James Hedges *
Hyattsville, Md.

EXPANDED JOINTS AND OTHER PERIGLACIAL PHENOMENA ALONG THE NIAGARA ESCARPMENT

Abstract

The crest of the Niagara escarpment from Fayette county, Iowa southeastward into Carroll county, Illinois is dotted with small sinkholes. The face of the escarpment is mantled with cryoturbates and the foot of it locally is fringed by solifluction aprons.

Those of the sinkholes which are open give access to deep, narrow fissures partly filled with pieces of shattered dolomite. The fissures were produced by the development of ice wedges in joint openings parallel with the local face of the escarpment and the resultant displacement of joint-bounded dolomite blocks outward across the top of the Maquoketa shale.

Several feet of broken, disturbed dolomite are visible above the bedrock and below the soil in many exposures along the Niagara escarpment. In some places, this "head" has moved downslope by solifluction and has accumulated at the base of the escarpment as terrace-like jumbles of blocky material. The finer constituents occasionally have been piped out of these solifluction aprons, leaving small, cavernous openings among the blocks.

Periglacial phenomena in Wisconsin are restricted to drifts of pre-Valderan ages; they are common in areas invaded by Rockian ice. This, together with the presence of Woodfordian drifts on three sides of the Niagara Escarpment area in Iowa, Wisconsin, and Illinois, suggests that the expanded joints and other periglacial phenomena developed in Woodfordian time.

INTRODUCTION

For many years, cave explorers have been aware of the numerous small openings which occur along the crest of the Niagara escarpment in Iowa and in adjacent portions of Illinois and Wisconsin. These fissures rarely extend more than 100 feet, laterally, between the furthest limits of human accessibility. They tend to be drab in color and to be barren of travertine deposits. Many, however, are deeper and of greater technical difficulty than are most other caves of the Driftless Hill Land. Consequently, they have attracted the attention of persons interested in underground mountaineering.

The internal morphology of these openings is conspicuously different from that of most other caves in the region. High, narrow, rubble-choked, they rarely contain evidence either of solutional erosion or of solutional

* 8218 Sherrill, Hyattsville, Md., U.S.A.

deposition. Howard (1963, pp. 63–64), in the only formal discussion of them which has thus far been published, sought to interpret them as portions of a dynamic karst landscape. He postulated that rapid groundwater circulation from the loess- and drift-mantled, gently-rolling Niagara cuesta towards the strongly-dissected, retreating face of the Niagara escarpment leads to their ongoing, non-cyclic development. This "stream piracy" theory, that cavern development occurs as a result of locally steep water-table gradients brought about by the (geologically) sudden shifting of surface drainage divides, is set forth in detail by Woodward (1961).

For a number of reasons, Howard's solutional hypothesis fails satisfactorily to provide an explanation of the development of the fissures. Nor can a solutional hypothesis encompass the minor stratigraphic and relief features associated with the fissures along the Niagara escarpment. The ubiquitous mantle of "head", the solifluction aprons with their piped-out, cavernous interiors, the thermoclastic talus of the retreatal gorges, all these and the fissures, too, cry out their allegiance to environmental conditions now changed and to geomorphic processes no longer effective in this region.

It has come to be realized during the last two decades that the Driftless Hill Land and adjoining portions of the Upper Mississippi Valley region display a variety of relict physiographic features which developed under the influence of periglacial climates accompanying the incursions of the Wisconsinan and earlier ice sheets. Straw (1966) recently proposed a periglacial origin for the fissures, "head" mantle, and solifluction aprons which occur along the Niagara escarpment in Grey county, Ontario. Extending the periglacial hypothesis to the fissures and to the associated minor relief and stratigraphic features of the Niagara escarpment in the Driftless Hill Land, as is done in the following pages, permits the origin of these phenomena to be encompassed within a theory of landscape modification already known to have applications within this region.

GEOGRAPHY

LOCATION

Niagaran rocks in northeastern Iowa first emerge from beneath overlapping younger formations near the common boundary of Winneshiek and Fayette counties. From there for over 125 miles into Stephenson county, northwestern Illinois the ragged edge of these and of the subjacent Alexandrian dolomites form an east- and north-facing escarpment along the western and southern margins of the Driftless Hill Land (Trewartha and Smith, 1941, pp. 43–44). Residual "mounds" capped by Silurian dolomites persist in southwestern Wisconsin in Grant (Sinsinawa mound), Lafayette (Platte mounds), and Dane (Blue mounds) counties.

STRATIGRAPHY

Four units comprise the pre-Pleistocene stratigraphy of the Niagara Escarpment area. In ascending order, these are: (1) the Maquoketa formation (plastic and fissile shales, dolomite, and limestone) of Upper Ordovician (Cincinnatian) age, of which only the uppermost plastic member (Brainard, or Upper Maquoketa) is of present interest; (2) the Edgewood formation (thick-bedded dolomite with nodular chert, locally in Fayette county a crystalline limestone) of Lower Silurian (Alexandrian) age; (3) the Kankakee formation (thin-bedded dolomite with much bedded chert) of Lower Silurian (Alexandrian) age; and (4) the Hopkinton formation (heavy-bedded to massive, fossiliferous dolomite, rarely with nodular chert and with massive chert only in extreme southwestern Clayton county) of Middle Silurian (Niagaran) age. The Edgewood in Dubuque county (Brown and Whitlow, 1960, pp. 31–44) and both the Edgewood and the Kankakee in western Illinois (Willman and Reynolds, 1947, pp. 7–8) have been subdivided, as has the Hopkinton in Jackson county (Savage, 1905, pp. 609–619).

The Hopkinton is rarely seen in the face of the escarpment, although in wells and other artificial exposures it is often found to be present as a thin cap on the salients and as a thicker mass a few thousands of feet back from the scarp crest. The Maquoketa is exposed only in the floors and banks of gullies cutting up the face of the escarpment; its presence is everywhere betrayed by the less steeply-sloping lower portion of the (topographic) Niagara escarpment, across which numerous blocks of dolomite have been strewn by frost heaving and by solifluction. The tattered margins of the two Alexandrian formations are those which in most places comprise the "Niagara" escarpment in this region.

Pleistocene deposits in the Niagara Escarpment area will be described later when the ages of the periglacial features are discussed.

STRUCTURE

All Paleozoic rocks in the southwestern part of the Driftless Hill Land dip south and west from the Wisconsin dome toward the Illinois and Forest City basins at an average inclination of about 20 feet per mile (Trowbridge, 1934). A few local structures are present in the area of the Niagara escarpment, notably the Savanna-Sabula (Preston) anticline, which trends east and west through those cities. The Savanna-Sabula anticline has a maximum dip on the steeper (northern) flank of 40 feet per mile.

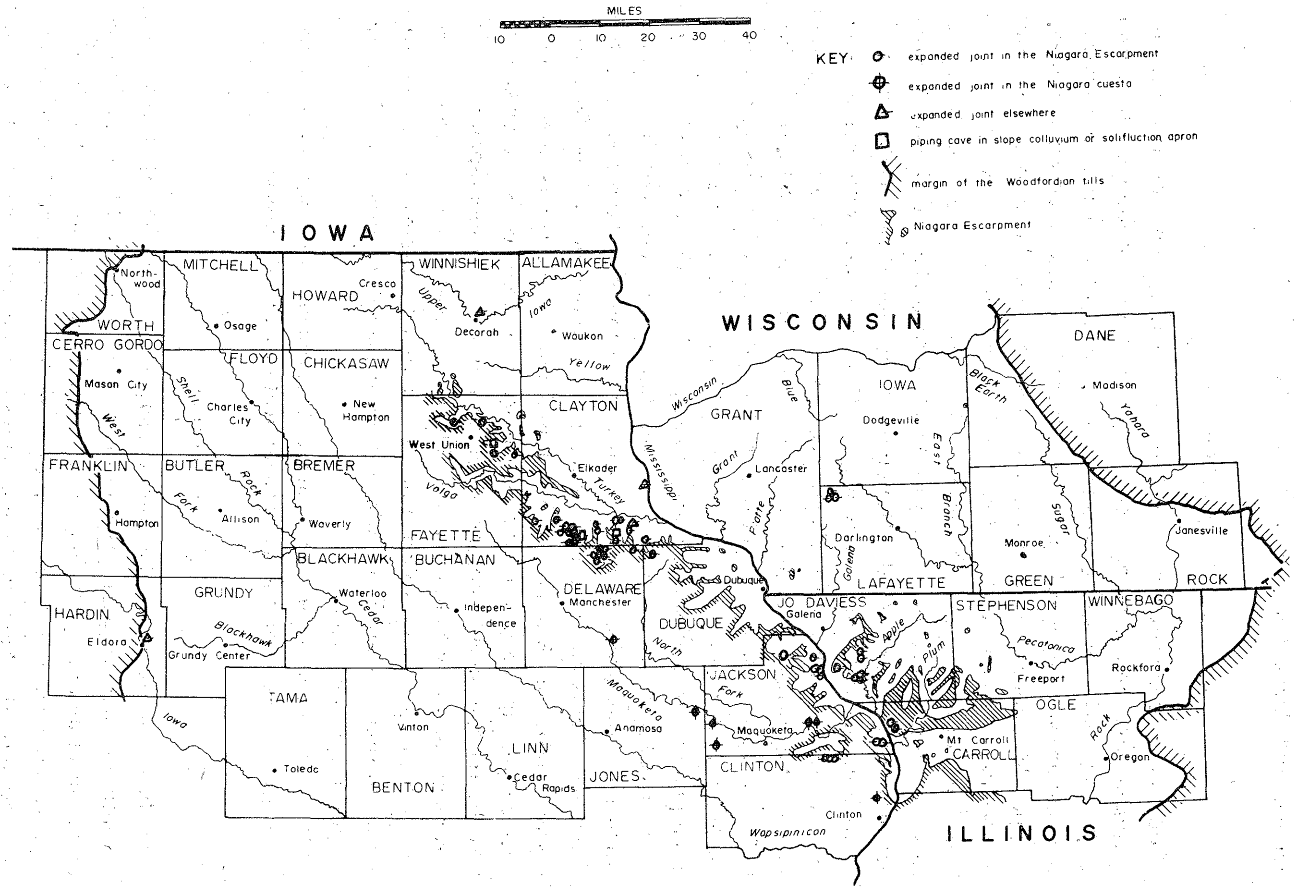


Fig. 1. The Niagara Escarpment Region of Iowa, Illinois and Wisconsin

Redrawn from maps of the U. S. Geological Survey and of the three state Geological Surveys

Two east-west anticlines comparable in size to the Savanna-Sabula anticline and many smaller structures were mapped in the Ordovician rocks of the zinc-lead district by Heyl, *et al.* (1959). These or other, similar, disturbances may have affected the Niagara Escarpment area. A few broad, shallow folds within the Niagara cuesta in Iowa were mapped by Howell (1935), using the top of the St. Peter sandstone as a datum plane. The scarcity of well records and the presence of settled blocks at many points along the face of the escarpment have prevented more-detailed structural investigations in this area.

TOPOGRAPHY

The Niagara escarpment forms the southern and western boundaries of the Driftless Hill Land (Trewartha and Smith, 1941, pp. 43–45). To the north and east lies a predominantly erosional topography, developed on indurated pre-Pleistocene rocks, the local relief of which often is measureable in hundreds of feet. To the south and west is a region of drift-mantled to drift-depositional topography where local relief rarely exceeds 200 feet.

West of Dubuque, the Niagara escarpment is a monolithic barrier. Its summit stands 600 feet above the alluvial floor or the Mississippi trench, being everywhere nearly 1200 feet AT. There are no gaps through it below 1000 feet AT. The rise of the cuesta is virtually imperceptible when viewed from the gently-rolling Kansan drift plain. As seen from the Driftless Hill Land, however, the scarp rises sharply two to three hundred feet above the Galena plateau.

The declining structural elevation of the cuesta south of Dubuque brings the crest of the escarpment down to about 900 feet AT in the vicinity of Sabula. To the east, in northwestern Illinois and southwestern Wisconsin, the Niagara cuesta is reduced to a multitude of outliers, some sinuous, others digitate, still others in the form of typical, conical mounds. Summit elevations east of the Mississippi range from about 900 feet AT at Savanna upwards to 1241 feet AT on Charles Mound, Illinois and culminate at 1719 feet AT on west Blue Mound, Wisconsin, depending upon the structural elevations of the individual remnants. The main body of Silurian rocks in northern Illinois is concealed beneath younger formations in the Illinois basin.

The immediate area of the Niagara escarpment includes four distinct topographic elements: (1) the gently-rolling, drift-mantled Niagara plateau, (2) the Niagara escarpment, (3) the Maquoketa apron, and (4) the dissected Galena Plateau.

The gently-rolling summit of the Niagara cuesta generally has been considered to be a remnant of the summit (Dodgeville) peneplain of the Driftless Hill Land (*cf.* Bates, 1939, Fig. 4). However, outcrop and well-log data include very few bedrock elevations above 1100 feet AT throughout the Iowa portion of the cuesta. The only important exception is the large salient north of Holy Cross, which is set off sharply from the remainder of the cuesta by an abrupt rise to 1200 feet. The uniformly high elevation of the land west and north of this salient is owed largely to drift deposits[1].

