposures of white, pink, and green tuffaceous claystones, siltstones, and sandstones constitute the Barstovian part of the section in the Yellowstone Valley. These beds are distributed along the east side of the Yellowstone River from ~25 km north of Gardiner to the Emigrant Road Junction ~30 km farther north (Figs. 2, 4). Informally these rocks have been called "lake beds," "White Cliffs" (Horberg, 1940; Montagne, 1982a), and "Chalk Cliffs" (Wood and others, 1941; McKenna, 1955). Here they are formally designated as the "Hepburn's Mesa Formation." The distinctive lithology of the beds and their restriction to the Yellowstone Valley precludes recognizing them as part of the Sixmile Creek Formation, which has been considered widespread in intermontane valleys of southwestern Montana (Fields and others, 1985).

The type section at Hepburn's Mesa (Fig. 2; NE¼, SE¼, Sec. 23, T. 6 S., R. 7 E., lat. 45°18′N, long. 110°49′W) includes measured sections CC-North and CC-South (Figs. 4, 5). Maximum thickness is in the type area, where more than 150 m of section is exposed. Outcrops to the north near Wanigan (CC-4, Fig. 4A) and to the south (CC-9 and CC-10, Fig. 4B) generally are <20 m thick.

Attempts to date the deposits near Hepburn's Mesa by K^{40}-Ar^{40}, Ar^{39}-Ar^{40}, and fission-track techniques have not been successful, because the tuffs are too fine grained, highly zeolitized, and contain detrital zircon (C. C. Swisher and D. W. Burbank, 1988, personal commun.). Recovery of several hundred fossil-mammal specimens, however, permits relatively precise biostratigraphic placement within the North American Land-Mammal Age chronology, which has been calibrated to the radiometric and paleomagnetic time scales at a number of localities (Woodburne, 1987). The stratigraphic distribution of fossil mammals at Hepburn's Mesa documents that the bottom of the formation is of earliest Barstovian or latest Hemingfordian Land-Mammal Age, whereas the top falls in the early part of the Late Barstovian (Table 1).

The Hemingfordian-Barstovian boundary correlates with 16.0 Ma, the division between early and late Barstovian with 14.8 Ma, and the end of the Barstovian with 11.5 Ma (MacFadden and others, 1989; Tedford and others, 1987; Berggren and others, 1985). Hence the Hepburn's Mesa Formation spans at least 16.0 through 14.8 Ma. The bottom of the formation is 20 m below the biostratigraphically derived date of 16.0 Ma, and most likely dates to about 17.0 Ma, as suggested by a K^{40}-Ar^{40} date of 17.6 ± 0.7 Ma on vitric tuff (Hughes, 1980; corrected according to Dalrymple, 1979) and paleomagnetic analyses underway by D. W. Burbank. The K^{40}-Ar^{40} date is from CC-4 (Fig. 4A), where it is impossible to determine superpositional relationships with the cliffs of Hepburn's Mesa 25 km to the south (Fig 4B). The age should be considered a maximum inasmuch as the glass on which it was based may have included reworked as well as fresh shards.

