2003. Proceedings of the Indiana Academy of Science 112(2):97–108

WESLEY CHAPEL GULF REVISITED: GEOMORPHIC PROCESSES AFFECTING DEVELOPMENT OF THE ALLUVIAL FLOOR OF A KARST GULF IN ORANGE COUNTY, INDIANA

James M. Durbin, **Ryan C. Jones**, **Brendan Anslinger**, and **Danise Guy**: Department of Geology, University of Southern Indiana, 8600 University Blvd. Evansville, Indiana 47712 USA

ABSTRACT. Wesley Chapel Gulf is a karst collapse depression with an alluvial floor that was deposited as a subterranean passage of the Lost River rose to the surface (rise) and flooded the bottom of the structure. This paper reports on the thickness of floodplain sediments, the sedimentological character of fluvial deposits, and the generalized stratigraphic framework for the alluvial floor. Two stratigraphic units were identified from sediment cores, each with a fining upward trend. Unit 1 extends from −10.8 m to −3.5 m, and exhibits thin (1–5 mm) varve-like bedding, with interbedded clayey silt, sand, and occasional organic-rich silt. Thin bedding and a lack of well-developed paleosols indicate that frequent flooding early in the gulf's history allowed floodplain deposition of Unit 1 to outpace pedogenesis. The contact between Units 1 and 2 is marked by a distinctive color transition and a change in sediment deposition. Unit 2 extends from −3.5 m to the surface, and is a sandy to silty loam, with silt beds and some organic rich zones. Minor paleosols (entisols) in Unit 2 indicate short-lived depositional hiatuses linked to rising floor elevation due to vertical accretion of the floodplain that allowed pedogenesis to outpace deposition. Sediments in both units become finer-grained away from the rise, indicating that the Lost River rise is the dominant source of floodplain alluvium. Stratigraphic and sedimentologic evidence indicate a shift in sedimentation style, attributable to climatic and/or anthropogenic influences. Coarse sediments dominant at the base of Unit 1 suggest a linkage to climatic transitions during the terminal Pleistocene through the Holocene. The abrupt transition from Unit 1 to Unit 2 indicates a change in flood stage and sediment load of the gulf rise a probable response to anthropogenic modification of the Lost River drainage.

Keywords: Karst gulf, alluvium, stratigraphy, sedimentation, paleosol, climate change

Wesley Chapel Gulf, called Elrod Gulf on the French Lick 7.5 minute topographic quadrangle, is located approximately 7.2 km (4.5 mi.) southwest of Orleans, Indiana in Orange County (Fig. 1). While mapping the Lost River karst drainage system, Clyde A. Malott (1922, 1932, 1952) studied the caves adjacent to Wesley Chapel Gulf in detail, and formally defined a karst gulf as "a karst feature . . ." that is "a steep walled depression that characteristically has an alluviated floor and that always contains a stream that rises and sinks within it" (1932). Based on the planview shape of the gulf and the steep walls, he hypothesized that the gulf formed due to coalescing sinkholes created as caves collapsed (1932). Malott (1952) hypothesized that the karst spring in the depression at the southern end of Wesley Chapel Gulf (known as Boiling Spring or Wesley Chapel Gulf rise – Fig. 2) connected to the Lost River system, in part because of the extent of the subterranean stream system he documented while mapping the local cave system. Powell (1987) included Wesley Chapel Gulf in an investigation of the hydrology of the Mitchell Plain karst system. The thickness of alluvium has never been measured directly. Measurements at the rise pit by Malott (1932, 1952) were used to estimate that the alluvial floor of the gulf was between 10.6 m (35 ft) and 12.2 m (40 ft) thick.

Alluvium in Wesley Chapel Gulf provides information on the nature and effect of climatic and anthropogenic influences on the Mitchell Plain in that the fill represents a partial record of subterranean Lost River discharge and sediment load. Paleosols in the floodplain provide insight into the frequency of flooding, the nature of sedimentation, and the degree of pedogenesis that occurred on the floor between depositional events.

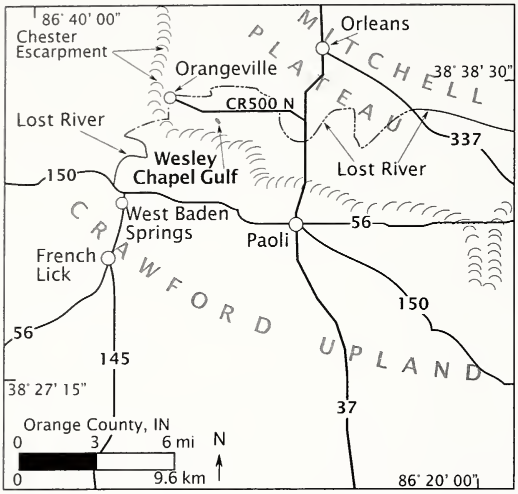


Figure 1.—Location map of Wesley Chapel Gulf, Orange County, Indiana, with respect to the physiographic provinces. Map modified from Gray et al. (2000).

Our research had five goals: 1) determine the thickness of the alluvial fill; 2) describe the stratigraphy and sedimentology of the sediments; 3) determine a relative age sequence for the alluvium; 4) interpret the geomorphic processes and causative factors that allowed the Gulf to develop its present morphology; and 5) hypothesize about the rates of sedimentation and timing of deposition.