The most common aspect of the Niagara escarpment is that of a densely-wooded bluff inclined between 15° and 20° above the horizontal, near the top of which are discontinuous low outcroppings of disturbed, frost-riven dolomite. Cliffs more than 15 or 20 feet in height usually are found only where the retreat of cavern entrances has led to the development of deep re-entrant gorges within the escarpment, as at Dutton's Cave near West Union (Hedges, 1967).

At the base of the escarpment lies a rolling meadow which often is littered with large blocks of "float" derived from the scarp face above. The meadow, which is underlain by an apron of Maquoketa shale, has a pronounced slope north and east toward the Galena plateau. Previously it, too, was forested (McGee, 1891, Pl. 22), but by now the White settlers and their descendants have cleared much of it. In the Dubuque South quadrangle, areas underlain by mapped Alexandrian rocks (the escarpment) form slopes of 15° to 20° while areas underlain by mapped Maquoketan rocks (the meadow) form slopes of 8° to 12°. The break in slope between escarpment and meadow is conspicuous.

The transition from the Maquoketa apron to the Galena plateau is not especially striking. Level upland areas below the (topographic) Niagara escarpment and above the rims of the inner gorges frequently are underlain not by the Galena, directly, but by a few feet to a few tens of feet of bevelled Maquoketa shale, as is well shown by Brown and Whitlow (1960, Pl. 1). It is not until after one passes north of Turkey river that the plateau is developed predominantly on the Galena. For this reason (and others), Trowbridge (1914, p. 208) and some, but not all, other writers have referred to this plateau as the subsummit (Lancaster) peneplain.

[1] Consequently, it would appear better to extend the exhumed pre-Desmoinesian surface of Hedges and Darland (1963, Fig. 2) to the crest of the escarpment and to place the intersection of the projected pre-Desmoinesian and Dodgeville surfaces slightly north of the present scarp crest, except on the salient north of Holy Cross. A recent (unpublished) study of the subdrift topography of Maquoketa River valley shows that our earlier conclusion was incorrect.

PERIGLACIAL CLIMATE

Areal extent of the periglacial climate

A map of North America showing the "Hypothetical Life Zones of the Wisconsin Ice Age" was compiled by Dillon (1956, Fig. 11). Dillon's "tundra zone" encompasses all of the Driftless Hill Land, the adjacent ice--free portions of Wisconsin, Minnesota, and northern Illinois, and all but the extreme southwestern corner of the ice-free portion of Iowa. Brunnschweiler (1962, 1964), also, postulates the former existence of a tundra climate throughout this region. The tundra zone, as mapped by Dillon, in the United States is essentially coterminous with the zone in which periglacial features are found (Smith, 1962, p. 336).

Tundra conditions may not have existed far south of the Niagara escarpment, however, despite the extension of the glaciers hundreds of miles beyond. Frye and Willman (1958, p. 524) note: "Based on the foregoing data and the observations of Sharp (1942), permafrost existed for a relatively short time, and perhaps locally, adjacent to the margins of Wisconsin glaciers in northern Illinois, but became ineffective in central Illinois. It is judged that an arctic climate was not prevalent at the Wisconsin glacial border in its southernmost extent". They (p. 520) state that Frye and Leonard found relatively mild climatic conditions to have occurred in close proximity to the glacier front during the maximum extent of both the Nebraskan and Kansan glaciations in their type region, also.

The forest beds

At many places in Iowa, "forest beds" consisting of wood fragments and occasionally other forest debris occur immediately beneath drift sheets. The forest beds of the Des Moines lobe were studied recently by Ruhe and others. Ruhe and Scholtes (1959, p. 589) state: "It seems unreasonable to demand that the glacier ice, as it overrode the landscape, always incorporated old wood in its basal till. It seems more appropos to conclude that the Iowan ice overode a forested landscape...". Species identified include hemlock, larch, yew, and spruce (p. 592). In the Scranton No. 1 exposure, tree stumps still are rooted in place. The trees were broken off toward the south. Black (1965, p. 218) mentions "...that the Rockian ice advanced over a spruce forest...". There also is a forest bed below the Cary drift (Ruhe, Rubin, and Scholtes, 1957, p. 681).

From this, and from evidence in Ohio that living forests there were overwhelmed by the advancing ice (Brunnschweiler, 1962, p. 16), one might be tempted to conclude that climatic conditions adjacent to the ice at glacial maxima were boreal rather than tundra, even in relatively northern areas of central United States. Thornbury was convinced of this (1954, pp. 414–415): "Undoubtedly many areas peripheral to ice sheets had temperatures many degrees lower than those that exist today in the same areas. However, remarkably few features which can be associated with former permanently frozen ground or intensified frost riving have been described in such states as Ohio, Indiana, and Illinois. The evidence seems to suggest rather that climatic conditions in those states were not much more severe during glacial times than at present. Such pollen analyses as have been made of bogs in this part of the United States indicate that there was no tundra belt peripheral to the ice sheets as seems to have been true in central Europe. The evidence suggests that conifers grew right up to the ice sheets. Probably the presence of periglacial features in Wisconsin, Pennsylvania, and Missouri, if they be such, may be accounted for by higher altitudes of the areas in which the periglacial features exist". Black, in a review of recent periglacial studies in the United States (1964b, p. 21) tends to agree that paleobotanic evidence does not support the concept of an extensive tundra zone bordering the ice sheets: "A series of pollen studies indicate that last Wisconsin glaciations did not affect the vegetation much beyond their termini in northcentral United States...".

Geomorphic indicators of former periglacial climate

That periglacial landforms and soil structures are less extensively developed in North America than they are in Europe is freely acknowledged. Smith (1962, p. 336) attributes this difference in development to the difference in latitude between ice-marginal Europe and ice-marginal North America. Because glaciers extended much further south in North America than they did in Europe, "it is not surprising that the zone of more severe periglacial conditions should have been more restricted on the former continent than on the latter, with tighter squeezing of displaced climatic zones". There are no periglacial features in central North America south of the southern border of Wisconsinan and Illinoian drifts in Illinois, and there are none south of the southern border of Kansan and Nebraskan drifts in Missouri with the possible exception of hill slopes in the St. François mountains.

For more northerly areas of the continent, however, an abundance of geomorphic evidence is at hand showing that a truly arctic, tundra climate has existed in the past. This is particularly true at high altitudes and, also, in the Driftless Hill Land where the presence of Wisconsinan ice of the Des Moines and Lake Michigan lobes on three sides of the region in post-Rockian time probably led to unusually severe climatic conditions there during the interval from the Rockian advance to the Cary retreat.

The initial report of relict periglacial features in north-central United States *qua* periglacial features was made by Sharp (1942), who described soil involutions in northeastern Illinois. Since that time, many other geomorphic and pedologic phenomena whose origin is owed to the former activity of periglacial agencies have been described and several localities mentioned in older literature have been re-interpreted as exhibiting periglacial effects.

Fossil ice wedges have been described from southwestern Wisconsin (Black, 1957, 1961, 1964a, 1965; Black and Wittry, 1959), from east central Iowa (Wilson, 1958; Ruhe and Scholtes, 1959, p. 591; Ruhe, *et al.*, 1965, p. 18), and from northern Illinois (Horberg, 1949; Frye and Willman, 1958). Involutions have been reported to occur in Kansan drift in southern Iowa (G. M. Schafer, 1953; Lees, 1926, p. 335 *cf.* Sharp, 1942, p. 126) and it has been suggested that the planation of the Kansan drift surface in southern Iowa and northern Missouri was accomplished by solifluction (Schrader and Hussey, 1953). Stone stripes and rock monuments in Wisconsin are mentioned by Black (1964a, pp. 21 and 28; 1964b, p. 21). Smith (1949, p. 211) has suggested that the colluvial/alluvial terraces of the Kickapoo (*cf.* Bates, 1939, pp. 870–876) were developed under periglacial conditions. Deposits in southwestern Wisconsin originally described by Sardeson (1897) as the moraines of a "local glaciation" were re-interpreted by Smith (1949, p. 197) as solifluction flows.

As suggested by Chamberlin (1897), it is likely that the "local glacial deposits" of Squire (1897, 1898, 1899, 1908) in western and southwestern Wisconsin also were produced by solifluction, or by another extra-glacial process associated with glaciation. It also seems likely that some of the northeastern Iowa features discussed by McGee (1891) might be re-interpreted as periglacial in origin. For instance, his figures nos. 53, 100, 101, 102, 103, and 114 include layers resembling "head". He states (p. 513): "The common situation of such accumulations is in deep, steep-sided and tortuous valleys, transverse or oblique to the direction of ice flow and about the drift margin; and there are few such valleys so placed that are not at least partially lined with deposits of this character". McGee's figures nos. 76 and 119 appear to portray fossil ice wedges. He remarked (p. 556). "It is noteworthy that such witness to the energy of ice-work increase in number and in the eloquence of their testimony nothward..."

The ubiquitous loess deposits of central North America are not usually considered in discussions of periglacial geomorphology (Brunnschweiler's papers are about the only exceptions). At least since the work of Shimek (*cf.* 1895), however, they have been recognized as being aeolian in origin and as having developed under former periglacial conditions.

It is true that some "ice-wedge casts" are open to re-interpretation as soil tongues (Yehle, 1954) and that some soil involutions are known to be relatively recent in age (Wright, 1961, p. 943). The witness of the shattered, entombed forests cannot be denied, nor can that of the bog pollen. Yet, the preponderance of the evidence permits of no other explanation than the periglacial. I am confident that when more and better absolute dates are available for Pleistocene events in this region, and that when the meteorological values and the temporal spans of the several Pleistocene climatic phases have been firmly established, the apparent contradiction between the geomorphic and the paleobotanic testimony will melt away.

EXPANDED JOINTS IN THE NIAGARA ESCARPMENT

PHYSIOGRAPHIC SETTING

The principal subjects of this paper are the host of deep, narrow fissures which underlie small sinkholes dotting the crest of the Niagara escarpment. These fissures, literally "expanded joints", occur only along the margins of bluffs which form the local face of the escarpment. They are found only where valley-cutting has exposed Maquoketa shale at the bases of the bluffs and are usually located within two hundred feet of the crest of the bluff, often beneath the bluff face itself.

INTERNAL APPEARANCE

From within, the fissures are seen to be of uniform width. They are so in vertical as well as in horizontal section. Their floors almost invariably consist of fallen rocks, forest litter, and/or deposits of material referred to by splunkers as "heteropolycrapite" – empty beer cans, old fence wire, tread-bare automobile tires, the broken skeletons of unwary animals, etc. The explorer also notices, usually with alarm, that the ceilings of the fissures also consist of loose stones. Solid rock is visible only in the walls of the caves.

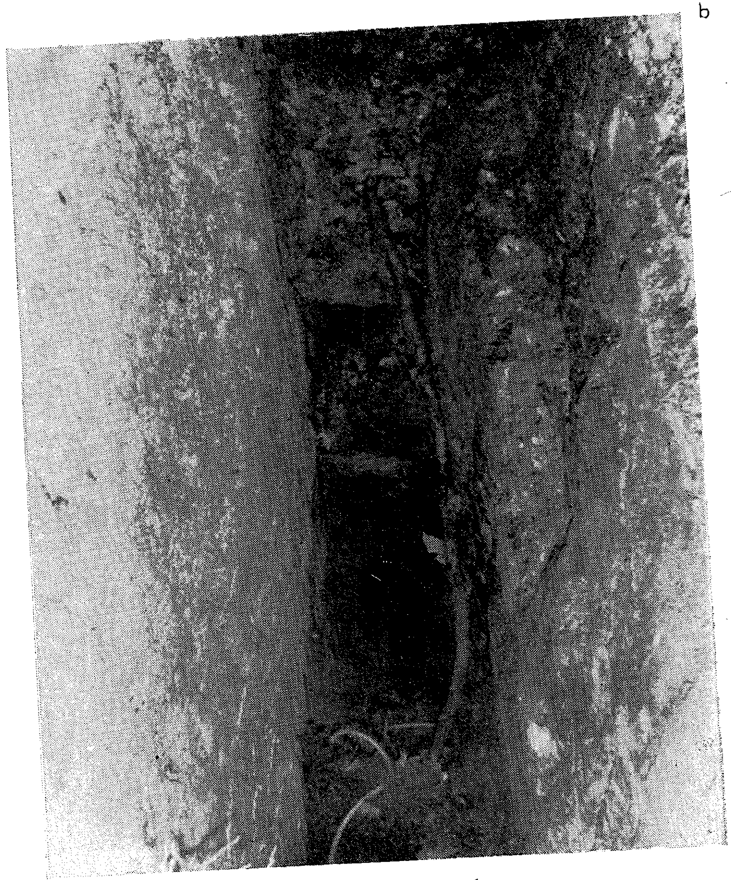
The walls upon close inspection are quite often found to be mirror images of each other. The lumps and hollows on the one side match exactly the hollows and lumps on the other. This mirror imagery not infrequently includes details as small as the matching broken ends of chert nodules and other minor rock structures. There is no suggestion of bilateral symmetry, of the complementary receding and protruding in opposite passage walls of beds varying in solubility, nor is there any trace of bedding-plane enlargement, as are customarily found in caves developed by solution.