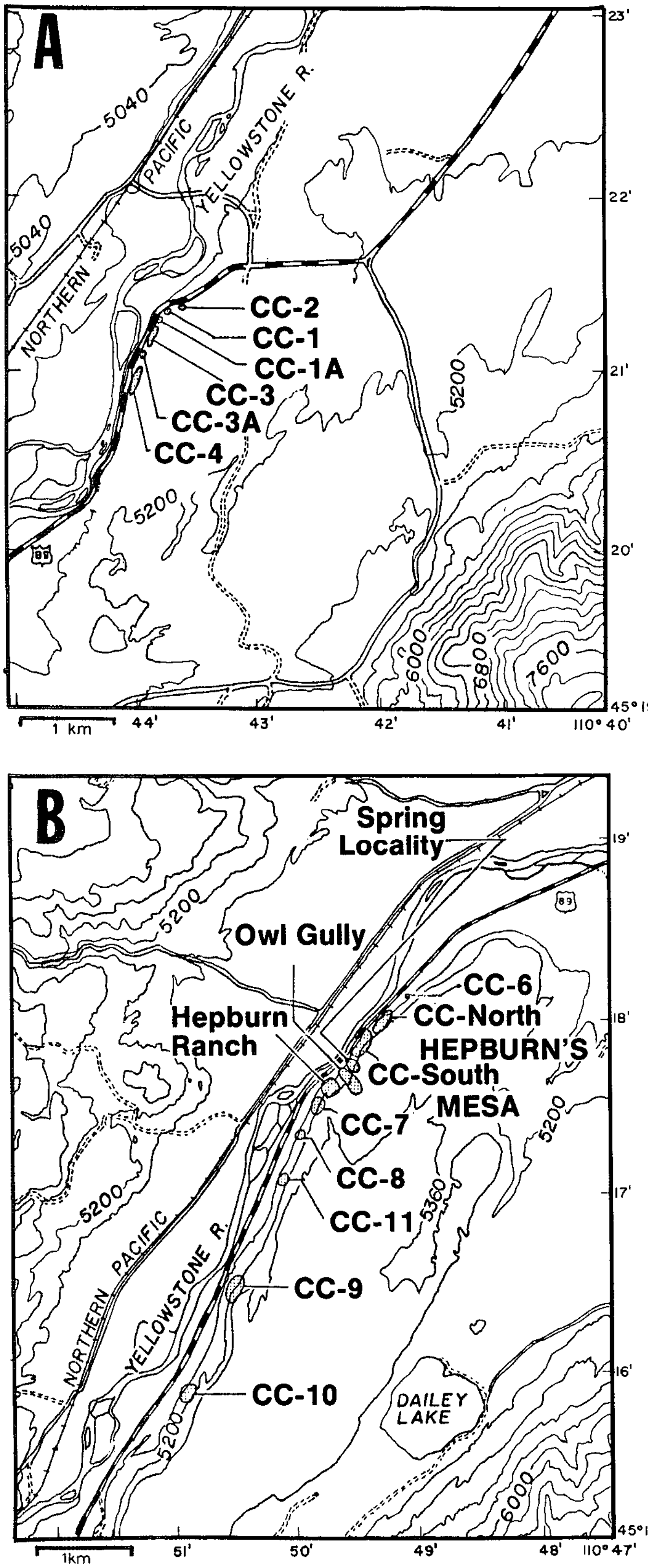

Figure 4. Outcrops of Hepburn's Mesa Formation (stippled areas). See Figure 2 for location of A (northern exposures) and B (southern exposures) within the Yellowstone Valley. Each outcrop is named (for example, CC-North) for convenience of discussion. Outcrops from CC-North to Hepburn Ranch are referred to collectively in the text as "near Hepburn's Mesa." Base map for outcrops is traced from U.S. Geological Survey Fridley Peak 15′ quadrangle.