BACKGROUND

Wesley Chapel Gulf lies in the Mitchell Plateau (formerly the Mitchell Plain of Malott [1922]) physiographic province, a non-glaciated and karstified region developed on Mississippian limestone bedrock that has numerous sinkholes, blind valleys, swallow holes, and caves (Gray 2000). The gulf formed by dissolution and collapse of the Ste. Genevieve Limestone (Malott 1932), a high calcite (>95% $CaCO_3$) unit in the Blue River Group (Carr et al. 1978). Water in Wesley Chapel Gulf rise probably emanates from a complex series of underground conduits, similar to those mapped by Malott (1932). Dye tracing studies conducted along sinking streams near the gulf link subsurface flow from the Lost River eastward to the Wesley Chapel Gulf rise, and eventually to the Rise of the Lost River (Murdock & Powell 1968; Bassett & Ruhe 1973; Bassett 1974; Ruhe 1977; Powell 1987). The contributions of meteoric surface water, stream water, and subterranean flow to the karst drainage network is complex and the subject of continued research (Bayless et al. 1994; Lee & Krothe 2001). Weathering and erosion of local bedrock and loess east of Wesley Chapel Gulf are the logical source of alluvium filling the gulf (Ruhe 1977; Ruhe & Olson 1980; Wingard 1984).

The study area is surrounded by farmland, and there has been limited construction. However, clearing of forested tracts continues in the region; and new construction and agriculture in the Lost River basin increase the potential for erosion of unconsolidated material, and subsequent deposition in the gulf. Anecdotal accounts by Malott (1932) indicate the gulf floor was cleared of trees and shrubs for growing crops and grazing cattle as early as 1915, although a few larger trees estimated to be 75–100 years old are still present on the gulf floor. No crop farming has occurred on the gulf floor floodplain for at least the last 25 years, and probably longer, but pasture was maintained from at least 1983–1996 when ownership was turned over to Hoosier National Forest (Forest Service website). Under Forest Service jurisdiction, the gulf is reverting to an uncultivated state, with thick herbaceous vegetation and small woody plants growing on the floodplain.

METHODS

Sediment cores were collected with a Giddings soils probe at two locations (Fig. 2) along the long axis of the gulf to determine floodplain stratigraphy and the relationship between the rise and alluvium particle size. Cores were sited as far from the gulf walls as possible to limit the influence of surface streams and colluvium. Cores were collected from the following sites: A− 76.2 m (~250 ft) from the rise, and B− 15.2 m (~50 ft) from the rise (Fig. 2). Core holes were backfilled with sand and capped with a 0.6 m (2 ft) bentonite plug.

Sediments were described and analyzed for stratigraphic contacts, paleosols, grain-size trends, bedding features, and mineralogy in the Geomorphology and Hydrogeology lab at the University of Southern Indiana (Figs. 3, 4; Table 1). Standard techniques were used to describe soils in the cores (Birkeland 1999). Sediment samples representative of each identified layer or paleosol horizon from both