There has been no perceptible vertical displacement of the strata in opposite

Pl. 1. Entrance to Mossy Cave, looking down

a

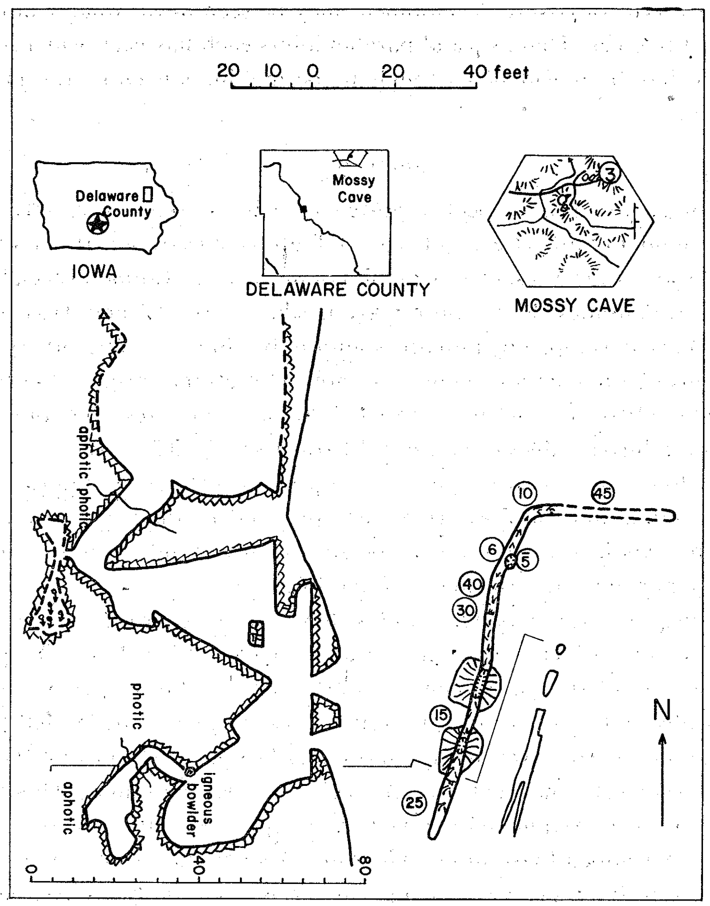
b



Pl. 2. Views of Mossy Cave: (a) north passage, looking inward, (b) north passage, looking down

passage walls. Fig. 2 is a map of a typical expanded joint. Pls. 1, 2, and 3 contain typical scenes from that cave.

While the fissures normally are developed along a single joint, Route Three Cave, northern Delaware county, has developed along at least three. In the face of the cut on Hwy. 3 west of Elk creek is an entrance about 18 inches wide and 10 feet high. A few tens of feet within, the passage doubles in height. Toward the rear, the upper and lower levels are superposed, but



C. R. G. Grade 6 survey prepared by James Hedges, David Jagnow, and Howard Wellemeyer; 26 August 1967

Fig 2. Mossy Cave, Delaware County, Iowa

NE/NE/NW 11 90 N 4 W 960′ A. T.; N.S.S. Standard Map Symbols (1961) toward the front, the upper level is angled to the west at 15°. The upper level ends in a terminal breakdown near the face of the cut. Fig. 3 is a map of this portion of the cave. Pl. 4 is a photograph of the entrance.

A few tens of feet down the hill from the cave entrance is a third opening, this one filled with rubble from road construction, which probably represents a lower level of cavern development. Only one fissure is present in the bluff at any given elevation. Wedging evidently occurred along several different joints at different elevations.

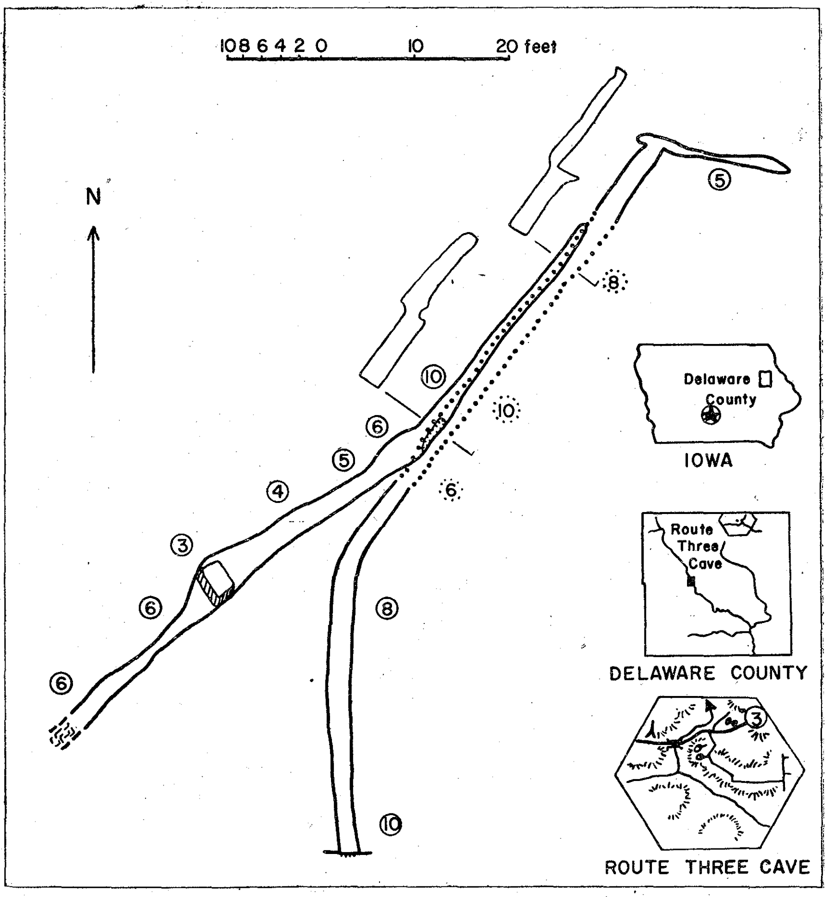
A third type of fissure development may be seen in the quarry along Hwy. 3 east of Elk creek. Here, several parallel joints each has been widened slightly but none has been widened enough to permit human entry for more than a few feet.

DISTRIBUTION

Expanded joints have been reported to exist in Wisconsin only in north Platte Mound, in Lafayette county east of Platteville (Potter, 1961). In Illinois, expanded joints are present in Carroll county near Savanna, especially in the Mississippi Palisades. Bob Upton Cave is said by Bretz and Harris (1961, p. 84) to have developed by phreatic solution, but Bat Cave and others reported to exist in the same area evidently are not of solutional origin. Trowbridge and Shaw (1916, p. 114) mention small "elongate depressions joined end to end" on a Silurian ridge in the central part of sec. 22 T27N R1E in JoDaviess county. Other sinks in JoDaviess county are located on the Silurian uplands about Blanding and on a Silurian ridge one and one-half miles southeast of Woodbine. These probably are associated with expanded joints. I have not had an opportunity to visit any of these localities.

In Iowa, several expanded joints have been recorded in northeastern Jackson county (Spelunking, many short articles 1957–61). The Dubuque county section of the escarpment is poorly known from a speleological viewpoint, probably because of the proximity of the large, well-decorated Dubuque mines and caves. However, many sinks along Middle Fork of Little Maquoketa river are shown at the center of the Holy Cross quadrangle. This area was discussed by Grant and Burchard (1907, p. 8), who thought that the sinkholes had developed by solution. The sinkholes here are mostly on the plateau rather than in the bluff, which is uncharacteristic of sinkholes associated with expanded joints.

Expanded joints are best developed in that portion of the Niagara escarpment which lies along the common boundary of Delaware and Clayton counties, especially along Elk creek. The deepest known fissure, Big Breathing Hole (est. 125 feet deep below the entrance, Hedges, 1963a), occurs here and the longest know fissure system, Yew Ridge Cave (estimated to extend about



C. R. G. Grade 4 survey prepared by George Darland, Ed Dvorak, James Hedges, and Dan ver Ploeg; 28 March, 20 June, 25 July 1959

Fig. 3. Route Three Cave, Delaware County, Iowa

NE/NE/NW 10 90 N 4 W 920′ A. T.; N.S.S. Standard Map Symbols (1961)

700 feet beyond the entrance, Barnett, 1967a; 1967b, p. 26), is in Dubuque county about a mile east of the Delaware county line.

Only a few expanded joints have been reported to occur along the Niagara escarpment northwest of the Elk Creek area. This no doubt is an artifact of areally haphazard exploration. It seems probable that fissures are rather evenly distributed along the lenght of the escarpment.

Expanded joints have also been reported to occur in southeastern Jackson and in northeastern Clinton counties near Sabula (Hedges, 1961) and Preston (Barnett, 1967c). These caves lie near the outcrop of Maquoketa shale along the crest of the Savanna-Sabula anticline.

Surface exposures of Maquoketa shale in the backslope of the Niagara cuesta have been reported in Dubuque county at Washington Mills (Norton, 1911, p. 379) and in Delaware county at Rockville (Calvin, 1894, p. 41). Neither of these areas have been examined for expanded joints.

The internal morphology of Liivaama's Cave, adjacent to Maquoketa river south of Springbrook, closely resembles that of the expanded joints along Elk creek. Published reports and unpublished well records of the Iowa Geological Survey for the counties of Clinton, Jackson, Jones and Delaware reveal the presence of several large subdrift valley systems incised into the backslope of the Niagara cuesta. Maquoketa shale is exposed beneath the drift along many of these buried valleys. Unfortunately, well records are too few in the area of Liivaama's Cave to permit one to know whether or not there is a buried valley adjacent to the cave deep enough to have reached the top of the Maquoketa shale.

The locations of expanded joints known or reported to exist in the Niagara escarpment are shown in Fig. 1.

GENETIC HYPOTHESES

Solution

Most expanded joints in the Niagara escarpment contain very little evidence of solution. This is as true of those at Griersville Rock, Ontario (Straw, 1966, p. 372) as it is of those in the Driftless Hill Land. The only important exception to this rule is Mossy Cave (Fig. 2, Pls. 1, 2, 3) where, at the lowest point in the cave, a prominent solutional half-cylinder and many scallops are cut into the east wall. The general absence of travertine deposits in the fissures is also an indication, albeit not so convincing, a one, that solution has not played an important rôle in their development. The half-cylinder and scallops at Mossy Cave are probably due to local, exceptional processes which may have operated at any time before, or after, the expansion of the joint opening, although it is of course possible that solutional features once were common in the fissures and that they were destroyed by frost spalling at the time when the cave breccias were formed.

I have previously described the occurrence of a vertical shaft or "domepit" at Dutton's Cave, Fayette county (Hedges, 1967, Fig. 4b). This vertical shaft is similar to the vertical shafts in Mammoth and other caves of the Central Kentucky Karst region and developed by the solution of Edgewood limestone beneath a caprock of cherty Kankakee limestone. That the expanded joints, also, are high and narrow, are developed in the Kankakee and Edgewood formations, and are located near bluff-lines suggest that they, too, may be vertical shafts.

The expanded joints, however, lack the vertical grooves or lapies which are characteristic of vertical shafts. At Dutton's Cave, the vertical shaft is developed largely beneath the insoluble cherts of the Kankakee, while individual expanded joints are as well developed in the Kankakee as they are in the Edgewood. Vertical shafts possess bilateral symmetry, whereas the walls of fissure caves form mirror images of each other. In vertical shafts, moreover, the ceilings and floors of the openings usually consist of solid rock, which is definitely not true in expanded joints.

Alan Howard proposed a few years hence that these fissures are a solutional form intermediate between cutters (Subkutan, or Bodenbedeckter, Karst) and true caves (Subterran Karst) (Howard, 1963). For a time, I was inclined to agree with him (1967, p. 84) but, as noted above, the majority of fissures contain few if any features of solutional origin. They are as wide at the base as they are at the top, which is not true of cutters and is rarely true of solutional caves. Howard's concept includes subterranean drainage piracy by cave passages leading from beneath a caprock-covered area (the Niagara plateau, underlain by cherty Kankakee dolomite) to the heads of valleys which have been incised through the caprock. The expanded joints, however, lie parallel with the walls of adjacent valleys. They do not intersect valley walls except where two valleys join (where they unavoidably must intersect valley walls). Howard's illustration of an idealized cave shows a cave passage developed in the Edgewood formation, above which the Kankakee is intact excepting a few sinkholes. While it is true that the expanded joints are roofed across at most points, their roofs ordinarily consist of loose mantle rock, not of undisturbed bedrock. Also, individual expanded joints are not confined to the Edgewood but are developed in the Kankakee as well.

Gravity

Solution as the primary cause of the fissures appears unlikely. The initial, antecedent joint openings seem definitely to have been expanded, physically widened, rather than to have been enlarged by solution. Wind, waves, and vulcanism, the agencies commonly held responsible for the development of non-solutional caves, rather obviously were not involved. Caves are sometimes formed by gravity, however, as during the accumulation of talus. Gravity also produces caves as a result of the slumping of joint-bounded blocks adjacent to cliffs.

The fissures are not talus caves. The rock on both sides of the passages is still essentially whole and undisturbed. No falling, landsliding, rotation, fragmentation, distintegration, or accumulation has taken place. These are not random cracks and crannies among superficial bowlders; they are substantial, regular voids within the depths of the earth.

Slumping movements, in which a slab of rock adjacent to a cliff moves *down in toto* and *out* at its base to form a toreva block, are not evident in the fissures of the Niagara escarpment. Toreva-block caves possess triangular cross-sections widest at the base and strata in opposite walls of the opening are vertically displaced relative to each other, whereas the fissures in the Niagara escarpment are of uniform width and the beds in opposite walls are equal in elevation.