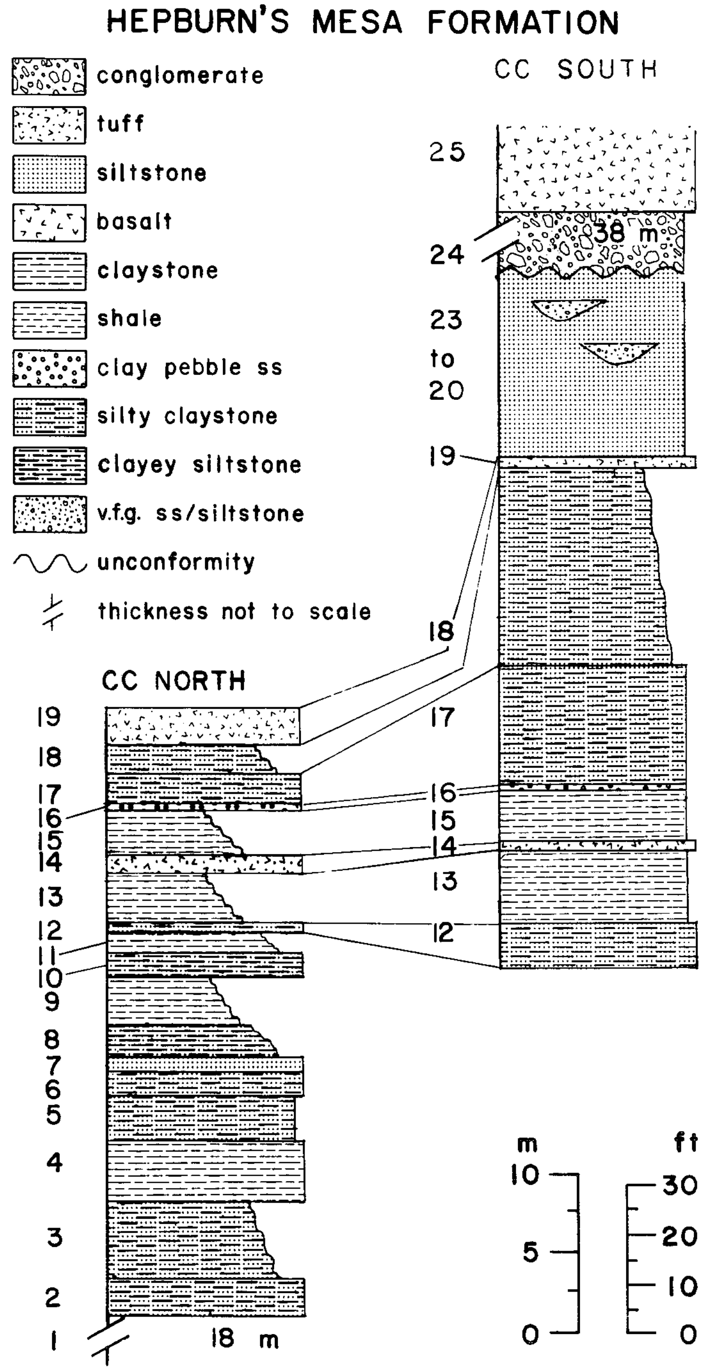

Figure 5. Measured type section of Hepburn's Mesa Formation at Hepburn's Mesa. The fault shown in Figure 3 causes part of CC-North to be repeated at CC-South.

Barstovian Environment in Yellowstone Valley

Detailed mapping of facies, studies of thin sections, and X-ray diffraction analyses were undertaken to determine the depositional environment of the Hepburn's Mesa Formation. The finest grain sizes, almost exclusively clay and silt, prevail near Hepburn's Mesa (Figs. 2, 3) at measured sections CC-North, CC-South, Owl Gully, and Hepburn Ranch (Fig. 4B). Laminations 1 mm to several centimeters thick, beds of claystone and clayey siltstone a few centimeters to a few meters thick, and thin beds (<0.5 m) of zeolitized tuff are found in this area. To the north (CC-1 through CC-4, Fig. 4A) and south (CC-8 through CC-11, Fig. 4B), grain

TABLE 1. FOSSIL MAMMALS IDENTIFIED IN HEPBURN'S MESA FORMATION

Taxon	Stratigraphic Position 1	2	3	4	Geologic range elsewhere PH	EH	LH	EB	LB	PB
INSECTIVORA										
Talpidae (moles)										
Scalopoides	X	X	X		·····	·····	·····	·····	·····	·····
Proscalopidae (moles)										
Mesoscalops		X				·····	·····	·····		
LAGOMORPHA										
Hypolagus (rabbits)		X	X				·····	·····	·····	·····
Oreolagus (pikas)		?	X	X		·····	·····	·····	·····	
RODENTIA										
Sciuridae (squirrels)										
Petauristodon lg.			X	X			·····	·····	·····	·····
Petauristodon sm.			X				·····	·····	·····	·····
Spermophilus		?	X	X					·····	·····
cf. *Tamias*			X						·····	·····
Mylagaulidae (mountain beavers)										
Mesogaulus douglassi		X								
Mesogaulus sp.	X		X			·····	·····	·····		
Aplodontidae (mountain beavers)										
Tardontia			X	X					·····	
Heteromyidae (pocket mice)										
Mookomys			X		·····	·····	·····	·····	·····	
Proheteromys			X		·····	·····	·····	·····	·····	
Perognathus			X	X				·····	·····	·····
Cupidinimus			X	X		·····	·····	·····	·····	·····
Peridiomys	X	X	X	X				·····	·····	
Diprionomys			X	X					·····	·····
Geomyidae (gophers)										
Lignimus			X	X					·····	·····
Mojavemys			X	X				·····	·····	
Cricetidae (mice)										
Copemys			X	X				·····	·····	·····
Eomyidae (mice)										
Pseudotheridomys			X	X	·····	·····	·····	·····	·····	
Pseudadjidaumo			X						·····	
Zapodidae										
Schaubeumys			X	X	·····	·····	·····	·····	·····	
ARTIODACTYLA										
Camelidae (camels)	X	X			·····	·····	·····	·····	·····	·····
Paleomerycidae (deer)										
Dromomeryx		X					·····	·····	·····	
Blastomeryx		X	X	X	·····	·····	·····	·····	·····	
Antilocapridae (antelope)										
cf. *Merycodus*		X					·····	·····	·····	·····
PERISSODACTYLA										
Equidae (horses)										
Parahippus		X			·····	·····	·····	·····		
Merychippus	X	X	X				·····	·····	·····	·····
Hypohippus	X						·····	·····	·····	·····

1 = CC-4. 2 = Below Unit 16. 3 = Unit 16. 4 = Unit 23. PH = pre-Hemingfordian, EH = early Hemingfordian, LH = late Hemingfordian, EB = early Barstovian, LB = late Barstovian, PB = post-Barstovian. Fossils are housed at The Carnegie Museum of Natural History Section of Vertebrate Fossils.

sizes are slightly coarser than those near Hepburn's Mesa. These distal exposures are composed predominantly of massive beds of well-sorted, homogeneous, friable silt and very fine-grained sand, and they contain paleosols as well as tuff beds of relatively fresh glass shards. The distribution of grain sizes and bedding structures suggests that the formation represents a lake in the Hepburn's Mesa area, with northernmost and southernmost exposures probably representing aeolian deposits near the lake.

