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## Kaolinite From Warsaw Geodes, Keokuk Region, Iowa

JOHN B. HAYES<sup>1</sup>

*Abstract.* White, compacted masses of kaolinite are common in the famous geodes from the upper mudstone unit of the lower Warsaw Formation around Keokuk, Iowa. X-ray diffraction studies of the fine silt-size, pseudohexagonal grains show the kaolinite to be exceptionally well crystallized and free from impurities. Stacking faults are produced by grinding the kaolinite, which have a marked effect on X-ray patterns. The relationship of kaolinite to other mineral euhedra in geode cavities indicates that kaolinite formed early in geode history. Several lines of evidence suggest that the kaolinite originated from the fine-grained silicate residue remaining after acid dissolution of early diagenetic calcareous concretions, the progenitors of geodes. The kaolinite formed in an acid environment into which aluminum was introduced to react with the layer silicate "wreckage" from the concretions.

Geodes from the Mississippian Warsaw Formation near Keokuk, Iowa, have long attracted the fancy of geologists, mineralogists, and amateur enthusiasts. In addition to being abundant, the geodes are noted for containing in their cavities well formed crystals of a variety of minerals, about twenty in number. Papers by Van Tuyl (1925), Tripp (1959), and Hayes (1961) deal with the mineralogy of Warsaw geodes, the emphasis being mostly on description. It is the purpose of the present paper to examine the crystallography and genesis of one of the common geode minerals, kaolinite. Also, it is hoped to draw attention to an Iowa occurrence of exceptionally pure and well crystallized clay mineral.

### GEOLOGIC SETTING

Strata exposed around Keokuk, Iowa, include the Burlington, Keokuk, Warsaw, Salem, and St. Louis formations, all marine rocks of Mississippian age. Figure 1 presents a composite columnar section of the strata. Some geodes occur in thin, argillaceous dolomite beds near the top of the Keokuk Formation, but the bulk are found in the lower member of the Warsaw Formation as defined by Van Tuyl (1925).

In the Keokuk area the lower Warsaw can be divided into three lithologically distinct and laterally traceable units. The lowest averages 12 feet in thickness and consists of massive, argillaceous dolomite which, at some places, is interbedded with

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Stratigraphic Classification                      Composite Columnar Section

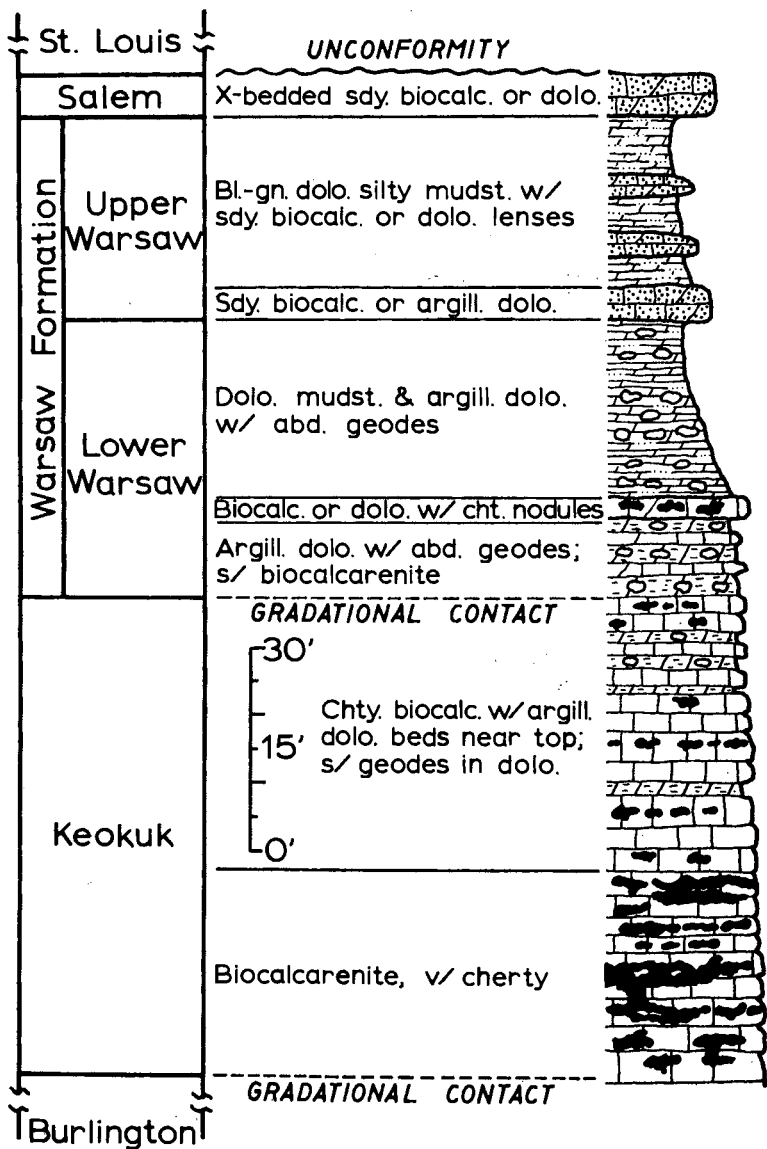


Figure 1. Classification and composite columnar section of Mississippian rocks near Keokuk, Iowa. Kaolinite occurs almost exclusively in geodes of the upper mudstone unit of the lower Warsaw.