The outward and downward slipping of glide blocks detatched from scarp faces across inclined shale surfaces lubricated with water has occurred in the Chuska mountains of northwestern New Mexico (Watson and Wright, 1963; Wright, 1964, 1965; Michael, 1965) and in the Colorado plateau (Flint, 1957, p. 196). Where a thick, blocky regolith exists, as along the Niagara escarpment, crevices between the youngest glide blocks might for a time remain roofed over as a result of soil creep. The top of the Maquoketa shale is well lubricated (it causes a prominent spring line at the base of the escarpment). However, the contact plane between the Maquoketan and the overlying Alexandrian rocks dips slightly into the face of the escarpment, not outward toward the adjacent vale. Block-gliding does not appear to have been the cause of the expanded joints in the Niagara escarpment.

The presence of slumped blocks along the escarpment has been reported by several authors. Willman and Reynolds (1947, p. 7) mention that in northwestern Illinois, "the position of Silurian outcrops at the top of steep slopes and overlying the soft Maquoketa shale, commonly results in slumping of large Silurian blocks. Slumped blocks can usually be recognized by tilting of the strata." Settled blocks of Niagara dolomite in the area of the Lancaster and Mineral Point quadrangles were noted by Grant and Burchard (1907, p.8). Calvin and Bain (1899, pp. 446–448) state that the Illinois Central Railway line near Peosta "at one point cuts through Maquoketa shales at a depth of 70 feet below the contact with the Niagara; and only 30 rods to the southwest, without essential change of level, the track enters a cut in the lower quarry beds of the overlying limestone. A mass of Niagara of unknown length, and full 100 feet in thickness, has here been allowed to settle, on account of flow of the soft shales, through a vertical distance of not less than 80 feet".

I do not mean to deny that plastic deformation of the Maquoketa with consequent slumping of the overlying Niagara, has occurred. It has. We are dealing here with two separate phenomena: with (1) slumped blocks which may easily be recognized by their vertical displacement and/or tilting and (2) with other blocks which have been forced outward without either vertical displacement or tilting. Probably, slumping is the normal process in the destruction of the escarpment. If so, the expanded joints are transient, ephemeral irregularities brought about by unusual climatic conditions associated with former glaciers. Smith (1953; 1962, p. 329) points out that in the "rock cities" near Olean and Salamanca, New York, displaced blocks nearest the escarpment are unaffected by gravity but that those farther out from it "display varying degrees of tilting and downward movement, grading into landslide masses".

With respect to this distinction, Route Three Cave is crucial. Here, several different joint planes at different elevations within the escarpment have been widened, yet slumping could not have occurred because each level of the cave is bounded at top and (possibly excepting the lowest level) bottom by undisturbed dolomite. Route Three Cave might, however, have arisen from tensions produced by cambering of the dolomite outward into the adjacent valley of Elk creek. If no joints extending throughout the dolomite section were present at this locality, several discontinuous joints might have been opened each at a different elevation.

Caves within the Dachstein massif of Austria are thought to have resulted from gravity movements although they, like Route Three Cave, have floors and ceilings of undisturbed bedrock. Groom and Coleman (1958, p. 16) state that Arnberger "suggests that such caves originated tectonically during the movements which formed the Dachstein nappe, and again during later orogenic stages. At such times the limestone beds were subjected to local tensions in a direction generally coincident with the dip. The bedding planes then became planes of slipping movement down which detatched packets of rock slid. This produced cave spaces between the moved blocks and the rock which resisted tension and remained in place".

If the typical expanded joints in the Niagara escarpment were due to plastic deformation and slumping, Route Three Cave might have arisen from tensions produced by cambering of the dolomite caprock outward toward the valley of Elk creek. The accompanying photograph (Pl. 4) shows, however, that the beds remain horizontal on both sides of the entrance to Route Three Cave. Cambering has not occurred here.

Although the expanded joints of the Niagara escarpment in the Driftless Hill Land do not appear to be explainable by slumping or cambering, Straw has attributed the expanded joints at Griersville Rock, Ontario to gravity movements. He (1966, p. 371) found pulverized, crumpled Cabot Head shale to be intruded upward into the lower portions of the openings. "Displacement of the dolomite blocks appears to be a result of mass-movement under gravitational influences, and separation and movement of the dolomite blocks was clearly facilitated by failure of the upper part of the Cabot Head shale and by the presence of master joints. Lateral flow of broken shale toward the scarp edge seems to have resulted from the presence of excessive quantities of groundwater" associated with a periglacial climatic regime.

Gravity movements are also supposed to have been responsible for the origin of the many deep, narrow openings adjacent to cliffs in central Montana. Campbell (1968) writes that these caves are developed in thick, massive limestones overlying shales, both of which have been exposed by canyon--cutting in the Little Belt and other dissected mountain ranges. Most of the openings occur within a few hundred feet of canyon rims, on the up-dip sides of the canyons, and are oriented parallel with the local trend of the bounding cliffs. Shale often is exposed in the floors of the fissures; they are roofed over by loose rocks and vegetation. Some of the fissures have been slightly modified by solution.

Campbell mentions (p. 27) that Hubbert and Ruby found that a body of rock lubricated and partially buoyed up by water is capable of travelling along a surface inclined at only one or two degrees. He concludes (p. 28): Smokeholes nos. 1 and 2, and other caves like these," were probably formed by just such a sliding process. Water seeping through the limestone would reach the impermeable shale and travel along the limestone-shale interface producing a lubricating effect which would allow blocks of limestone to break loose and slide down dip".

These fissure caves of Montana are remarkably similar to the expanded joints of the Niagara escarpment in Ontario and in the Driftless Hill Land.

The gravity hypothesis is that which customarily is advanced by European writers in discussing expanded joints. Warwick (1962a, p. 26) and Straw (1966, p. 374) mention open fissures in Derbyshire and Yorkshire, England which occur near the edges of gritstone escarpments underlain by shales. Warwick (1962b, p. 147) further describes the Yorkshire examples as having "formed after the thawing out of the ground during the latter part of the Ice Age when the clays underlying the limestones became saturated with water, and were squeezed out by the weight of the limestones above".

In France, enlarged joints in limestone overlying shale in Provence and in the gorges of the Tarn are described by Gèze (1953, pp. 3–4): "Dans les régions où des grandes masses calcaires reposent sur des couches plastiques avec un certain pendage, sous l'influence de la gravité, des masses calcaires peuvent se disjoindre et s'éloigner légèrement les unes des autres, en glissant sur leur substratum. En très grand, se phénomène est appelé "tectonique en banquise". En plus petit, des blocs peuvent basculer sur les pentes au voisinage des gorges où de dépressions topographiques, tout en restant reliés à la masse principale".

Subglacial

Martin (1928) described fissures in bedrock which were formed by ice-wedging beneath glaciers (subglazialer Frostsprengung). These fissures are narrower than the expanded joints of the Niagara escarpment, they wedge out downwards, and they were filled with ground moraine concurrently with their development.

At Griersville Rock, Ontario, the salient which the fissures are developed faces the direction from which the ice advanced. In the Driftless Hill Land, fissures occur in bluffs lying in many different orientations to the direction of previous ice movements. None of the fissures are filled with glacial debris. They are overlain by a mantle of locally-derived cryoturbates, not by drift. They do not wedge out downwards. The subglacial hypothesis is not applicable to the expanded joints of the Niagara escarpment.

Ice Wedging

Some expansive process must have taken place within joint openings parallel with the local face of the escarpment as a result of which blocks of dolomite underlying the scarp face gradually were propelled outward across the top of the Maquoketa shale without undergoing either settling or tilting. The agency best qualified for the accomplishment of this task is ice wedging.

Contraction of the ground under intense cold, followed by the development of ice wedges in the resulting contraction fissures (in our case, within pre-existing joints slightly widened by contraction of the adjacent rock) and repeated annually over a number of years as described by Leffingwell (1915)[2], gradually would lead to the creation of long, narrow openings uniform in cross section the walls of which would possess mirror imagery. The force which caused displacement of the valley-ward blocks was not the pressure of crystallization of the ice. It was the pressure of the rock itself as it expanded

[2] An earlier report of this phenomenon was published by Thoreau (1966, p. 181). Thoreau states that during his sojourn at Walden Pond, Massachusetts about 1845, "..I was awakened by the cracking of the ground by the frost, as if someone had driven a team against my door, and in the morning would find a crack in the earth a quarter of a mile long and a third of an inch wide".

upon warming for, the rock being yet frozen although many degrees warmer, ice remained in the contraction fissures and the pressure developed by expansion could be relieved only by outward displacement of the blocks between the ice--filled fissures and the scarp face.

In Hunsberger Cave, north of Elgin and in many other caves, it may be seen that the joint opening which is the cave is uniform in width where parallel with the bluff face but where the joint turns sharply toward the face it immediately becomes almost closed. This is precisely what one would expect to find had the hillside been moved bodily into the adjacent valley. Frost-induced motions in the overlying mantle of "head" maintained a roof across the developping fissures, except locally where a fissure became too wide for the available subsoil fragments to bridge across it.

Leffingwell (1915, p. 652) estimated that an ice wedge three meters in diameter could form in northern Alaska in only 600 years. Even if the development of the expanded joints in the Niagara escarpment should have required 6000 years, this length of time is not excessively long in view of the number of glacial advances and accompanying phases of periglacial climate to which the escarpment has been subjected.

Black (1964a, p. 25) reports that, by analogy with ice wedges in northern Alaska, 1000 to 3000 years would sufficed to produce the ice wedges of Wisconsin. "An average rate of growth of one millimeter per year was found at Barrow, Alaska and is probably not out of line for Wisconsin". The maximum width of the fossil ice wedges of Wisconsin is about twice that of the expanded joints in the Niagara escarpment. If the Cary advance lasted 3500 years, and if out of that time only 1000 years saw periglacial conditions in southern Iowa, there would still have been adequate time for the development of the expanded joints during the Cary interval alone. Most of the expanded joints in the Niagara escarpment are less than one meter in width.

The actual amount of expansion need have been only a few inches in some instances. One sometimes finds tongues of rock projecting into accessible passages (Pl. 2 b), beyond which two closely-spaced fissures continue too narrow for human entry. Thermoclastic processes may have caused spalling of the cave walls and partial destruction of the intervening partition, with resultant creation of a single, relatively wide, accessible passageway.

In a few places, several parallel joints may be seen each to have been widened slightly instead of a single joint having been widened a great deal. This is well shown in and near the quarry on Hwy. 3 east of Elk creek.

Ice wedging previously has been suggested as the cause of joint expansion in other regions. Although Straw (1966) attributes the fissures in the Niagara escarpment at Griersville Rock, Ontario primarily to gravity movements, he states that in addition to cant and creep as the shale was extruded "the joints may have been forcibly widened by the growth of ice veins during former more severe climatic conditions" (p. 372). The "Rock Cities" near Olean and Salamanca, New York were attributed to ice wedging by Smith (1953, 1962). Smith (1962, p. 329) holds that the blocks "obviously were moved to their present position by a force acting only in a horizontal direction, and the effects of frost or ice wedges under former periglacial conditions are believed to provide the only adequate explanation".

Koteff (1967) reports that vertical joints in the upper part of a granodiorite knob near Little Quitacas Pond, Massachusetts were enlarged by ice wedging. Ice in horizontal joints furnished a lubricated sliding plane over which the blocks moved.

The presence of a sliding plane, whether of shale or of ice or of some other material, is essential. Where one is wanting, the action of frost on indurated rocks leads variously to the development of bedrock polygons with shattered edges, as at Brönlund Fjord, northern Greenland (Davies, 1961), to frost valleys, as in central Labrador (Twidale, 1956, 1958), or to compaction and loss of pore space and the development of many tiny crevices, as in chalk (Cailleux, 1943).

EXPANDED JOINTS ELSEWHERE IN THE UPPER MISSISSIPPI VALLEY REGION

Deep, narrow fissures more or less like the expanded joints in the Niagara escarpment occur at widely scattered points in and adjacent to the Driftless Hill Land. Some of these are analogous in origin to the expanded joints in the Niagara escarpment and logically should be discussed in connexion with them.

FROST-RIVEN BLUFFS IN THE NIAGARA CUESTA

Deeply-riven bluffs are found in the Niagara cuesta at a few places where valley floors adjacent are not known to have been eroded deeply enough to expose the top of the Maquoketa shale.

In one of these bluffs, three miles south of Springbrook, is developed Liivaama's Cave. The interior of the cave resembles the interior of an expanded joint. However, attempts have been made to blast the cave shut and these may have altered its appearance. In the absence of proof that shale is exposed beneath the drift which partly fills the adjacent valley of Brush creek, Liivaama's Cave cannot be classified as a (genetic) expanded joint.

The right bank of Maquoketa river just upstream from Canton is rent by numerous fissures. In the deepest portion of these is the Canton Ice Cave. Typical expanded joint morphology is lacking in this cave. Maquoketa River valley is not known to be entrenched as deeply as the top of the shale at this point. On the other hand, there is no suggestion of solutional activity in the development of the cave. The entire bluff in the area of the cave is nothing but a chaotic mass of huge blocks. It seems likely that Canton Ice Cave owes its origin to periglacial weathering processes, but it is not a (genetic) expanded joint.