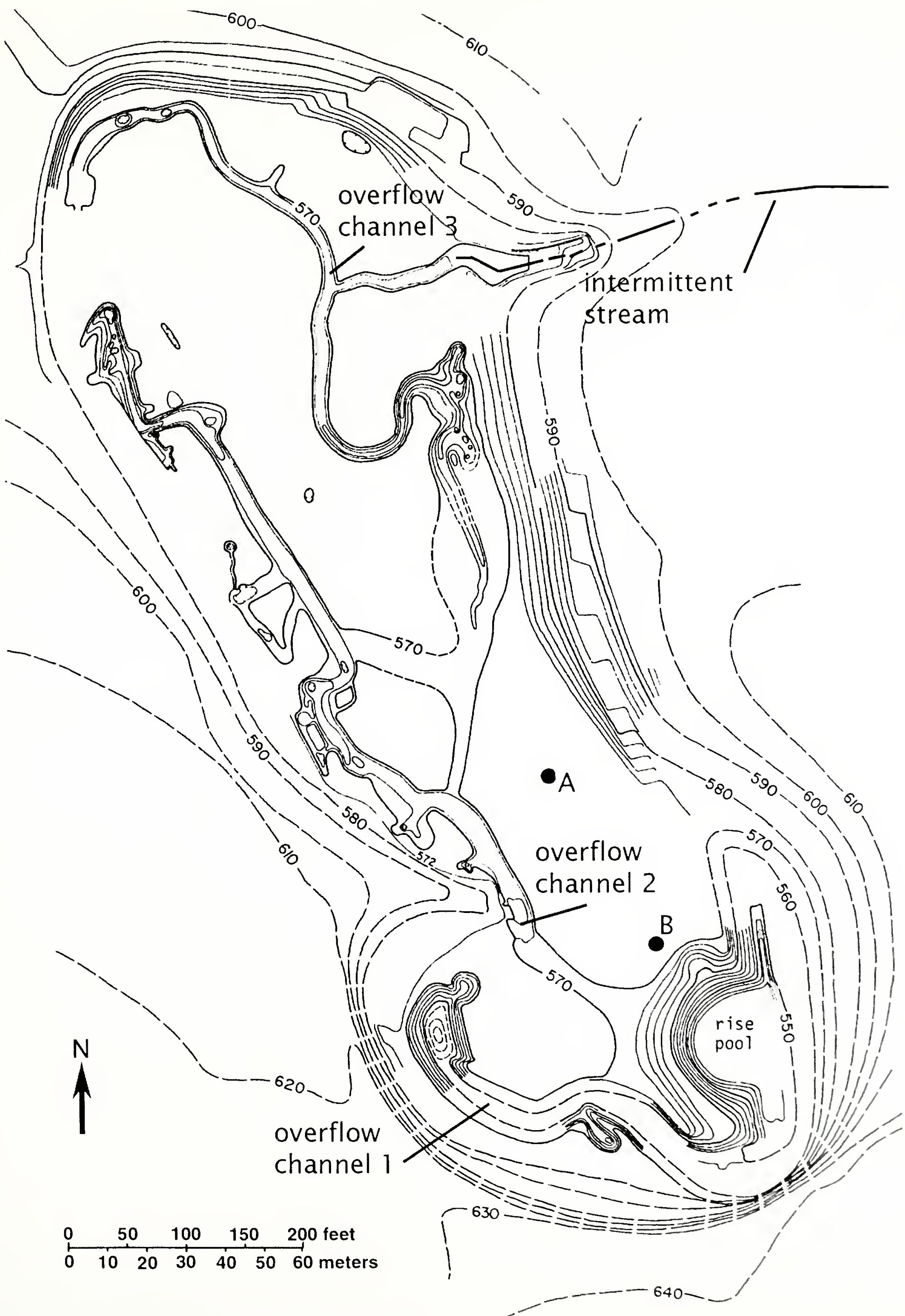


Figure 2.—Topographic map of Wesley Chapel Gulf showing the location of core extraction Sites A and B with respect to other gulf features. Plane table and alidade survey map by R.L. Powell, S.D. Maegerlein & J.L. Basset, 1970; modified from Powell 1981.

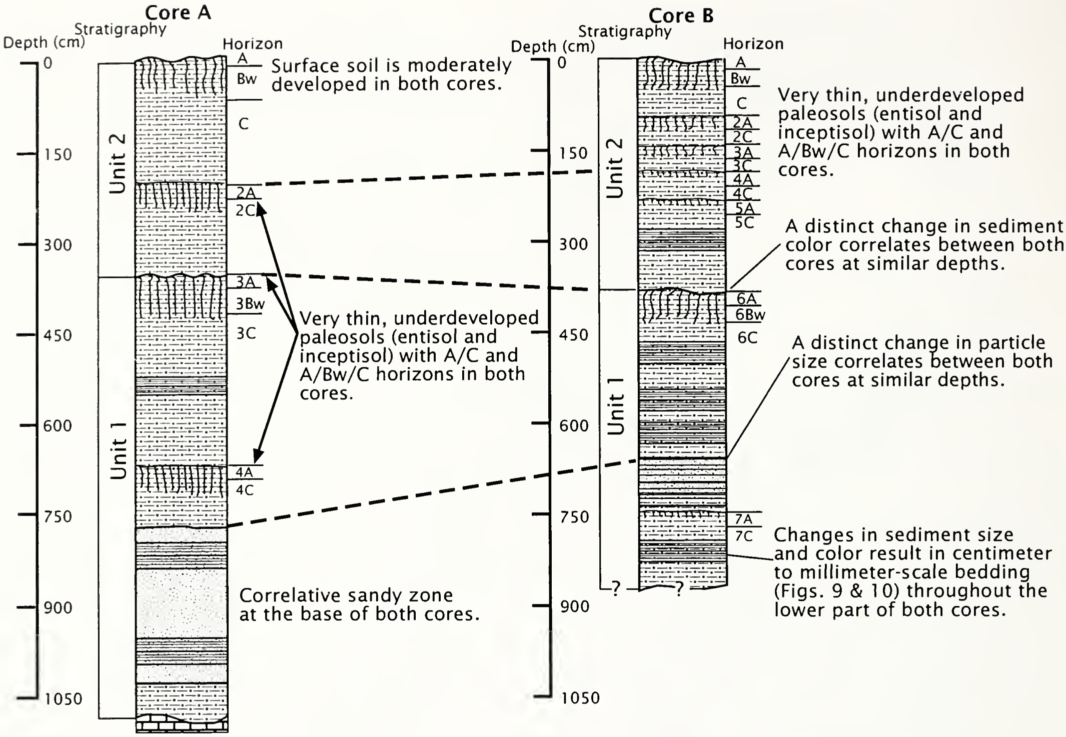


Figure 3.—Diagram showing floodplain stratigraphy and bedding, and soils developed in alluvium from Wesley Chapel Gulf. Dashed lines represent correlative layers between the two cores.

cores were collected and processed for particle size using a standard pipette method (Singer & Janitzky 1986). All particle size data were tabulated and plotted versus depth to identify trends and correlate soil descriptions between cores (Fig. 4; Table 1). Mean particle size percentage was determined for each core and for the two stratigraphic units in the cores by summing the values for the sampled intervals and dividing by the number of intervals. Bedding features and mineralogy were determined by visual inspection using a hand lens and binocular microscope.

OBSERVATIONS

The floodplain has limited topographic relief (Fig. 2) except near the rise and where overflow channels have formed around the perimeter of the gulf floor (Figs. 2, 5, 6). Previous observations of the gulf during floods suggested that muddy storm water resurging from the rise was a source of fine-grained sediment on the floor (Figs. 5, 6), although perhaps not the only source. An intermittent stream enters the gulf at the northern end, where the gulf has expanded along the thalweg of a shallow valley and has cut a channel into the bedrock rim (Fig. 2). Limited amounts of sediment, consisting mostly of occasional broken pieces of limestone and rare chert pebbles, exist in the channel, although fine-grained sediment from nearby fields is transported during heavy rainfall (Powell pers. comm. 2003). The gulf floor has little in the way of material larger than sand size except where rare, partially buried limestone boulders have fallen from the gulf walls. Floodplain sediments exposed at the surface close to the rise are coarser than at locations farther away from the rise (Figs. 7, 8).

Core refusal at Site A occurred at −10.8 m (−35 ft) when the push tube encountered limestone interpreted as bedrock or collapse rubble. Coring at Site B penetrated 10.3 m (34 ft), but no sediment was recovered from −8.8 to −10.3 m (~29 to 34 ft). Coring was terminated at 10.3 m (34 ft) due to a lack of sediment recovery when either the saturated material flowed out of the tube upon extrac-

Table 1.—Particle size data from sediment cores at Sites A and B, Wesley Chapel Gulf.