Thin sections revealed a significant tuffaceous input into the lake. Vitroclastic structure, remnants of sharp glass fragments, bubble shards, and euhedral and anhedral phenocrysts of quartz, plagioclase, and biotite were evident in most samples. Relatively pure tuff beds were characterized by vitroclastic structure in >75% of their thin sections, lack of welding, thinness, and fine grain-size, all of which indicate an airfall origin and distant source for the ash. The nearest known contemporaneous vent area is in Jackson Hole (Barnosky, 1984), but other volcanic centers were distributed throughout the Columbia Plateau and Great Basin (Christiansen and McKee, 1978; Fields and others, 1985).

X-ray diffraction methods (Cook and Morton, 1982) clarified the relative abundance of tuffaceous material and the nature of the lake in which the deposits accumulated. Analyses were conducted on a suite of samples that spanned the entire formation in the Hepburn's Mesa area. Selected samples were powdered, sieved through a 98-micron screen, mounted on petroleum jelly, and exposed to copper radiation on GE diffraction equipment. Separated size fractions of <2 microns, 2–5 microns, 5–10 microns, 10–20 microns, 20–40 microns, and 40–98 microns were analyzed for Unit 19 at Owl Gully.

Clinoptilolite constitutes between 30% and 59% of all samples and is especially abundant in the 20- to 40-micron size range. The prevalence of this zeolite, which is derived from the diagenetic alteration of volcanic glass (Sheppard and Gude, 1973), supports the thin-section evidence for a major tuffaceous component. Diagenetic K-feldspar and cristobalite (<2% combined), both indicative of volcanic detritus (Sheppard and Gude, 1973), also are present. Clay minerals (smectite), found mostly in the <2-micron size fraction, are nearly as abundant as clinoptilolite (Table 2). Small amounts of calcite, quartz, halite, gypsum, and albite also are present.

Clinoptilolite can form in either fresh or saline water (Sheppard and Gude, 1973), but it most commonly is indicative of moderately saline, alkaline lakes with a pH in the range of 8.4 to 9.5 and a high ratio of Si/Al (Hay, 1966; Mumpton, 1981). The presence of halite and gypsum confirms that the water at Hepburn's Mesa was at least periodically saline. Alkaline water would be consistent with the absence of fossil pollen and diatoms in the deposit. More than 30 samples from a range of lithologies were processed by a variety of applicable techniques. Most were barren of microflora or at best produced very few pollen grains (C. W. Barnosky and B. Sherrod, 1988, personal commun.), despite the fact that the sampled lithologies included laminated lake sediments, which usually preserve pollen and diatoms in fresh-water environments. Pollen and diatoms, however, have less chance of preservation in alkaline environments (Traverse, 1988, p. 379).

Typically, deposits that formed in saline lakes exhibit lateral zonation of fresh glass near the shore grading to zeolites and finally to K-feldspar toward the center (Hay, 1966; Mumpton, 1981). Zonation from fresh glass in northernmost and southernmost exposures to zeolites in outcrops near Hepburn's Mesa (Fig. 4B) is evident; within the Hepburn's Mesa area, however, clinoptilolite is the prevalent zeolite throughout. There is no horizontal or vertical gradation to other zeolites or K-feldspar. The lack of zonation suggests that Hepburn's Mesa is a cross section cut near the shore of the lake, and that salinity and alkalinity remained relatively constant through the >1.2 m.y. represented by the outcrops.

Saline lakes today form only in basins that lack external drainage in arid climates. They can be perennial or ephemeral bodies of water. A

TABLE 2. MINERALS IDENTIFIED BY X-RAY DIFFRACTION IN REPRESENTATIVE BEDS NEAR HEPBURN'S MESA

Unit	2	4	7	8	12	14	19
	Percentages						
Clinoptilolite	41	55	59	30	48	41	44
Clay	45	53	34	47	34	37	35
Calcite	4	1	8	2	<1	15	12
Halite	<1	<1	2	<1	<1	<1	1
Quartz	9	5	5	2	3	3	2
Other (includes gypsum, K-Spar, albite, cristobalite)	<1	..	..	19	15	4	6
Total*	~100	115	108	~100	~100	~100	100

*Percentages exceeding 100% reflect imprecise readings because of amorphous glass and/or preferential orientation of clay.

perennial saline lake (for example, the Great Salt Lake, Utah) is defined as "a surface body of brine that persists for many years (tens, hundreds, or even thousands) without drying up" and may be shallow (meters) or deep (hundreds of meters) (Hardie and others, 1978, p. 23). Fluctuations in water level cause dry mudflats to form around the lake (Hardie and others, 1978, p. 21). Ephemeral saline lakes, also called "playa lakes," are characterized as "a shallow body of water . . . that at least once every few years dries up, leaving in the low central area an exposed layer of salt(s) that precipitated out as the brine evaporated" (Hardie and others, 1978, p. 21). Water level in both perennial and ephemeral saline lakes fluctuates through the year. Seasonal or sudden precipitation and/or spring activity can produce "catastrophic expansion (with freshening) and gradual contraction (with increasing salinity)" (Hardie and others, 1978, p. 22).