On the left bank of Raccoon creek, below Dancehall Cave in the Maquoketa Caves State Park, is a narrow defile billed the "Fat Man's Misery". Maquoketa shale does not crop out here. The slab of rock which stands between the trail and the creek bed probably was pried out from the adjacent cliff concurrently with the development of the West Union talus on the floor of the valley. The origin of the defile occurred during a time of periglacial climate, but "Fat Man's Misery" is not a (genetic) expanded joint.

FROST-RIVEN BLUFFS IN THE GALENA PLATEAU

Before the turn of the century, the most famous cave in this region was probably the Decorah Ice Cave (Kovařik, 1898). It is formed by a toreva block of Galena limestone along the left bank of Upper Iowa river at Decorah. White (1870, v. 1, p. 80) assumed that the base of the block had slipped out across the top of the underlying Decorah shale under the influence of gravity.

The bluff contains several other openings similar to the Ice Cave, although none is so large or so deep. Gravity, aided by undermining of the cliff by Upper Iowa river, may have been competent to enlarge these fractures to their present size, but periglacial ice wedging also would have been competent to do so. If these openings were produced by ice wedging, the thin blocks prized out may have been incapable of standing unsupported after melting of the ice wedges and, fortituously, they settled back against their parent ledge instead of toppling into the river. The ice which now seasonally occurs in the Ice Cave is, of course, not left over from glacial times. It melts completely each autumn and forms anew each spring.

South of Clayton, on the left bank of Buck creek at its mouth, is developed a cave resembling the expanded joints of the Niagara escarpment. Cricket Cave might be a gash vein. It is developed at the proper stratigraphic horizon in an area where gash veins occur. However, it is partly filled with breakdown, lacks the laminated silts and the travertine deposits characteristic of gash veins, and it is not mineralized. Decorah shale is intersected by the face of the bluff. It seems permissible to interpret Cricket Cave as an expanded joint developed by periglacial ice wedging.

Some caves beneath sinkholes in the Galena plateau are high, narrow joint openings which in gross outline resemble expanded joints. However, these enlarged joints typically occur in the interior of the plateau, miles from the nearest valley deep enough to expose Decorah shale, and they possess many morphologic details of solutional origin. These joints were enlarged by solution, not expanded by ice.

"Fissure Caves" dissolved out by artesian water ascending to outlets beneath ancient valleys were described by Brod (1964) from the east flank of the Ozark dome, Missouri. These caves somewhat resemble expanded joints. Most, however, are developed below the Decorah shale, in Plattin limestone. Only a few are found in the Kimmswick limestone above the shale. One (Ehler's Pit, p. 94) passes from Kimmswick through the Decorah into Plattin limestone. These "fissures", like the "Fissures" in the interior of the Galena plateau, are clearly solutional in origin.

FROST-RIVEN BLUFF NEAR ELDORA

In one of the mural cliffs of Eldora (Pennsylvanian: Desmoinesian) sandstone along Iowa river near Eldora is a deep, narrow opening locally known as Rattlesnake Cave. According to Raby (1957), it is developed on a single vertical plane parallel with the bluff. There are vertical and lateral branches at various levels. The depth of the cave is about 60 feet, or very nearly the total height of the bluff above the shale, unit no. 3, in the bluff at Eldora (Beyer, 1899, p. 272). Beyer describes the sandstone bluffs here as "weathered and shattered".

Rattlesnake Cave lies one mile east of the Bemis (Altamont of Beyer) moraine, at N42°20′. It seems very likely that ice wedging and frost riving occurred here. Rattlesnake Cave, like the expanded joints in the Niagara escarpment, occurs adjacent to and parallel with a valley deeply enough incised to have exposed an underlying shale.

DRIFT-FILLED JOINT IN THE BARABOO RANGE

Weidman, in discussing supposed glacial deposits west of the Cary moraine near Devil's lake, Wisconsin, mentions a "kettle" in the quartzite range southwest of North Freedom (1904, p. 103). Thwaites (1935, p. 403) re-examined the feature and concluded that it is a drift-filled fissure. Ice wedging might have been active here. An ice wedge might have developed in a steeply-inclined bedding plane and forced out a block across a layer of ice in an underlying joint plane, after the fashion of Koteff's (1961) examples in Massachusetts, although it is not possible to tell for certain on the basis of published information.

OTHER PERIGLACIAL PHENOMENA ALONG THE NIAGARA ESCARPMENT

Periglacial weathering effects along the Niagara escarpment are not limited to expanded joints. Four additional kinds of periglacial features have been described in the escarpment area. These phenomena, and two associated erosional forms, are discussed below.

BLOCK STREAMS

Smith (1949, pp. 208–210) described a block stream and other blocky accumulations on the flanks of the Blue mounds, Dane county, Wisconsin. The Blue mounds are outliers of the Niagara cuesta. These scattered blocks are similar to the "float" of dolomite blocks strewn over the Maquoketa apron in Iowa and Illinois. On the Blue mounds, blocks occur on slopes as low as 3 to 4°. They can be found nearly a mile from the summit and as much as 600 feet below it. Smith observes, "The individual blocks ... appear to be stranded where they lie and to be undergoing gradual breakdown .... Irregular cracking of the rock is common, and detatched fragments are seen on and around many of the blocks". Were the blocks a lag deposit, had they remained behind after removal of the less resistant underlying shale, those at the greatest distances from the summit should be more decayed than are those nearer the summit. This is not the case. Smith (p. 210) inclined toward the view that the blocks were transported by solifluction, although he allowed that "the largest of the blocks, particularly on the east slope, represent unit movement" (seasonal frost heave) "rather than mass movement".

According to Smith (p. 206), Büdel found that blocks in a part of Europe had moved as much as two kilometers and on slopes as low as 2°. "Postglacial stability was found to be characteristic of block accumulations having slopes up to more than 17°". Black (1964b, p. 12) states that Hamilton found large dolomite blocks in an area adjacent to the late Wisconsin moraine to be stationary at present, although "hillside rubble moved three to fifteen millimeters during freezing and thawing in one winter" and "Some slumpage seems to occur in the steps or terracettes". These findings are in agreement with the observed distribution of float blocks on the Maquoketa apron and with the overgrown, decaying appearance of the modern Niagara escarpment.

Float blocks in Iowa are sometimes found along the margin of the Galena plateau as much as a mile from the Niagara escarpment. Calvin and Bain (1899, p. 447) mention having found one near Twin Springs which had thus crossed the entire Maquoketa apron and plunged 60 feet into the Galena limestone gorge of Little Maquoketa river.

SOLIFLUCTION APRONS

At several places in Clayton and Fayette counties blocky, terrace-like benches are found at the base of the Niagara escarpment. These solifluction aprons consist of a jumble of angular blocks embedded in angular dolomite gravel and silt. The benches do not have level surfaces, although they are roughly of uniform height. Many blocks protrude above them. One such bench, its side partly eroded by the stream, occurs at the base of a north-facing slope below McCreary's Cave, northwest of West Union (Hedges, 1967, pp. 86–87). Another is that in which is developed Sampson Ice Cave, northwest of Colesburg (Hedges, 1963b). The latter bench lies at the base of a south-facing slope along Pine creek. Sampson Ice Cave consists of a passage 30 feet long developed by the piping out of fines from among the blocks.

SLOPE COLLUVIUM

J. P. Schafer (1962) recently re-emphasized the fact that "Hillsides in northeasternmost Iowa ... are almost entirely covered by Pleistocene slope colluvium ...". This mantle of "head", or congelifractate, is gradational with the underlying, parent materials and with the overlying solifluction aprons. "Head" deposits also are common in Wisconsin. Black (1964a, p. 22) states: "Mass movement of soil is evident practically everywhere in the state. In the southwest especially where little drift was deposited, mass movements have been an effective process in removing the soil from bedrock hills and filling the valleys with thick accumulations of debris; practically all exposures illustrate such movements..."

Hamilton found limited movement of hillside rubble to occur during Wisconsin winters at the present time (Black, 1964b, p. 12). Downslope motion of the colluvium was greatly accelerated during former times of intense frost action. Most of the material no doubt was produced under former periglacial regîmes rather than under modern climatic conditions.

Peltier's (1949, p. 48) description of the rubble slopes along Susquehanna river could be applied almost without change to slopes in the Driftless Hill Land. There, as here, the slopes "are covered by an evenly distributed deposit of rubble which extends into the valleys and beneath the sediments of the small creeks. ... In drainageways on mountain slopes the finer constituents are removed from the rubble surface and a blockfield remains".

The face of the Niagara escarpment frequently displays no naked exposures of dolomite *in situ*, except in the floors of gullies. Its most common aspect is that of a densely-wooded bluff lying between 15 and 20 degrees above the horizontal, near the top of which may be discontinuous dolomite outcroppings a few feet in height. Most of these "outcrops" upon close examination are found merely to be large chunks of disturbed rock.

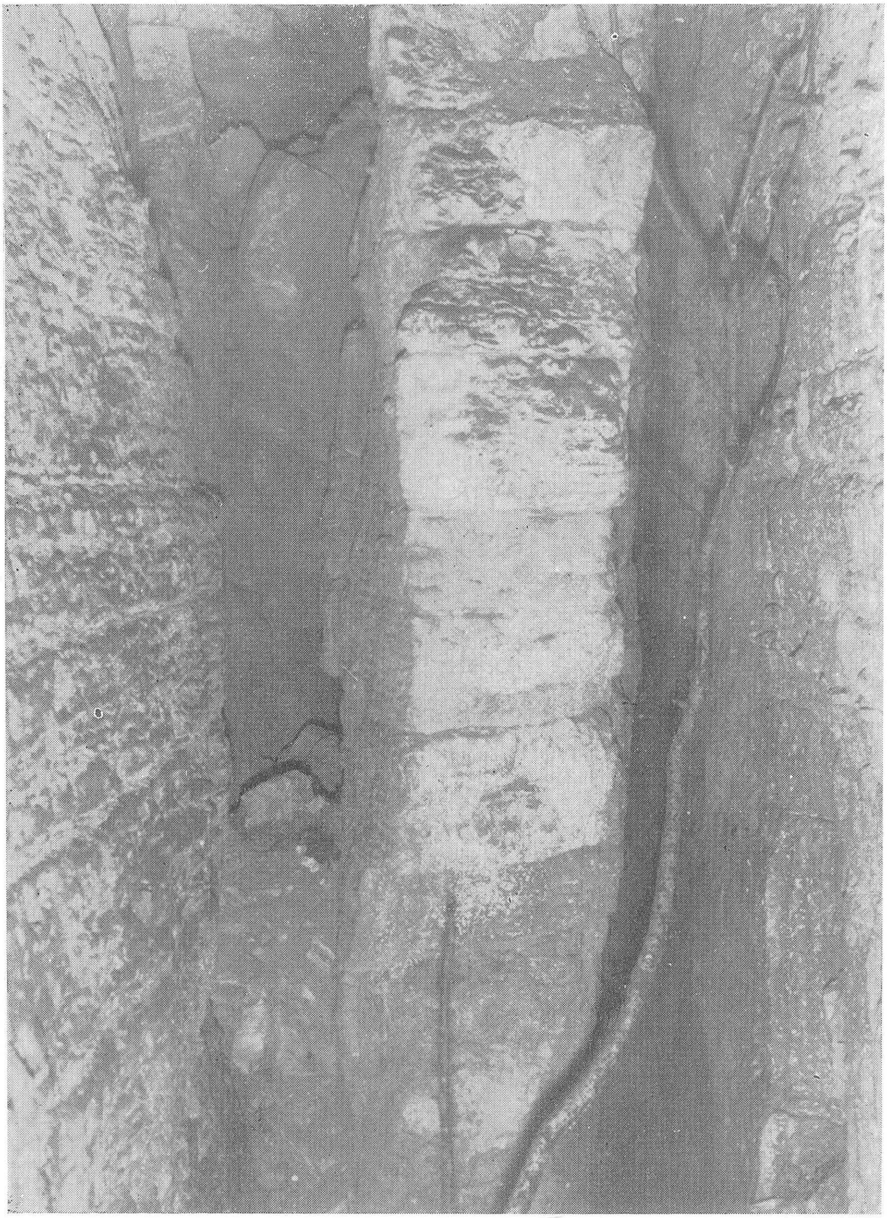
As seen in road cuts, the scarp face consists of a steeply-bevelled pediment of dolomite clothed in a mantle of frost-riven stone, talus, and soil 10 to 15 feet in thickness. The general absence of unencumbered exposures suggests that the few which are present, excepting those in stream beds and, possibly, those along the gorges of large, high-gradient, through-flowing streams such as the Volga below Fayette, are relicts.

Two particularly noteworthy occurrences of slope colluvium are those in which are developed the Bixby and the Brainard ice caves. At Bixby State Park, the south side of Paradise valley is littered with large blocks partly buried in gravel and silt. An opening among the blocks 25 feet in length constitutes the Bixby Ice Cave. Late in winter, after most of the ice has sublimated, it is possible to crawl a few feet further. Near the former site of the hamlet of Brainard, the bluff south of Otter creek is littered with blocky debris in a matrix of gravel and silt. The Brainard Ice Cave is a narrow crevice about 10 feet in length between two large blocks.