	Depth		Percent particle size by weight				Depth		Percent particle size by weight		
Site A	cm	in	% clay	% silt	% sand	Site B	cm	in	% clay	% silt	% sand
Unit 2	0	0	12	64	24	Unit 2	0	0	14	52	34
	3	1	18	52	30		—	—	—	—	—
	25	10	20	75	5		25	10	20	68	12
	152	60	17	62	21		152	60	19	69	12
	234	92	22	73	5		234	92	16	66	18
	262	103	18	69	13		—	—	—	—	—
	292	115	20	65	15		292	115	15	61	24
	350	138	20	75	5		350	138	15	59	26
Mean			19	76	15	Mean			16	62	21
Unit 1	386	152	19	68	13	Unit 1	—	—	—	—	—
	488	192	19	66	15		488	192	16	54	30
	541	213	18	72	10		—	—	—	—	—
	549	216	23	66	11		—	—	—	—	—
	597	235	18	73	9		597	235	14	62	24
	625	246	23	58	19		615	242	17	56	27
	670	264	21	68	11		678	267	2	37	61
	595	274	21	72	7		696	275	5	27	68
	777	306	16	36	48		777	306	54	18	28
	869	342	14	26	60		869	343	5	42	53
	939	370	17	36	47		—	—	—	—	—
	975	384	31	56	13		—	—	—	—	—
	984	388	38	61	1		—	—	—	—	—
	998	393	22	74	4		—	—	—	—	—
	1017	400	20	70	10		—	—	—	—	—
	1025	403	10	83	7		—	—	—	—	—
	1036	408	11	82	7		—	—	—	—	—
Mean			20	63	17	Mean			16	42	42
Mean	Site A		19	69	16	Mean	Site B		16	52	31

tion, and/or sediments flowed around the tube during pushing and never filled the coring device.

Two distinctive unconsolidated units were identified based on color and sedimentation style: a lower unit (Unit 1), and an upper unit (Unit 2). Unit 1, a 7 m (23 ft) thick, dark colored sandy silt, possesses distinctive intervals in which bedding is preserved. Coring disrupted the original morphology of the layers somewhat; however, bedding appears horizontal and thin, ranging from 0.5 mm to 2 cm thick and averaging about 1 mm. Unit 1 bedding in both cores had alternating light and dark colored bands (Figs. 9, 10), with lighter bands commonly consisting of slightly coarser material (fine sand/coarse silt) and darker bands composed of finer particles (fine silt and clay) and in some instances organic materials. Sediments throughout the unit lacked calcareous materials, but they did have some iron and manganese oxides in the darker layers.

Unit 2, by contrast, is a lighter colored sandy silt about 3.5 m (11.5 ft) thick, with rarely preserved bedding and minor paleosols at several intervals. Unit 2 also lacked calcareous materials, but contained zones of abundant mottled sediments. Poorly developed, thin paleosols (entisols) consisted of an A-horizon with weak soil structure (peds) overlying an oxidized C-horizon with no intervening B-horizon. A-C-horizonation is commonly associated with entisols, a soil taxonomic classification that denotes limited development owing to, among other things, limited time of subaerial exposure in the soil-forming environment (Birkeland 1999). The surface soil consisted of an A-Bw-C-horizon sequence, characterized as an inceptisol, that also suggests limited soil development.

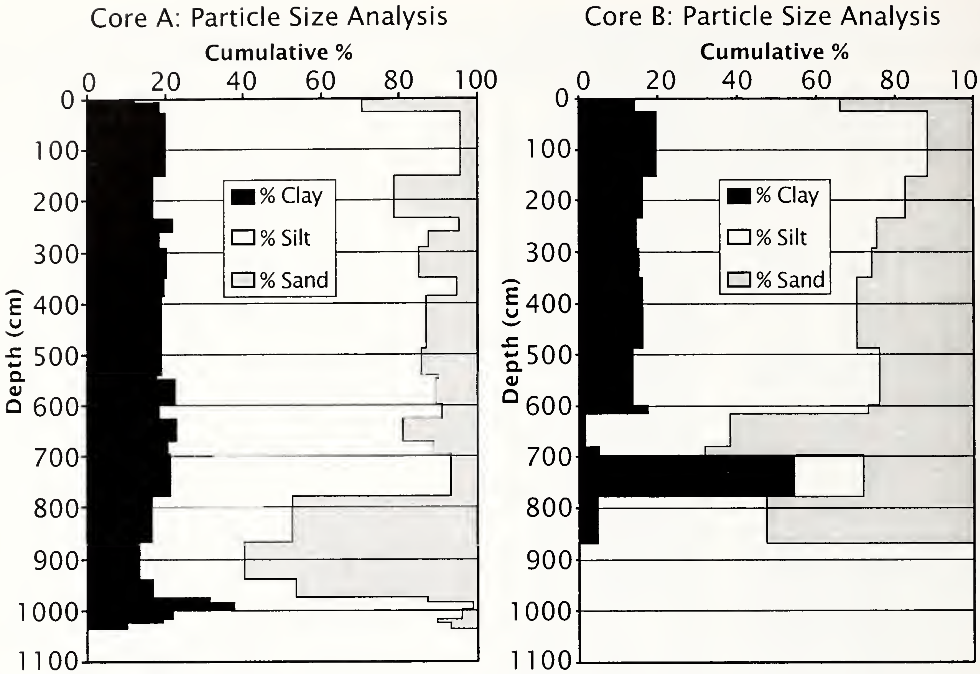


Figure 4.—Particle size analysis for cores from Sites A and B. Plots show cumulative percent of sand, silt, and clay size fractions *vs.* depth. Black is percent clay, gray is percent sand, and white is percent silt as determined by standard pipette method (Singer 1986).