Paleontologic and sedimentologic criteria suggest that sediments in the Hepburn's Mesa area represent mainly the dry mudflat and shallow perennial lake subenvironments (Hardie and others, 1978). Bones of small and large mammals sparsely distributed through the outcrops provide direct evidence that the water level periodically dropped to expose a dry mudflat. Among the sedimentologic features diagnostic of mudflats are mudcracks, graded laminae a few millimeters thick, thin beds (1–10 cm thick), ripple marks (for example, unit 23 at CC-South), insect burrows, and thin lenses (<0.5 m thick) of pebbly sandstone. The lenses consist of pebble-sized lithic fragments, mud flakes, and rounded mud pellets supported in a matrix of clay and silt (for example, units 16 and 23). Such pockets indicate reworking of brittle mud crusts by wind or water (Hardie and others, 1978). Pebble-sized particles and flakes of dried mud are picked up on the mudflat and then dropped at the edge of the lake. At Hepburn's Mesa, transport by water instead of wind is implied by the inclusion of abundant rodent-sized mammal bones in parts of some lenses, which suggests that the sediment load incorporated bone-laden fecal or regurgitation pellets of mammalian and avian carnivores. Such bone-laden pellets are generally too large to be moved by wind, but they can float in water and be sorted by settling velocity to result in bone-rich areas within the sedimentary deposit.

Criteria suggestive of a perennial-lake subenvironment include (1) the predominance of massive or finely laminated claystone without mudcracks or other features that would indicate subaerial exposure between depositional events and (2) lateral continuity of beds, particularly thin beds of highly zeolitized tuff. Varve-like bands, most clearly evident in outcrop in unit 23 but also recognizable in thin sections of other units, are composed of ~1-mm-thick green layers alternating with 4-mm-thick white layers. Similar layers result from the alternating deposition of clastic particles (green) and chemical precipitates (white) that is characteristic of perennial saline lakes. The massive structure characterizing most of the claystone suggests recrystallization that takes place in the bottom brine and destroys the original laminated fabric, or mixing of the bottom sediments by wind-generated waves in a very shallow lake (Hardie and others, 1978, p. 25–26).

The outcrops near Hepburn's Mesa do not exhibit the sedimentary features definitive of an ephemeral saline lake. Massive claystones are found in the saline mudflat subenvironment of ephemeral lakes, but, unlike the claystones of Hepburn's Mesa, they are "full of salt crystals" and associated with couplets composed of a "thin mud layer (millimetre scale, black, iron sulfide-rich, crowded with salt crystals) overlain by a thicker crystalline salt layer (centimetre scale)" deposited in the salt pan (Hardie and others, 1978, p. 23). Features that are diagnostic of spring deposits, for example, travertine, tufa, oolitic sandstone, or silica sinters are not present, either.

The fauna of the Hepburn's Mesa Formation is consistent with the dry climate mandated by a saline lake. Geomyoid rodents (families Geomyidae and Heteromyidae) are the most abundant taxa both in numbers of genera and numbers of specimens (Table 1). Extant geomyoids are found mainly in arid and semiarid environments (Hall, 1981). ("Arid" and "semiarid" are used *sensu* Cooke and Andrew, 1973, p. 8; and Meigs, 1953.) The explosive radiation of the geomyoids during the Barstovian may have coincided with the onset of increasingly dry climates, as was postulated for an earlier (Arikareean), but less pronounced, radiation of geomyoid taxa (Wahlert and Souza, 1988; Wahlert, in press). Insectivores from the Hepburn's Mesa Formation include the proscalopid mole *Mesoscalops*, which was adapted to digging in sandier drier substrates than modern talpid moles (Barnosky, 1981). The most abundant horse is a relatively high-crowned grazing form, *Merychippus*, which is consistent with the somewhat open landscapes characteristic of arid and semiarid environments. Large tortoises that must have required winter temperatures much warmer than today's in the Yellowstone Valley are also present.