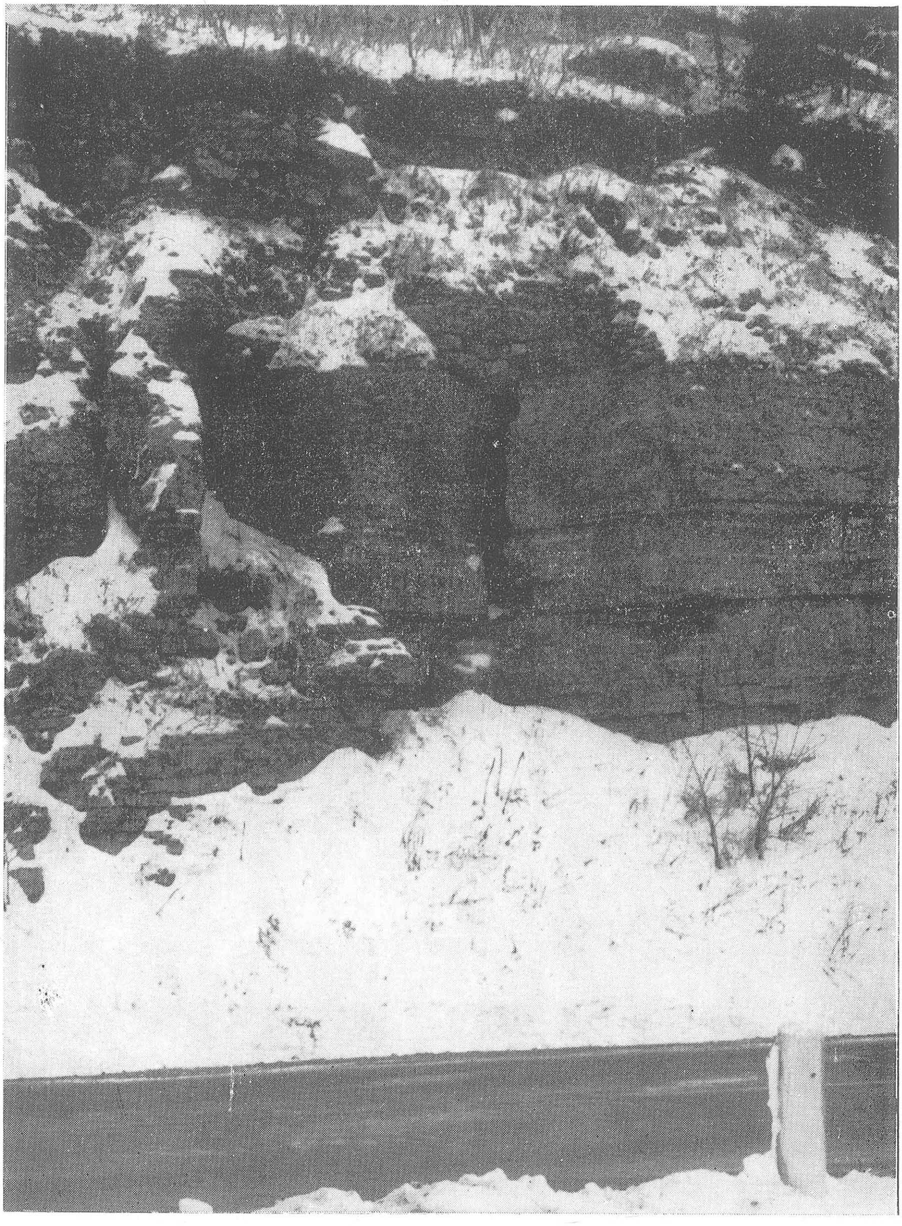
THE WEST UNION TALUS

The West Union talus (Hedges, 1967, pp. 86–88) is a morphostratigraphic unit produced by the progressive collapse and retreat of cavern entrances during several stages of the Pleistocene, from the Kansan to the Present. Pre-Iowan portions of the deposit are covered by alluvial terrace materials. The West Union talus is related by time of origin and by its connexion with frost action to the slope colluvium which clothes the face of the Niagara escarpment throughout its length. However, West Union talus is, specifically, the chaos of blocks formed by cavern collapse and the development of retreatal gorges. It is not a congeliturbate, although frost action may have disturbed it after deposition, and it is not colluvium.

It is perhaps significant that retreatal gorges and West Union talus are associated only with caves which are developed parallel beneath minor surface valleys and whose entrances, consequently, open beneath ephemeral waterfalls. Meltwater gathered in these valleys, upon trickling down over ledges above the cave entrances may have enhanced frost wedging at these entrances relative to entrances not so located.



Pl. 3. Mossy Cave: south passage, looking inward



Pl. 4. Entrance to Route Three Cave

If frost wedging had brought about the retreat of these entrances, one would expect that the rapidity of their retreat would be related to the directions in which they face. This is not so. The north-facing mouth of Searryl's Cave has not retreated at all since Yarmouthian time, while the mouth of Soward's Cave, which also faces the north, has retreated some 500 feet since Iowan time. Factors such as the thickness of the span between cave roof and valley floor above, the frequency of partings within the rock, and the length of the span of the cave entrance apparently are more important in controlling the rate of retreat than is the direction of exposure of the entrance.

CAVERN BREAKDOWN

The mouths of caves in the Upper Mississippi Valley region commonly are much larger in cross-sectional area than are the passages beyond. Bretz (1956, p. 345) attributed the enlarged mouths of Missouri caves to frost action. Periglacial frost action probably was also responsible for the enlarged mouths of caves in the Upper Mississippi valley.

In the caves of northern Europe, periglacial frost wedging produced extensive frost breccias (Flint, 1957, pp. 196, 216; Gèze, 1965. pp: 73–74; Schmid, 1958, p. 16; Trombe, 1952, p. 294; Warwick, 1962c, p. 109) but in the caves of the Upper Mississippi Valley region thermoclastic breccias are not found for any distance beyond the entrances of the caves. Here, as elsewhere in temperate United States, cave breccias usually were produced as a result of loss of buoyant support by cave roofs during the transitions of the caves from water-filled to air-filled conditions (White and White, 1963). This is well shown by straigraphic sections in caves of the Maquoketa River valley. Maquoketa River valley occupies much of the area immediately south of the Niagara escarpment (Hedges and Darland, 1963, pp. 301–303).

One possible exception is Dancehall Cave, Jackson county. The virtual absence of solutional features in the main passage of Dancehall Cave while solutional features are profusely developed in the principal side passages of the cave may be due to their having been destroyed in the main passage by frost spalling. No point in the main passage is more than 275 feet from an entrance at the present time. The cave stream freezes in winter over much of its length even under the present relativelly mild climate. If a thermoclastic breccia once was present in Dancehall Cave, it may since have been removed by stream action or by excavating done in connexion with development of the cave for tourists.

CAVES IN THE SLOPE COLLUVIUM AND THE SOLIFLUCTION APRONS

Smith (1949) described block fields near the Blue mounds and in the Baraboo range, parts of which have been washed free of their silt matrix by small streams. He, and J. F. Quinlan, Jr., have suggested that the entrances to the expanded joints and the small caves of the slope colluvium and of the solifluction aprons, respectively, were developed by soil piping (personal communications, 1968)[3].

Piping, of course, is not a periglacial phenomenon. The development of piping caves in the "head" and in the solifluction aprons has taken place since glacial times.

Like the ice in the Decorah Ice Cave, the ice which is found in the piping caves melts each autumn and forms anew each spring. It is not left over from the Ice Age.

AGES OF THE PERIGLACIAL FEATURES

It seems likely that each of the several kinds of periglacial features may include examples of widely disparate ages. Tundra climatic conditions must have accompanied the southward advance of each glacier. Typically periglacial processes of weathering and erosion should have occurred in areas adjacent to the ice during each glacial stage. Although each type of periglacial feature which I describe, except the West Union talus, may be and is discussed as though all members of the group are of the same age, it is very likely that additional field work will reveal examples which are of different ages.

METHODS OF DATING

Stratigraphy

The solifluction aprons are not overlain by loess. Many areas of "head" are loess-free, also, as are the block streams. A post-Peorian age is thus indicated for most of these features.

The jumble of debris which occupies the lower portions of the expanded joints is not known to include stratigraphic divisions or distinctive materials by which these fissure fills might be correlated with specific members of the Pleistocene section. Straw (1966, p. 374) thought that the unfilled openings at Griersville Rock, Ontario are post-periglacial in age while the filled openings are of periglacial age. According to this reasoning, filled openings developed

[3] For a general review of soil piping, see Parker (1963).

while cryoturbation (hence, downslope movement of regolith materials into the fissures) was yet in progress. Unfilled openings, therefore, must have developed after cryoturbation had ceased.

That some of the fissures are completely filled while others are not does suggest that some of the joints may have been expanded more recently than others. However, if as seems likely, the openings were produced by ice wedging, the presence of ice in them at times when cryoturbation was active would have prevented their becoming filled by regolith materials. All of the fills must be post-periglacial in age, but not all of the fills may postdate the same interval of periglacial activity. That is to say, some fissures may have developed during a pre-Valders glaciation and were filled by later soil movements while other fissures are of Valderan age (according to Straw, all are of Valderan age) and are as yet unfilled.

Radioactivity

Radio-carbon dating has not been employed in the Niagara escarpment area. In the absence of clear stratigraphic divisions in the fills of the expanded joints, and because cryoturbation has affected many of the surficial periglacial deposits, any radio-carbon dates which might be obtained from these materials would not be very dependable.

Fossils

A few of the expanded joints contain modern skeletal remains such as *Bos* and *Canis*. Detailed paleontological explorations have not been made in any of them. I am not aware that fossils have been recovered from any of the periglacial deposits.

Topographic relationships

Known exposures of the West Union talus occur in valleys ranging in age from late-Wisconsinan to Sangamon. In places as yet undiscovered, the talus may extend backward in time as far as the Kansan stage. The West Union talus thus transgresses time. It is the only type of periglacial feature presently known to do so.

Expanded joints and "head" are found along valleys excavated as recently as the Sangamon stage. Ice-wedging and cryoturbation must have occurred, then, as recently as the Wisconsin stage in the Niagara escarpment area.

Relationships with drift sheets

Had the localities in which these periglacial features occur been covered by a glacier subsequently to their development, one would expect drift from that glacier to have accumulated in the expanded joints and to overlie the periglacial deposits. Because the fissures do not contain drift[4], and because the periglacial deposits are not buried beneath later morainic accumulations, the periglacial features must be younger than the youngest drift in their respective areas.

All of the Niagara escarpment in Iowa was glaciated in Nebraskan time (Trowbridge, 1966). The boundary of the Kansan drift sheet is roughly coincident with the escarpment north to the Clayton county line, beyond which it lies east of the escarpment[5]. Most of the Niagara escarpment in northwestern Illinois probably was glaciated in Kansan or Nebraskan time, as indicated by the presence of very old drift near Hanover (Trowbridge and Shaw, 1916, pp. 86–88) and of Kansan drift near Savanna (Shaffer, 1954, p. 446).

Illinoian drift in Illinois mantles the Silurian outliers in Stephenson and in eastern JoDaviess counties, and in all but the extreme northwestern part of Carroll county. No Illinoian drift occurs in southwestern Wisconsin. A map by Donn, Farrand, and Ewing (1962, Fig. 3) showing Illinoian glaciation of the "Driftless Area" is incorrect (Farrand, pers. comm. 1968). Illinoian drift in Iowa does not occur much north of Clinton county.

Glaciers of Wisconsinan ages did not encroach upon the Niagara escarpment in Iowa except near Savanna, where Tazewell drift occupies part of the area between the Goose Lake channel and the Mississippi. The border of the "Iowan drift" is roughly coincident with the escarpment at and northwest of Delaware county, and lies only a short distance south of it between Delaware county and the Mississippi. However, Ruhe, *et al.* (1965) rather thoroughly explain away the "Iowan drift" and replace it with an "Iowan

---

[4] It might be supposed that during times of glaciation, the expanded joints would have been protected from filling with drift by ice wedges developed in them before the glacier arrived, or by frozen, impermeable ground above them. This is not necessarily so, for the ground beneath glaciers is not normally frozen. Glacier ice insulates the ground from sub-freezing temperatures at the surface and previously-frozen ground is thawed by geothermal heat. Horn (1947) has shown that solution and the development of extensive caverns may occur in limestone beneath glaciers.

[5] Ruhe considers all of the older drift east of the Iowan border in northeastern Iowa to be of Kansan age because, as has long been thought, the Nebraskan and Kansan drifts in Iowa are lithologically indistinguishable (Ruhe and Scholtes, 1959, Fig. 1; Ruhe, Rubin, and Scholtes, 1957, p. 672). Separation of the two drifts customarily is predicated on an intervening paleosol, on the degree of weathering of surface exposures, or, in northeastern Iowa, on topographic position. However, Schmaltz (1962) has found lithologic differences between the two drift sheets in north-central Missouri.

surface". They state (p. 11), "We... emphasize that it is a subaerially evolved feature and that glacier ice is unequivocally precluded from any association with it". The voluminous earlier literature upon the Iowan must be disregarded.

There seems to be little doubt that the "Driftless Area" of southwestern Wisconsin, in which are located the Sinsinawa, Platte, and Blue mounds outliers, was glaciated (Black, 1960, 1965). Black has tentatively assigned the scattered patches of drift and the other evidences of glaciation found there to the Rockian substage (early Wisconsinan). The details of the advance, especially its relationship with the mounds, have not, however, been established. Wisconsinan glaciers in Illinois are not known to have encroached upon the Niagara escarpment.