Particle size data from Unit 1 varied between cores (Table 1), with silt dominating at Site A (mean silt percentage of 63%) and mean particle size at Site B split equally between sand and silt, with values of 42%. In both cores, Unit 1 showed greater sand content at Site A when compared to Site B. The lower 2.1 m (7 ft) of Unit 1 contains a higher percentage of sand than the upper part at both core sites.

Unit 2 particle size data showed a trend (Fig. 4; Table 1) similar to that of Unit 1, with mean silt-sized fractions comprising 76% at Site A, and 62.5% at Site B. Sand content ranged from 15% at Site A, to 21% at Site B. When comparing mean particle size percentages for the entire core, silt dominates the alluvium at Site A, with a mean percentage of 69.5% of the total sediment; sand and clay were 16% and 19.5%, respectively.

Although silt (52%) also dominates alluvium at Site B, the mean percent particle size is greater than at Site A, as sand is 32% and clay is 16% (Fig. 4; Table 1). Sediments collected from both units at Site A were finer-grained compared to the coarser-grained sediments in both units at Site B (Fig. 4; Table 1), confirming that surface sediments become finer-grained with distance away from the rise.

INTERPRETATIONS AND DISCUSSION

Lack of core barrel refusal at Site B indicates alluvium close to the rise is greater than 10.3 m (34 ft), which is consistent with estimates calculated from earlier measurements by Malott (1932, 1952). At Site A, core barrel refusal at 10.3 m (34 ft) depth was interpreted as bedrock or collapse rubble. Attempts to auger deeper were unsuccessful and turned up limestone fragments. The alluvium is thinner at Site A, probably due to irregular topography of the underlying bedrock or collapse rubble created as the gulf formed. Relatively lesser amounts of collapse debris should occupy locations where dissolution created pre-collapse caves in the limestone. The resultant collapse-created surface allowed deposition of thicker alluvium at some locations, and thinner amounts of fill in other locations. It is not known how long it would have taken the underground system to begin flooding into the

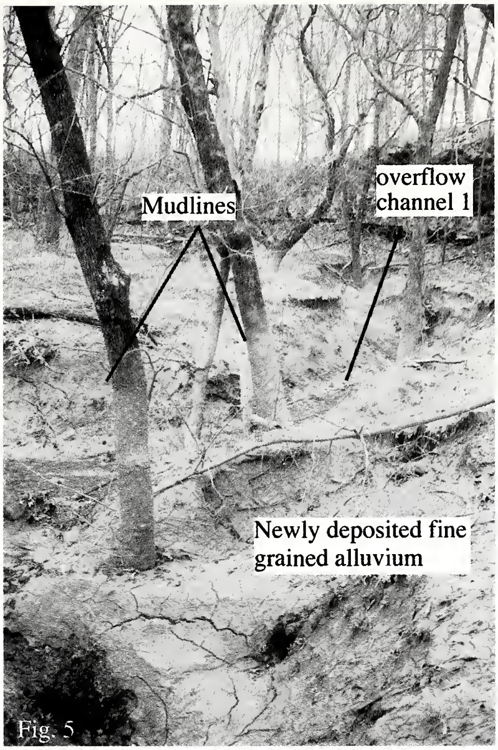


Figure 5.—Photograph of Wesley Chapel Gulf after a flood in February 1986 showing the overflow channels (foreground), documenting the stage of the flooding (mudlines on tree trunks), and illustrating the nature of fine-grained sediment deposition on the floodplain. Note the lack of thick vegetation on the floodplain, as the property owners routinely maintained floor at the time of the photos. Photo courtesy of Dr. Paul Doss.

gulf, or if pre-collapse loess and/or residuum contributed to the gulf alluvium.

As the intermittent stream cut down into the bedrock wall on the northeast side of the gulf (Fig. 2), it created a series of bedding plane benches and a small waterfall onto the gulf floor. Channel modification in the intermittent stream occurs primarily by dissolution, creating a dissolved sediment load. However, during heavy rainfall, the stream also carries suspended sediments. Minor amounts of sediment are stored in or along the bedrock channel indicating that most of the suspended sediment is transported into the gulf. The lack of an alluvial fan or fan delta at the base of the waterfall suggests material transported into the gulf via this stream has been carried into the subsurface via one of many sinkholes in the overflow channel, or is redistributed over the floodplain by flood waters associated with the gulf rise. Although no cores were collected from the north end of the gulf, both depositional scenarios likely occur under specific circumstances. The surface stream discharge occurs more frequently than does overflow flooding in the gulf (Powell pers. comm. 2003). Discharge in the intermittent stream does not always create a waterfall, as some of the flow is diverted into fractures before reaching the rim of the gulf (R. Armstrong pers. comm. 2003). When discharge exceeds the capacity of the fractures to divert the water, the flow over the waterfall is directed into the overflow channel 3 at the north end of the gulf where it is then diverted underground through a series of swallow holes (Fig. 2). In order to vertically-accrete sediments on the floodplain, surface stream discharge must exceed the capacity of the overflow channel swallow holes to carry the water. Overbank flooding attributable to the intermittent stream does occur, but less frequently than flow in the stream. When the gulf rise floods the overflow channels, water from the surface stream flows into the already flooded gulf, where suspended sediments are probably redistributed as a floodplain rather than a delta or fan.