JACKSON HOLE

Geologic Setting

Jackson Hole is topographically and structurally an intermontane basin, with its long axis oriented just slightly more north-south than the axis of the Yellowstone Valley. Jackson Hole has been sinking through most of the Cenozoic (Love and Reed, 1971; Love and Christiansen, 1973). Since the late Miocene, this movement has been accentuated along the Teton fault, which borders the west side of Jackson Hole. Love and Reed (1971, p. 9) likened the movement to that of "two adjoining giant trapdoors hinged so they would swing in opposite directions," with Jackson Hole swinging down and the Teton Range swinging up. Westward dip of Neogene strata in Jackson Hole reflects this motion.

Major volcanic episodes occurred locally during the Eocene, leaving the widespread volcanic units in the Absaroka Mountains. Within Jackson Hole, late Miocene through Pleistocene volcanism is represented in the Teewinot Formation (9.4 Ma), Conant Creek Tuff (5.8 Ma), Huckleberry Ridge Tuff (2.0 Ma), and Lava Creek Tuff (0.6 Ma). The compositions of tuffs in the Eocene volcanic fields indicate compressional tectonics, whereas the late Miocene to Pleistocene ones reflect regional extension (Lipman and others, 1972; Christiansen and Lipman, 1972). Less prolific but nevertheless persistent volcanism also took place throughout the Miocene and deposited the Colter Formation, which consists of more than 1,500 m of pyroclastic flow, lahar, surge, and coarse air-fall deposits that were erupted from vents mainly in the northwest part of Jackson Hole. Chemical analyses of tuff from these deposits and biostratigraphic dating suggest that a transition from compressional to extensional tectonics took place between 18 Ma and 13 Ma (Barnosky, 1984, 1986). The inferred tectonic transition corresponds with a lithologic boundary that separates the upper member of the Colter Formation from the lower member.

Barstovian Deposits

The upper two-thirds of the Colter Formation, the Pilgrim Conglomerate Member, constitutes the Barstovian part of the section. Within it, calc-alkaline rhyolite tuffs provide the earliest evidence of extensional tectonic regimes in Jackson Hole (Barnosky, 1984). The Pilgrim Conglomerate Member is 1,000 m thick and is underlain by 500 m of late Arikareean (ca. 24 Ma) through early Hemingfordian (ca. 18 Ma) Crater Tuff–breccia Member. The Pilgrim Conglomerate Member is dated as late Barstovian, probably not much younger than 13.6 m.y. and not older than

14.8 m.y. Early Barstovian strata may be represented by the 400 m of undated section that separate late Barstovian mammals from early Hemingfordian ones.

The contact between the two members is covered but probably coincides with a relatively short period of erosion in view of the Pilgrim Conglomerate Member's local unconformable relationship with underlying Paleocene rocks and its abundance of rounded quartzite cobbles (Barnosky, 1984). Overlying the Pilgrim Conglomerate Member, there is 1,800 m of tuffaceous lacustrine deposits, the Teewinot Formation. Within the Teewinot, there are pure vitric tuff beds several meters thick, some of which contain obsidian pebbles that suggest nearby volcanoes continued to supply volcaniclastics. The unconformable contact between the Colter and Teewinot Formations, lying several hundred meters below a K^{40}-Ar^{40} date of 9.4 Ma (Evernden and others, 1964), probably corresponds with the onset step-faulting that cuts the Colter Formation and drops blocks eastward.

Barstovian Environment in Jackson Hole

The depositional environment of the Pilgrim Conglomerate Member has been treated in detail elsewhere (Barnosky, 1984, 1986). The lithology, distribution of facies, sedimentary structures, and microscopic textures indicate deposition on and near the flanks of volcanoes.

The fossil mammal fauna, like that of the Yellowstone Valley, indicates an arid or semiarid environment by the dominance of seven genera of geomyoid rodents (Barnosky, 1986). Only two of the 22 Barstovian genera found in Jackson Hole are not also present in the Yellowstone Valley. These include a camel (*Aepycamelus*) and a beaver (*Monosaulax* cf. *curtus*).

DISCUSSION

Climate

Aridity in the Greater Yellowstone region during Barstovian time is consistent with the paleoclimatic pattern postulated for all of southwestern Montana and adjacent Idaho. Thompson and others (1981, 1982) and Fields and others (1985) drew the conclusion of regional aridity from studies of Barstovian sediments in at least 13 intermontane basins. Among the supporting evidence they cited were (1) rapid sediment production, indicating sparsely vegetated landscapes; (2) internal drainage within basins; (3) predominance of smectite-rich paleosols; and (4) abundance of extinct horses, camels, antelope, hares, and tortoises.