The distribution of the drift sheets shows that all of the expanded joints in Iowa must be of post-Nebraskan age; those in and north of Delaware county must be post-Kansan in age. The expanded joints in Illinois are of post--Nebraskan or post-Kansan age (none are known to exist in the area of Illinoian drift). Expanded joints in Wisconsin (the Platte Mound caves) probably are post-Rockian in age.

The mantles of "head", the associated solifluction aprons, and the block trains are not overlain by any later glacial deposits. They must be at least as young as the expanded joints.

AGES OF THE PERIGLACIAL FEATURES

The age of the most recent ice invasion in any given area places a limit upon the maximum age which the unfilled fissures and surficial periglacial deposits in these areas may have. It may be possible to use drift relationships in a slightly different way to obtain highly precise dates for the development of the periglacial features.

Wright (1961, pp. 939–940) concluded that because periglacial phenomena in Europe are found largely in areas marginal to drift sheets, or below a younger drift in which periglacial features are absent, cold climates cause glaciation and that climatic amelioration begins before retreat of the ice. Periglacial features, therefore, may develop during glacial advances but cannot develop during the retreatal phases of glaciations. This hypothesis is borne out by the distribution of periglacial phenomena in the Driftless Hill Land[6].

[6] Paleotemperature data indicate that surface waters in the North Atlantic ocean became continuously colder throughout the Wisconsin epoch. Donn and Ewing (1966, p. 1708) concluded that a cold climate remained for several thousand years after deglaciation – for about 5000 years in the case of the Wisconsinan. Brunnschweiler (1962, p. 25), in order to explain the sub-drift forest beds of Ohio, adopted the viewpoint that cold climates

Relationships to terminal moraines

Periglacial tundra climates are most severe and of longest duration within a belt a few miles wide just beyond the ice border at glacial maxima. This is the place, and time, most favorable for ice-wedging and for mass movement. It is interesting to observe that many of the best-developed periglacial phenomena in the Upper Mississippi Valley region lie within such narrow belts adjacent to the terminal moraines of the middle- and late-Wisconsinan glaciers.

Of the 22 known occurrences of ice-wedge casts in Wisconsin, one is adjacent to the Valders border, nine are adjacent to the Cary border, and 12 although more than 20 miles from the Cary border are less than 50 miles from it (Black, 1965, p. 188). The notable talus accumulations in the Devil's Lake area end abruptly at the Cary moraine of the Green Bab lobe, suggesting that they developed in ice-free areas during the maximum stand of the ice. The block trains on the Blue mounds and the block fields in the Baraboo range are near the Cary moraine, also.

Involutions in Grundy county, Illinois are associated with the Valparaiso (Cary) moraine system (Sharp, 1942). Horberg's (1949) possible fossil ice wedge in Bureau county, Illinois is associated with the Bloomington moraine of Cary age. Involutions in Henry and Peoria counties, Illinois are associated with the Shelbyville (Tazewell) moraine (Frye and Willman, 1958). Expanded joints in northwestern Illinois all are located within a few miles of the border of the Tazewell drift.

In Iowa, the frost-riven bluffs at Eldora lie scarcely a mile from the Cary moraine of the Des Moines lobe. Expanded joints near Sabula, Preston, and Spragueville all are near the Tazewell margin.

The Woodfordian substage

Where periglacial features are closely associated with a particular terminal moraine, as at Devil's lake and Eldora, there can be scarcely any doubt that their origin coincided in time with the building of the moraine. Unfortunately, such intimate relationships seldom are found. Many of the periglacial phenomena of the Upper Mississippi Valley region lie 10 miles, 20 miles, or even further from the nearest terminal moraine.

follow glaciations, rather than preceding them: "The dated spruce forests of Ohio, evidently still alive when the Wisconsin ice first reached them, are difficult to explain in a periglacial environment. That a periglacial climate became effective shortly thereafter, however, can be interpreted from strong solifluctional evidence in the same area". As noted earlier, (p. 96), the question of the temporal relationship of cold climates to glacial maximae is as yet unresolved.

We have seen, however, that all of the periglacial phenomena of Wisconsin are necessarily post-Rockian (post-Altonian, or post-early-Wisconsinan) in age. From the axiom of uniformitarianism, we may suppose that those in the adjacent portions of Illinois and Iowa also are post-Rockian in age.

Only four known occurrences of periglacial-like phenomena are necessarily of Valderan (late-Wisconsinan) or Recent ages. These are: (1) fossil ice-wedge localities nos. 1 and 6 in Wisconsin (Black, 1965, p. 218), which are associated with the Valders moraine; (2) a local deposit near Trempealeau (Squire, 1908, p. 270) which post-dates the adoption by Mississippi river of the narrow channel between Trempealeau mountain and the west wall of the Mississippi trench, probably in Mankato time; and (3) an involution layer near North Branch, eastern Minnesota, which has been dated by radio-carbon analysis at 2520 years BP (Wright, 1961, p. 943).

All other periglacial features in the Upper Mississippi Valley region may be assigned to the Woodfordian (middle-Wisconsinan) substage. The Woodfordian substage, as presently defined, includes the Tazewell, Cary, and Mankato advances – stratigraphically, all of the units lying above the Farmdale soil and below the Two-Creeks forest bed (Frye, Willman, and Black, 1965, p. 51).

Of the three Woodfordian advances, the Cary was the most extensive. During the Cary maximum, about 20,000 years ago, the Driftless Hill Land and adjoining areas were bracketed by glaciers of the Green Bay and Des Moines lobes southward as far as latitude N 41° 45′. Within the region thus enclosed, climatic conditions were unusually severe. In fact, during the entire Woodfordian substage, "... for 17,500 years, from 30,000 to 15,500 years ago, no clue has been found that trees grew in the state" of Wisconsin. "It is assumed the climate was too cold and dry. ... Probably most ice wedges grew during that time as permafrost and buried ice of the Rockian advance were not destroyed in southern Wisconsin until after the Valders glaciation" (Black, 1964a, p. 26). It therefore seems likely that the expanded joints of the Niagara escarpment, the "head" and the solifluction aprons, and the other periglacial landforms and soil structures of the Driftless Hill Land and its environs are almost all of Woodfordian age.

A final note

Ruhe and Scholtes (1959, p. 591), on the basis of radiocarbon dates, concluded that Cary ice of the Des Moines lobe advanced 32 miles from the area of the Altamont moraine to Des Moines and subsequently retreated 107 miles from Des Moines to the Algona moraine in 1000 years. This is, roughly, the length of time available for the development of Rattlesnake Cave in the frost-riven bluff at Eldora.

Rattlesnake Cave is of about the same width as are the expanded joints in the Niagara escarpment, which thus need not have required more than 1000 years for their development, either. Corroborating evidence is at hand in northern Alaska, where modern ice wedges have a (measured) average increase in diameter of one millimeter per year. Even should the expansion of the joints in the Niagara escarpment have required more than 1000 years, the 17,500 years of the Woodfordian substage would have been entirely sufficient to have encompassed their development.

CONCLUSIONS

Detailed searches of the literature on caves and of the literature on periglacial phenomena have not been made. It is not known how frequently expanded joints analogous in origin to the expanded joints of the Niagara escarpment may occur in other regions.

Caves formed by ice-wedging in joint openings parallel with bluff faces should, however, be found everywhere that functionally analogous physiographic contexts occur. The requisite conditions are three: (1) a jointed, bluff-forming rock immediately underlying which is a plastic shale or another easily-deformed rock; (2) steep-walled valleys incised through the bluff-former into the yielding member; and (3) a region which has or which formerly had a periglacial tundra climate.

Other periglacial phenomena along the Niagara escarpment find many parallels in other regions. "Head", solifluction aprons, block trains, and thermoclastic cavern breakdown are ubiquitous wherever periglacial tundra climates formerly were effective.

Accumulations analogous in origin to the West Union talus probably occur in retreatal gorges associated with caves underlying minor valleys in other formerly periglacial regions. However, where caves are not developed parallel beneath minor valleys, and in perpetually warm regions beyond the limits of former tundra periglacial climates, it seems unlikely that such deposits could have developed.

Periglacial features in the Upper Mississippi Valley region are widely developed in materials of Altonian (early Wisconsinan) ages. They generally are absent from deposits of Valderan (late Wisconsinan) ages. These characteristics, together with the geographic concentration of periglacial phenomena within the area surrounded by middle-Wisconsinan drifts of the Des Moines and Green Bay lobes, suggests that periglacial phenomena in the Upper Mississippi Valley region are largely Woodfordian (middle Wisconsinan) in age.

However, direct, unequivocal evidence that the periglacial features along the Niagara escarpment are Woodfordian in age is lacking. Expanded joints are found along valleys excavated as recently as the Sangamon stage. West Union talus occurs in valleys younger than the Iowan. Slope colluvium and solifluction aprons evidently have formed since deposition of the Peorian loess. At least some of the periglacial features are, therefore, quite young, but their exact ages can only be inferred from data in other parts of the region.

ACKNOWLEDGEMENTS

Credit is due a great many individuals, members and friends of the Iowa Grotto (chapter) of the National Speleological Society who have assisted me with my field work over the past 11 years. Special notice must be accorded Mr. and Mrs. Aldrich Zobac of Cedar Rapids, Iowa who recorded the locations of many of the expanded joints during the course of their mushroom- and rock-hunting excursions in northeastern Iowa. I am also indebted to George Darland, now of Empire City, Minnesota, who for several years was a frequent partner of mine in exploration and in theoretical discussions.

The proximate stimulus for the preparation of this report was furnished by Fred Dorheim of the Iowa Geological Survey. Prof. Emeritus A. C. Trowbridge, The University of Iowa, kindly criticized a preliminary version of the manuscript. Assistance also was given by R. F. Black, W. E. Davies, W. R. Farrand, H. T. U. Smith, and others.

Bibliographie

Barnett, S., 1967 – Trip report. (Iowa Grotto, Nat. Speleol. Soc.) *Intercom,* vol. 3; p. 16, 26, 33.

Bates, R. E., 1939 – Geomorphic history of the Kickapoo Region, Wisconsin. *Geol. Soc. America, Bull.,* vol. 50; p. 819–880.

Beyer, S. W., 1899 – Geology of Hardin County. *Iowa Geol. Survey, Annual Reports,* vol. 10; p. 241–316.

Black, R. F., 1957 – Pleistocene climatic change recorded by ice-wedge polygon casts of Cary Age at River Falls, Wisconsin. *Geol. Soc. America, Bull.,* vol. 68; p. 1888–1889.

Black, R. F., 1960 – „Driftless Area" of Wisconsin was glaciated. *Geol. Soc. America, Bull.,* vol. 71; p. 1827.

Black, R. F., 1961 – Periglacial phenomena of Wisconsin, North-Central United States. *Abstracts of papers, VIth INQUA Congress, Warsaw 1961;* p. 87–88.

Black, R. F., 1964a – Periglacial phenomena of Wisconsin, North-Central United States. *Report of the VIth INQUA Congress, Warsaw 1961,* vol. 4; p. 21–28.

Black, R. F., 1964b – Periglacial studies in the United States 1959–1963. *Biuletyn Peryglacjalny,* no. 14; p. 5–29.

Black, R. F., 1965 – Ice-wedge casts of Wisconsin. *Wisc. Acad. Sci., Arts and Letters,* vol, 54; p. 187–222.

Black, R. F. and Wittry, W. L., 1959 – Pleistocene Man in South-Central Wisconsin. *Geol. Soc. America, Bull.,* vol. 70; p. 1570–1571.

Bretz, J. H., 1956 – Caves of Missouri. *Mo. Geol. Survey,* 2 nd series, vol. 39; 490 p.

Bertz, H. and Harris, S. E., Jr., 1961 – Caves of Illinois. *Ill. State Geol. Survey, Reports of Investigations,* no. 215; 87 p.

Brod, L. G., Jr., 1964 – Artesian origin of fissure caves in Missouri. *Nat. Speleol. Soc., Bull.,* vol. 26; p. 83–114.

Brown, C. E. and Whitlow, J. W., 1960 – Geology of the Dubuque South Quadrangle, Iowa-Illinois. *U.S. Geol. Survey Bull.,* no. 1123–A; 93 p.

Brunnschweiler, D., 1962 – The periglacial realm in North America during the Wisconsin Glaciation. *Biuletyn Peryglacjalny,* no. 11; p. 15–27.

Brunnschweiler, D., 1964 – Der Pleistozäne Periglazialbereich in Nordamerica. *Ztschr. f. Geomorph.,* N.F., Bd. 8; p. 223–231.

Cailleux, A., 1943 – Fissuration de la craie par le gel. *Bull. Soc. Géol. France,* sér. 3, vol. 13; p. 511–520.

Calvin, S., 1894 – Maquoketa shales in Delaware County. *Iowa Acad. Sci., Proc.,* vol. 2; p. 40–42.

Calvin, S. and Bain, H. F., 1899 – Geology of Dubuque County. *Iowa Geol. Survey, Annual Reports,* vol. 10; p. 381–666.

Campbell, N. P., 1968 – The role of gravity sliding in the development of some Montana Caves. *Nat. Speleol. Soc., Bull.,* vol. 30; p. 25–29.