The few boulders partly buried along the perimeter of the gulf floor indicate that rock falls contribute some debris to the gulf floor, although at the surface the amount of rock contributed to the alluvium via rock falls appears minor relative to the area of alluvium exposed on the gulf floor. However, the total volumetric contribution of boulders from rock falls and collapse events is beyond the scope of this paper.

The decrease in particle size with distance away from the rise is attributed to a decrease in floodwater energy. During flooding, water flowing up through the rise transports sediment as coarse as granule size (2–4 mm) and deposits the particles on the margin of the rise pit adjacent to the floor (Figs. 7, 8). When flood stage fills the rise pit, the water flows first into overflow channel 1 (Fig. 2) in the far southwest side of the gulf until the swallow holes can no longer divert the volume of water discharging into the channel. As the stage rises higher, it overflows in to overflow channel 2 (Fig. 2). As the flood stage exceeds the swallow hole capacity and the elevation of the

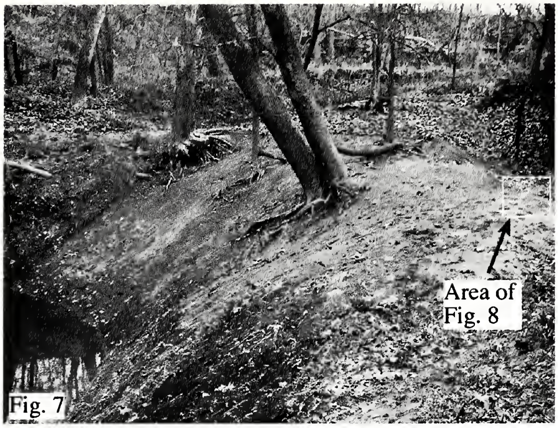
Figure 6.—Photograph of Wesley Chapel Gulf after a flood in February 1986 showing the mudlines on tree trunks. Note the lack of thick vegetation on the floodplain. Photo courtesy of Dr. Paul Doss.

overflow channel banks, it inundates the nearly horizontal floor of the gulf, flowing northwest as sheet flood (Powell pers. comm. 2003). When the water leaves the confines of the rise, energy levels drop, causing deposition on the gulf floor. The overflow water deposits coarse-grained sediments first, and then progressively finer-grained sediments with increased distance from the rise and correlative decrease in flow velocity as dictated by a Hjulström model (Hjulström 1939).

Couplets consisting of alternating bands of light and dark sediment at both sites are interpreted to be two parts of a single flood (Figs. 9, 10). A major flood in the gulf rises rapidly and has more energy than during the waning stages of the flood. During the maximum stage of flooding, the rise overflows the perimeter channels and the gulf becomes a steep-walled lake. As water levels fall, the overflow channels drain water on the floor back into swallow holes that line the overflow channels (Malott 1952). Coarser sediment deposited as a light band is transported more readily during the rising stage, whereas silt and clay-sized fractions that sometimes have a higher organic component are deposited as darker bands once the floodwater flow across the gulf abates.

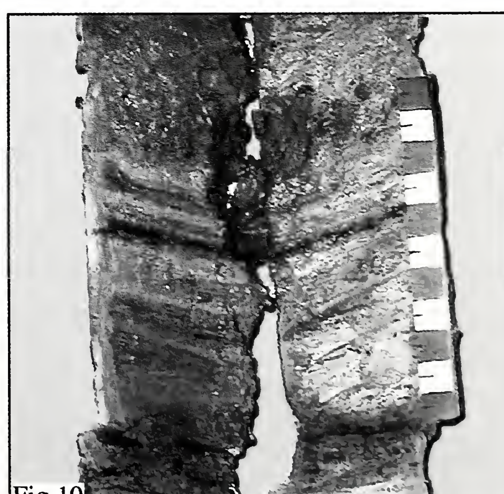
The lack of preserved bedding and the presence of paleosols in Unit 2 reflect a change in soil-forming conditions as compared to Unit 1, specifically the duration between depositional events. Pedogenesis acts as a homogenizing agent, disrupting original bedding features while creating soil horizons (Birkeland 1999). Original bedding is visible in Unit 1, which suggests sediment deposition outpaced soil development. Rates of pedogenesis were apparently more rapid or sedimentation rates were slower during deposition of Unit 2. Increased time between flooding should be expected as the elevation of the gulf floor increased due to sediment deposition and or incision occurred due to local base level lowering.

Correlation to climate and human influences.—The high percentage of coarse sediments at −7 m in both cores (Fig. 4) may represent climate change. Although pollen records or other climate proxy data for the region are scarce, some general statements can be made concerning past environmental conditions. Shortly after the Wisconsinan glacial maximum, climates in the midwestern U.S. were cold, with available moisture similar to modern levels but likely locked up as ice (Williams 1974; Bartlein et al. 1984). During the terminal Pleistocene conditions were still cool but warming, causing ice and snow to



Figures 7, 8.—Photographs of sediments proximal to the rise in Wesley Chapel Gulf. Figure 7 shows the morphology of sediments deposited by water flowing out of the rise. Figure 8 is a close up of sediments in the area indicated in Fig. 7. Pocket knife shown for scale (13 of 20 cm length exposed).

thaw and become more active again in the hydrologic cycle. By the early Holocene, conditions were warm, with fluctuating moisture levels from drier-than-modern to modern moisture conditions (Williams 1974; Holloway & Bryant 1985; Baker et al. 1992). The transitions from climates with cooler temperatures and some moisture to those with warmer temperatures and less or more effective moisture affected the vegetation (Williams 1974; Bartlein et al. 1984; Holloway & Bryant 1985; Baker et al. 1992), and altered discharge regimes and sediment loads for the karst system. Transitions from glacial climates and cold-tolerant arboreal vegetation to warmer climates with non-arboreal hardwood forests and grasses have been shown to increase soil development (Birkeland 1999), as well as sediment supply and discharge in streams (Langbein & Schumm 1958). The increased erosion in the Lost River basin and sediment transport initially favored deposition of coarser material on the gulf. However, with time the alluvial gulf floor increased in thickness



Figures 9, 10.—Photographs of small intervals of Core A (Fig. 9) and Core B (Fig. 10) taken from depths of 560 cm (~18 ft) and 457 cm (~15 ft), respectively. Sediments show distinctive bedding, accentuated by color differences between the various layers. Color changes are due largely to differences in particle size and variability in organic carbon content. Each square on the ruler (right) is 1 cm.

making deposition of coarser materials on the floodplain more difficult due to diminished amounts of coarse sediment and the increased gulf floor elevation.