Even stronger faunal indication of aridity, as well as of effectively dry loose substrates, is provided by the small mammals of the Yellowstone Valley, Jackson Hole, and the only other Barstovian fauna that has been studied in detail, the Anceney local fauna in the Three Forks Basin (Sutton, 1977; Dorr, 1956). The dominance of geomyoid rodents in these faunas suggests that the climate in southwestern Montana and adjacent areas became drier relative to the late Arikareean, even though late Arikareean climates also are thought to have been generally semiarid (Fields and others, 1985; Thompson and others, 1981, 1982).

Barstovian saline lakes have not previously been documented in the Northern or Middle Rocky Mountains, although Fields and others (1985, p. 18) noted that in the Beaverhead Basin and Middle Ridge area of the Lemhi Basin "bedded gypsum and marl suggest playa-like conditions." A Barstovian saline lake in the Yellowstone Valley suggests that the climate was drier than the present one in southwestern Montana. Presently saline lakes exist only in closed basins where regional moisture lost through evaporation and transpiration considerably exceeds incoming precipitation (Cooke and Warren, 1973, p. 218). The ratio of mean annual precipitation (in centimeters) to mean annual temperature (°C) can range from near 0, as in Death Valley, to 11.25 at Devil's Lake, North Dakota. Most commonly, however, this ratio for basins that hold ephemeral or perennial saline lakes does not exceed 5, as the following examples illustrate: Llano Estacado Playas, Texas, ratio = 4.08; Great Salt Lake, Utah = 3.50; Abert Lake, Oregon = ~3.13; Moses Lake, Washington = 3.00; Dead Sea, Israel = 2.88; Owens Lake, California = 2.00; Wilcox Playa, Arizona = 1.80; Lake Eyre, Australia = 1.60 (Mabbutt, 1977, p. 185; Johnson and others, 1985; Highsmith and Leverenz, 1962). Presently the precipitation:temperature ratio in the Yellowstone Valley is within the range required to hold a saline lake, approximating 3.12 at Gardiner and 4.94 at Livingston (Pierce, 1979, p. 8). The valley, however, is supplied with water year-around from through-flowing rivers and streams originating in highlands to the east, west, and south. In these highlands, the precipitation:temperature ratio is considerably greater than in the Yellowstone Valley itself; for example, 10.50 at Mammoth, 30.66 at the Gallatin Ranger Station, 59.09 at the Northeast Entrance to Yellowstone Park, and 75.00 at Norris Geyser Basin (Pierce, 1979, p. 8). To dry the Yellowstone Valley enough for a saline lake to form, effective precipitation would have to be considerably decreased in the surrounding areas, and stream gradients would have to be flattened so that the valley was a closed basin. Hence it seems likely that the Barstovian landscape in the Greater Yellowstone area was characterized by a regionally dryer climate and lower topographic relief than now exists there. Low topographic relief is substantiated by the absence of coarse clastics in the Hepburn's Mesa Formation.

The causes of increased aridity in the Barstovian have yet to be studied in detail, but drying relative to the Arikareean may involve a change in global air circulation and development of rain shadows as block faulting elevated much of the Basin and Range and the Northern Rocky Mountains. Uplift on such a scale can significantly alter global climatic patterns (Raymo and others, 1988, p. 649). The ejection of huge amounts of volcanic ash into the atmosphere may also have effected climatic changes.

Volcanism

As is the case in the Yellowstone Valley and Jackson Hole, Barstovian deposits in all of the intermontane basins of southwestern Montana and adjacent areas are composed largely of volcanic ash (Fields and others, 1985 and references therein). Only in Jackson Hole, however, is there evidence of a local vent complex. Major episodes of volcanism related to extensional tectonism also were occurring between 14 Ma and 17 Ma in the Great Basin, Snake River Plain, and Columbia Plateau (Christiansen and McKee, 1978). All of these are potential source areas for the tremendous volume of ash that was blown into the basins of southwestern Montana (Fields and others, 1985).

Barstovian Basin Filling

The mid-Tertiary (generally Hemingfordian) unconformity so widely evident in southwestern Montana and adjacent areas (Fields and others, 1985; Barnosky, 1984; Rasmussen, 1973) is recognizable in the Yellowstone Valley by the absence of Oligocene and earliest Miocene rocks. The lack of Oligocene sediments in the Yellowstone Valley sets it apart from most other intermontane basins of the Northern Rockies, which generally contain at least some Oligocene deposits (Fields and others, 1985, p. 18–19). Either mid-Tertiary erosion was more pronounced in the Yellowstone Valley, or Oligocene topography was such that sediment accumulation was negligible there.