Chamberlin, T. C., 1897 – Editorial footnote. *Jour. Geol.,* vol. 5; p. 825–827.

Davies, W. E., 1961 – Polygonal features on bedrock, North Greenland. *U.S. Geol. Survey, Professional Papers,* no. 424; p. D218–D219.

Dillon, L. S., 1956 – Wisconsin climate and life zones in North America. *Science,* vol. 123; p. 167–173.

Donn, W. L. and Ewing, M., 1966 – A theory of the Ice Ages III. *Science,* vol. 152; p. 1706–p. 1712.

Donn, W. L., Farrand, W. R. and Ewing, M., 1962 –Pleistocene ice volumes and sea-level lowering. *Jour. Geol.,* vol. 70; p. 206–214.

Flint, R. F., 1957 – Glacial and Pleistocene geology. New York City; 553 p.

Frye, J. C. and Willman, H. B., 1958– Permafrost features near the Wisconsin Glacial Margin in Illinois. *Amer. Jour. Sci.,* vol. 256; p. 518–524.

Frye, J. C., Willman, H. B. and Black, R. F., 1965 – Outline of Pleistocene geology of Illinois and Wisconsin. *in:* H. E. Wright and D. G. Frey (ed.) – The Quaternary of the United States. *Intern. Assoc. Quaternary Research, 7th Congress;* p. 43–62.

Gèze, B., 1953 – La génèse des Gouffres. *1er Congress Intern. Spéléol.,* vol. 2; p. 1–13.

Gèze, B., 1965 - La spéléologie scientifique. Paris; 190 p.

Grant, U. S. and Burchard, E. F., 1907 – Lancaster-Mineral Point Folio. *U.S. Geol. Survey, Atlas Folio,* no. 145; 14 p.

Groom, G. E. and Coleman, A., 1958 – The geomorphology and speleogenesis of the Dachstein Caves. *Cave Research Group, Occ. Publ.,* no. 2; 23 p.

Hedges, J., 1961 – Briefer notes. (Iowa Grotto, Nat. Speleol. Soc.) *Iowa Cave Book,* vol. 6; p. B25.

Hedges, J., 1963a – Big Breathing Hole. (Pittsburg Grotto, Nat. Speleol. Soc.) *Netherworld News,* vol. 11; p. 168.

Hedges, J., 1963b – Sampson Ice Cave. (Pittsburg Grotto, Nat. Speleol. Soc.) *Natherworld News,* vol. 11; p. 169.

Hedges, J., 1967 – Geology of Dutton's Cave, Fayette County, Iowa. *Nat. Speleol. Soc. Bull.,* vol. 29; p. 73–90.

Hedges, J. and Darland, G. W., Jr., 1963 – The Scotch Grove Strath in Maquoketa River Valley. *Iowa Acad. Sci. Proc.,* vol. 70; p. 295–306.

Heyl, A. V., *et al.,* 1959 – The geology of the Upper Mississippi Valley Zinc-Lead District. *U.S. Geol. Survey, Professional Papers,* no. 309; 310 p.

Horberg, L., 1949 – A. fossil ice wedge in Bureau County, Illinois. *Jour. Geol.,* vol. 57. p. 132–136.

Horn, G., 1947 – Karsthler i Nordland. *Norges Geol. Undersök.,* no. 165; 77 p.

Howard, A. D., 1963 – The development of karst features. *Nat. Speleol. Soc., Bull.,* vol, 25; p. 45–65.

Howell, J. V., 1935 – The Mississippi River Arch. *in:* A. C. Trowbridge (editor) – Guide book, 9th Annual Field Conference, Upper Mississippi Valley. Kansas Geol. Soc. Wichita, Kansas; p. 386–389.

Koteff, C., 1961 – A frost-wedged bedrock locality in southeastern Massachusetts. *U.S. Geol. Survey, Professional Papers,* no. 424; p. C57–C58.

Kovarik, F., 1898 – The Decorah Ice Cave and its explanation. *Scientific American Suppl.,* no. 1195; p. 19158–19159.

Lees, J. H., 1926 – Geology of Crawford County. *Iowa Geol. Survey, Annual Reports,* vol. 32; p. 239–362.

Leffingwell, E. de. K., 1915 – Ground ice wedges, the dominant form of ground ice on the North Coast of Alaska. *Jour. Geol.,* vol. 23; p. 635–654.

McGee, W. J., 1891 – The Pleistocene history of northeastern Iowa. *U.S. Geol. Survey, Annual Reports,* vol. 11; p. 189–577.

Martin, J., 1928 – Zum Problem der Glazialen Denudation und Erosion. *Abhandl. Naturwiss. Verein zu Bremen,* Bd. 26; p. 461–480.

Michael, E. D., 1965 – Origin of the lakes in the Chuska Mountains, northwestern New Mexico: Discussion. *Geol. Soc. America, Bull.,* vol. 76; p. 267–268.

Norton, W. H., 1911 – Underground water resources of Iowa. *Iowa Geol. Survey, Annual Reports,* vol. 21; p. 29–1214.

Parker, G. G., 1963 – Piping, a geomorphic agent in landform development of the drylands. *Berkeley, Intern. Assoc. Sci. Hydrology, Extract Publication,* no. 65; p. 103–113.

Peltier, L. C., 1949 – Pleistocene terraces of the Susquehanna River, Pennsylvania. *Pa. Geol. Survey, 4th series, Bull.* G-23; 158 p.

Potter, J. H., 1961 – Trip report. (Wisc. Speleol. Soc. Grotto, Nat. Speleol. Soc.) *WSS Newsletter,* vol. 2; no. 6; p. 2.

Raby, B., 1957 – Trip log. (Iowa Grotto, Nat. Speleol. Soc.) *Iowa Cave Book,* vol. 1, article no. 1957-8-7 and map file.

Ruhe, R., V., Rubin, M. and Scholtes, W. H., 1957 – Late Pleistocene radiocarbon chronology in Iowa. *Amer. Jour. Sci.,* vol. 255; p. 671–689.

Ruhe, R. V. and Scholtes, W. H., 1959 – Important elements in the classification of the Wisconsin glacial stage: A discussion. *Jour. Geol.,* vol. 67; p. 585–593.

Ruhe, R. V., *et al.,* 1965 – The Iowan problem. *Midwestern Friends of the Pleistocene, 16 th Annual Meeting;* 22 p.

Sardeson, F. W., 1897 – On glacial deposits in the Driftless Area. *Amer. Geologist,* vol. 20; p. 392–403.

Savage, T. E., 1905 – Geology of Jackson County. *Iowa Geol. Survey, Annual Reports,* vol. 16; p. 565–648.

Schafer, G. M., 1953 – A disturbed buried Gumbotil Soil profile in Jefferson County, Iowa. *Iowa Acad. Sci., Proc.,* vol. 60; p. 403–407.

Schafer, J. P., 1962 – Pleistocene frost action in and near Northeastern Iowa. *Geol. Soc. America, Special Papers,* vol. 68; p. 262.

Schmaltz, L. J., 1962 – Pebble lithology of Nebraskan and Kansan tills in North-Central Missouri. *Geol. Soc. America, Special Papers,* vol. 68; p. 262–263.

Schmid, E., 1958 – Höhlenforschung und Sedimentanalyse. *Basel, Schriften des Inst. f. Ur- und Frühgeschichte der Schweiz,* no. 13; 186 p.

Schrader, W. D. and Hussey, K. M., 1955 – Evolution of the level interfluvial divides on the Kansan Till Plain in Iowa and Missouri. *Iowa Acad. Sci., Proc.,* vol. 60; p. 408–413.

Shaffer, P. R., 1954 – Extension of Tazewell Glacial Substage of Western Illinois and Eastern Iowa. *Geol. Soc. America, Bull.,* vol. 65; p. 443–456.

Sharp, R. P., 1942 – Periglacial involutions in Northeastern Illinois. *Jour. Geol.,* vol. 50; p. 113–133.

Shimek, B., 1895 – A theory of the loess. *Iowa Acad. Sci., Proc.,* vol. 3; p. 82–89.

Smith, H. T. U., 1949 – Periglacial features in the Driftless Area of Southern Wisconsin. *Jour. Geol.,* vol. 57; p. 196–215.

Smith, H. T. U., 1953 – Periglacial frost wedging in the 'Rock Cities' of southwestern New York. *Geol. Soc. Americe, Bull.,* vol. 64; p. 1474.

Smith, H. T. U., 1962 – Periglacial frost features and related phenomena in the United States. *Biuletyn Peryglacjalny,* no. 11; p. 325–342.

Squire, G. H., 1897 – Studies in the Driftless Region of Wisconsin. *Jour. Geol.,* vol. 5; p. 825–836.

Squire, G. H., 1898 – Studies in the Driftless Region of Wisconsin II. *Jour. Geol.,* vol. 6; p. 182–192.

Squire, G. H., 1899 – Studies in the Driftless Area of Wisconsin. *Jour. Geol.,* vol. 7; p. 79–82.

Squire, G. H., 1908 – Peculiar local deposits on Bluffs adjacent to the Mississippi. *Wisc. Acad. Sci., Arts and Letters, Trans.,* vol. 16; p. 258–274.

Straw, A., 1966 – Periglacial mass movement on the Niagara Escarpment near Meaford, Grey County, Ontario. *Geogr. Bull.,* vol. 8; p. 369–376.

Thoreau, H. D., 1966 – Walden and civil disturbance. New York City, 424 p.

Thornbury, W. D., 1954 – Principles of geomorphology. New York City, 618 p.

Thwaites, F. T., 1935 – Physiography of the Baraboo District, Wisconsin. *in:* A. C. Trowbridge, (editor) – Guidebook, 9th Annual Field Conference. Kansas Geol. Soc., Wichita, Kansas; p. 395–404.

Trewartha, G. T. and Smith, G.-H., 1941 – Surface configuration of the Driftless Cuestaform Hill Land. *Annals Assoc. Amer. Geographers,* vol. 31; p. 25–45.

Trombe, F., 1952 – Traité de spéléologie. Paris; 376 p.

Trowbridge, A. C., 1914 – Preliminary report on geological work in Northeastern Iowa. *Iowa Acad. Sci., Proc.,* vol. 21; p. 205–209.

Trowbridge, A. C., 1934 – Upper Mississippi Valley structure. *Geol. Soc. America, Bull.,* vol. 45; p. 519–528.

Trowbridge, A. C., 1966 – Glacial drift in the "Driftless Area" of Northeastern Iowa. *Iowa Geol. Survey, Report of Investigations,* no. 2; 28 p.

Trowbridge, A. C. and Shaw, E. W., 1916 – Geology and geography of the Galena and Elizabeth Quadrangles. *Ill. State Geol. Survey, Bull.,* vol. 26; 233 p.

Twidale, C. R., 1956 – Vallons de gélivation dans le Centre du Labrador. *Rev. Géomorph. Dyn.,* vol. 7; p. 17–23.

Twidale, C. R., 1958 – Vallons de gélivation dans le Centre du Labrador. *Rev. Géomorph. Dyn.,* vol. 9; p. 84.

Warwick, G. T., 1962a – Caves and rocks. *in:* C. H. D. Cullingford (editor) – British Caving. London, 2 nd edition; p. 11–33.

Warwick, G. T., 1962b – British Caving Regions. *in:* C. H. D. Cullingford (editor) – British Caving. London, 2nd edition; p. 120–217.

Warwick, G. T., 1962c – Cave formations and deposits. *in:* C. H. D. Cullingford (editor) – British Caving. London, 2nd edition; p. 83–119.

Watson, R. A. and Wright, H. E., Jr., 1963 – Landslides on the east flank of the Chuska; Mountains, northwestern New Mexico. *Amer. Jour. Sci.,* vol. 261; p. 525–548.

Weidman, S., 1904 – The Baraboo Iron-Bearing District of Wisconsin. *Wisc. Geol. Nat. History Survey, Bull.,* no. 13; 190 p.

White, C. A., 1870 – Report on the geological survey of the State of Iowa. Des Moines. vols. 1 and 2.

White, E. L. and White, W. B., 1963 – Processes of cavern breakdown. *Nat. Speleol. Soc NEWS,* vol. 21; p. 17.

Willman, H. B. and Reynolds, R. R., 1947 – Geological structure of the Zinc-Lead District of northwestern Illinois. *Ill. State Geol. Survey, Reports of Investigations,* no 124; 15 p.

Wilson, L. R., 1958 – Polygonal structures in the soil of Central Iowa. *Okla. Geol. Notes* (Okla. Geol. Survey), vol. 18; p. 4–6.

Woodward, H. P., 1961 – A stream piracy theory of cave formation. *Nat. Speleol. Soc., Bull.,* vol. 23; p. 39–58.

Wright, H. E., Jr., 1961 – Late Pleistocene climate of Europe: A review. *Geol. Soc. America, Bull.,* vol. 72; p. 933–984.

Wright, H. E., Jr., 1964 – Origin of the lakes in the Chuska Mountains, northwestern New Mexico. *Geol. Soc. America, Bull.*, vol. 75; p. 589–598.

Wright, H. E., Jr., 1965 – Origin of the lakes in the Chuska Mountains, northwestern New Mexico: Reply. *Geol. Soc. America, Bull.*, vol. 76; p. 269–270.

Yehle, L. A., 1954 – Soil tongues and their confusion with certain indications of periglacial climate. *Amer. Jour. Sci.*, vol. 252; p. 532–546.