The modern climate of relatively mild wet winters and warm humid summers favored changes in chemical and physical weathering and in vegetation patterns, allowing lush hardwood forests to develop (Williams 1974; Bryant & Holloway 1985). The transition from humid to dry climates in the Holocene would have facilitated stripping of soils developed on the limestone bedrock during the wetter periods. As the landscape reached equilibrium with modern environmental conditions, coarse-grained materials that were already limited due to the stripped sandstone caprock of the overlying Pennsylvanian system of the Crawford Upland and the soluble Mississippian limestone bedrock of the Mitchell Plateau became scarcer. Increased proportions of finer-grained sediments from sources such as loess and terra rosa (insoluble limestone residuum) became the dominant sources of sediment entering the Lost River drainage network. The transition from sandy to silt-dominated alluvium in Unit 1 represents the likely response to the transition from the mid-Holocene warm and dry interval to modern warm humid climate.

Deforestation and tillage of the landscape beginning in the 1800's provided ample sources of fine-grained colluvium and soil that were transported into the surface and subterranean drainage networks. The shift in color and bedding characteristics observed in both cores at the contact between Units 1 and 2 may represent the influence of widespread deforestation and agricultural development of the region. The gulf floor was used sporadically for agricultural and grazing purposes during the 1900's through the 1990's, until the U.S. Forest Service took possession of the property in 1996. The presence of finer-grained sediments, the frequency of paleosol occurrence and the degree of paleosol development in the upper 3 m (10 ft) of the cores supports an anthropogenic influence on erosion and sedimentation in Unit 2 similar to that commonly observed in other drainage networks (Orbock-Miller et al. 1993; Taylor & Lewin 1997). Material deposited on the floodplain of the gulf meant future floods needed to attain higher stages to get out of the rise pit and overflow channels and onto the floodplain. Floods of sufficient magnitude to inundate the gulf floor would have occurred less frequently with each successive depositional event. Physical records of gulf flooding are not available, and anecdotal accounts are incomplete and/or contradictory (R. Armstrong pers. comm. 2003; Powell pers. comm. 2003). Modern flooding of the gulf occurs sporadically in terms of frequency and time of year over which flooding occurs (R. Armstrong pers. comm. 2003). Most floods occur in the spring, although floods can occur at any time of year. Over the interval in which data were collected (October 2000 through March 2003), the gulf flooded only once (March 2003), despite several episodes of intense rainfall and flooding of surface streams. The infrequent nature of flooding supports formation of minor paleosols in the upper 3 m (10 ft) of sediment during depositional hiatuses, as soil formation outpaced deposition.

Given the record of late Pleistocene and Holocene climate change, an alternative explanation may be that soil-forming conditions intensified and diminished episodically in conjunction with sediment deposition, creating conditions whereby soils formed either more quickly or more slowly. However, the number of, and degree of development of minor paleosols in the cores supports the variable rates of deposition scenario. The variability in the number of paleosols exceeds the documented shifts in late Pleistocene through Holocene climates, and speed at which said climatic changes could manifest themselves in the pedogenic record.

Modern flooding of the gulf occurs sporadically in terms of frequency and time of year (R. Armstrong pers. comm. 2003), but most flooding occurs in the spring. Over the interval in which data were collected (October 2000 through March 2003), the gulf floor flooded only once (in the Spring of 2003). Anecdotal accounts prior to 2000 suggest that such long intervals between gulf floor flooding may not be typical (Powell pers. comm. 2003).

ACKNOWLEDGMENTS

We thank the Hoosier National Forest, in particular Jim Dencour and Cyd Morgan, for their permission to conduct research in the gulf. We also thank the Geology Department at the University of Southern Indiana for use of laboratory resources and the Giddings soil probe. Thanks for the personal communication in August 2003 with R. Armstrong, Chairman of the Lost River Conservation Association. Thanks, also, to Mr. Ron Counts and members of the Geomorphology class of 2000 for their assistance collecting sediment cores. Lastly, I wish to express my gratitude to colleagues at USI and the Indiana Academy of Science whose thoughtful commentary during the editorial process was invaluable.

LITERATURE CITED

Baker, R.G., L.J. Maher, C.A. Chumbley & K.L. Van Zant. 1992. Patterns of Holocene environmental change in the midwestern United States. Quaternary Research 37:379–389.

Bartlein, P.J., T. Webb III & E. Fleri. 1984. Holocene climatic change in the northern Midwest: Pollen derived estimates. Quaternary Research 22:361–374.

Bassett, J.L. 1974. Hydrology and geochemistry of karst terrain, upper Lost River drainage basin, Indiana. MS Thesis, Indiana University. 102 pp.

Bassett, J.L. & R.V. Ruhe. 1973. Fluvial geomorphology in karst terrain. *In* Fluvial Geomorphology. (M. Morisawa, ed.). Papers presented at the 4th annual symposium on Geomorphology, State University New York, Binghamton, New York.