The onset of deposition in a saline lake, represented by the Hepburn's Mesa Formation, marks the time that the Yellowstone Valley became a closed, internally drained basin: ca. 16–17 Ma. To form an internally drained basin, the valley floor had to have dropped relative to what eventually became the Beartooth and Gallatin Ranges. We therefore interpret deposition of the Hepburn's Mesa Formation to signal the beginning of Neogene extension that slightly later became so clearly manifested in the Deep Creek–Luccock Park fault system. Subsidence continued for >1.2 m.y. The relative homogeneity of sediment deposited during this time indicates little change in tectonic style or environment.

In Jackson Hole, rates of Barstovian subsidence were faster than in the Yellowstone Valley and continued a local trend that had been going on throughout the Miocene. Jackson Hole, however, resembled the Yellowstone Valley in first reflecting extensional tectonism between 18 Ma and 13 Ma, and in faster subsidence during the Barstovian than during the Arikareean through early Hemingfordian. The Barstovian part of the section in Jackson Hole maximally spans 5 m.y. (probably substantially less) and is 1,000 m thick, whereas the Arikareean through early Hemingfordian part of the section represents a minimum of 6 m.y. and is only 500 m thick.

An interruption in the relatively constant rate of Barstovian basin subsidence is marked by an erosional contact of the Hepburn's Mesa Formation with overlying coarse conglomerates and by the angular unconformity between the Colter and Teewinot Formations. In the Yellowstone Valley, the coarse conglomerates signify uplift of the Yellowstone Park area relative to the valley, which increased stream gradients and triggered external drainage. Initially this caused erosion at the top of the saline-lake deposits and then deposition of conglomerate as the basin continued subsiding. The chronology of events is similar in Jackson Hole, where an angular unconformity marks erosion with contemporaneous westward tilting of the Colter Formation. Then rapid subsidence commenced as more than 1,800 m of Teewinot Formation accumulated (Love and others, 1973; Barnosky, 1984). These events presaged uplift of the Teton Range, which began no earlier than 9.4 Ma (Love and others, 1973).

Intermontane basins to the north and west of the Greater Yellowstone region also record post-Barstovian erosion, uplift, and establishment of external drainage (Fields and others, 1985). The timing of this deformational and erosional event, however, is less well constrained in most of these basins than it is in the Greater Yellowstone region, where it is bracketed between late Barstovian (ca. 11.5–14.8 Ma) and 9.4 Ma in Jackson Hole (A. D. Barnosky, 1984) and between late Barstovian and 8.6 Ma (date from Chadwick, 1982) in the Yellowstone Valley.

CONCLUSIONS

The Hepburn's Mesa Formation indicates that a perennial saline lake and associated mudflats occupied the Yellowstone Valley from >16.0 Ma to <14.8 Ma. Presence of a saline lake implies a climate more arid than today's in the Greater Yellowstone region. Dominance of small fossil mammals adapted to arid or semiarid environments from both the Yellowstone Valley and Jackson Hole corroborates this conclusion.

Extensional tectonism was underway in the Greater Yellowstone region by 16.0 Ma, when subsidence led to development of closed drainage in the Yellowstone Valley and the rate of subsidence increased in Jackson Hole. The earliest phases of extension throughout southwestern Montana may be signaled by the mid-Tertiary unconformity (ca. 17–18 Ma), which is particularly evident in the Yellowstone Valley by the absence of Oligocene through early Miocene rocks.

In Jackson Hole, the onset of extensional tectonism also is marked by a change in the nature of tuffs that were deposited on the flanks of local volcanoes. Trachyte, latite, and andesite in Hemingfordian and older deposits gave way to mainly calc-alkaline rhyolite in Barstovian and younger ones. The Jackson Hole volcanoes probably supplied some of the tuffaceous material so common in coeval deposits of nearby intermontane basins.

Between 16.0 and 14.8 m.y. in the Greater Yellowstone region, extensional tectonism was characterized by rates of basin subsidence that were slower in the Yellowstone Valley than in Jackson Hole, but relatively constant within each basin. After ca. 13–14.8 Ma, but before 8.6 Ma in the Yellowstone Valley and 9.4 Ma in Jackson Hole, uplift in the Yellowstone Park area was coupled with intrabasinal erosion and tilting. Then subsidence of basins continued at a faster rate.

A pattern of basin subsidence accompanied by extensional tectonism reaches from the Greater Yellowstone region through numerous intermontane basins in southwestern Montana and adjacent Idaho. Moreover, the timing (ca. 14–17 Ma) of this extensional deformation, and also of volcanism in Jackson Hole, coincides with (1) volcanism and initial extensional breakup of the basins and ranges in the central Great Basin and (2) extrusion of vast quantities of basaltic lava on the Columbia Plateau. This coincidence suggests that middle Miocene tectonism in the Greater Yellowstone area was probably a response to regional as well as to local tensional stresses, as can be accommodated in plate-interaction structural models (Christiansen and McKee, 1978; Eaton, 1979).

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