Bayless, R.E., C.J. Taylor & M.S. Hopkins. 1994. Directions of groundwater flow and locations of ground water divides in the Lost River watershed near Orleans, Indiana. Water Resources Investigations Report 94–4195, Reston, Virginia. U.S. Geological Survey. 25 pp.

Birkeland, P.W. 1999. Soils and Geomorphology. Oxford University Press, New York. 432 pp.

Carr, D.D., R.K. Leininger & M.V. Golde. 1978. Crushed Stone Resources of the Blue River Group (Mississippian) of Indiana. Dept. of Natural Resources, Geological Survey Bulletin. Bloomington, Indiana. 225 pp.

Forest Service Website. 2000. http://www.fs.fed.us/r9/hoosier/docs/wesley_chapel.htm

Gray, H.H., C.H. Ault & S.J. Keller. 1987. Bedrock Geologic Map of Indiana. Indiana Geological Survey Miscellaneous Maps, MM48.

Gray, H.H. 2000. Physiographic Divisions of Indiana. Indiana Geological Survey Special Report, SR61. 15 pp.

Holloway, R.G. & V.M. Bryant, Jr. 1985. Late Quaternary pollen records and vegetational history of the Great Lakes region: United States and Canada. Pp. 205–245, *In* Pollen Records of Late-Quaternary North American Sediments. (R.G. Holloway & V.M. Bryant, Jr., eds.) American Association of Stratigraphic Palynologists.

Hjulström, F. 1939. Transportation of detritus by moving water. Pp. 5–31, *In* Recent Marine Sediments, a Symposium. (P.D. Trask, ed.). American Association Petroleum Geologists, Tulsa, Oklahoma.

Jackson, M.T. 1997. The Natural Heritage of Indiana. Indiana Department of Natural Resources and Indiana Academy of Science, Indiana University Press, Bloomington, Indiana. 482 pp.

Janitzky, P. 1986. Particle size analysis. Pp. 11–16, *In* Field and Laboratory Procedures Used in a Soil Chronosequence Study. (M.J. Singer & P. Janitzky, eds.) U.S. Geological Survey Bulletin 1648. Reston, Virginia.

Lee, E.S. & N.C. Krothe. 2001. A four-component mixing model for water in a karst terrain in south-central Indiana, USA using solute concentration and stable isotopes as tracers. *In* Hydrochemistry of Springs. (B.G. Katz & C.M. Wicks, eds.). Chemical Geology 179:129–143.

Langbein, W.B. & S.A. Schumm. 1958. Yield of sediment in relation to mean annual precipitation. Transactions of the American Geophysical Union 39:1076–1084.

Malott, C.A. 1922. The physiography of Indiana. Pp. 59–256, *In* Handbook of Indiana Geology. (W.N. Logan, E.R. Cummings, C.A. Malott. S.S. Visher, W.M. Tucker & J.R. Reeves, eds.) Indiana Dept. of Conservation, Division of Geology, Spec. Publ. No. 21.

Malott, C.A. 1932. Lost River at Wesley Chapel Gulf. Proceedings of the Indiana Academy of Science 41:285–316.

Malott, C.A. 1952, The Swallow Holes of the Lost River, Orange County, Indiana. Proceedings of the Indiana Academy of Science 61:187–231.

Murdock, S.H. & R.L. Powell. 1968. Subterranean drainage routes of Lost River, Orange County, Indiana. Proceedings of the Indiana Academy of Science 77:250–255.

Orbock-Miller, S., D.F. Ritter, R.C. Kochel & J.R. Miller. 1993. Fluvial responses to land-use changes and climatic variations within the Drury Creek watershed, southern Illinois. Geomorphology 6:309–329.

Powell, R.L. 1981. Karst lands field trip of the Ecological Society of America. Field trip guidebook prepared for the American Institute of Biological Sciences. Geosciences Research Associates, Inc., Bloomington, Indiana.

Powell, R.L. 1987. The Orangeville Rise and Lost River, Indiana. Pp. 375–380, *In* North-Central Section of the Geological Society of America, Centennial Field Guide, Vol. 3, (D.L. Biggs, ed.), Geol. Soc. of America, Boulder, Colorado.

Ruhe, R.V. & C.G. Olson. 1980. The origin of terra rosa in the karst of southern Indiana. Pp. 84–190, *In* Field Trips 1980 from the Indiana University Campus, 14th Annual Meeting of the North Central Section, Geological society of America (R.H. Shaver, ed.).

Ruhe, R.V. 1977. Summary of geohydrologic relationships in the Lost River watershed, Indiana, applied to water use and Environment. Pp. 64–78, *In* Hydrologic Problems in Karst Regions (R.R. Dilamarter & S.C. Csallany, eds.), Western Kentucky University, Bowling Green, Kentucky.

Sengupta, S. 1979. Grain-size distribution of suspended load in relation to bed materials and flow velocity. Sedimentology 26:63–82.

Singer, M.J. & Janitzky, P. (eds.). 1986. Field and Laboratory Procedures Used in a Soil Chronosequence Study. U.S. Geological Survey Bulletin 1648. 48 pp.

Taylor, M.P. & J. Lewin. 1997. Non-synchronous response of adjacent floodplain systems to Holocene environmental change. Geomorphology 18:251–264.

Williams, A.S. 1974. Late-glacial postglacial vegetational history of the Pretty Lake region, northeastern Indiana. U.S. Geological Survey Prof. Paper 686–B. 23 pp.

Wingard, R.C. 1984. Soil Survey of Orange County Indiana. Soil Conservation Service, U.S. Dept. of Agriculture. 124 pp.

Manuscript received 9 April 2003, revised 3 September 2003.