biocalcarenite. Geodes are confined to dolomite beds and usually are large, rounded, and hollow, being well developed in a sense. The middle unit, about 3 feet thick, may be either argillaceous dolomite or biocalcarenite, characterized by nodules and lenses of chert; geodes are lacking. The upper unit is predominantly dolomitic mudstone, poorly laminated and indurated. The unit is about 24 feet thick in most exposures. Geodes are abundant in this mudstone unit, but they differ from those below in being smaller, flattened, and with little or no cavity. It is important to note here that kaolinite-bearing geodes are confined almost exclusively to the upper mudstone unit of the lower Warsaw.

#### GENERAL DESCRIPTION AND PREVIOUS INVESTIGATIONS OF GEODE KAOLINITE

Kaolinite occurs in solid geodes as compacted masses of snow-white, earthy powder, usually surrounded by coarsely crystalline calcite in the lower part of geode specimens. If the geode has a cavity, there may be a thin layer of kaolinite on the cavity floor, or the cavity may be completely, but loosely, filled with the white powder. The chalcedony shells of geodes from the mudstone unit of the lower Warsaw have tiny pockets of kaolinite throughout.

Photomicrographs show the kaolinite to be well formed plates with pseudohexagonal symmetry (Figure 2). Such crystals range in size from about 4 to 7 microns in long dimension. Vermicular aggregates of crystals (Figure 3) are common, and are several microns long. Examination of thin sections suggests that the

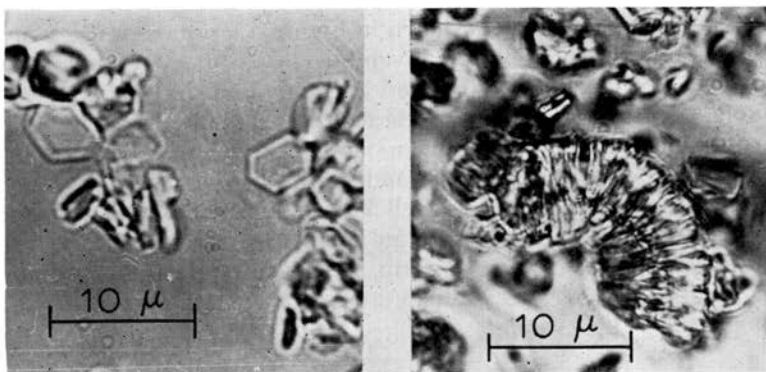


Figure 2. Photomicrograph of pseudo-hexagonal kaolinite crystals from Warsaw geode cavity. Ordinary light.

Figure 3. Photomicrograph of vermicular aggregate of kaolinite crystals from geode. The aggregate long axis is perpendicular to (001). Ordinary light.

kaolinite crystals and aggregates are in completely random orientation within the mineral pockets.

A number of mineralogists have made specific reference to

kaolinite of the Warsaw geodes. Ross and Kerr (1931, p. 159), in their classic study of the kaolin minerals, listed as a specimen of dickite, “. . . . fine-grained glistening white powder filling the cavity of a geode . . . . from vicinity of Keokuk, Iowa (U.S. National Museum No. 80551).” The present writer examined the numbered specimen while on a visit to the National Museum and found it to be a typical lower Warsaw geode. The writer obtained a sample of the white powder, which proved to be, by X-ray diffraction analysis, well crystallized kaolinite.

In a paper discussing dickite in Missouri, Tarr and Keller (1936) stated that they visited the Keokuk area in hopes of finding the material described as dickite by Ross and Kerr (1931). Tarr and Keller reported finding an abundance of the white powder in geodes, but that it all was kaolinite, according to X-ray identification. In a later paper, Tarr and Keller (1937) suggested that the kaolinite in Warsaw geodes was deposited from solutions in the cavities.

The possibility of dickite occurring in geodes of the Keokuk area is remote, considering that Tarr and Keller (1936, 1937) found only kaolinite, that the present writer found only kaolinite, and that the single specimen originally described as dickite by Ross and Kerr (1931) is kaolinite.

Striking relationships exist between kaolinite and crystals of other minerals with which it may be associated in geode cavities. Crystals of sphalerite, calcite, dolomite, and barite commonly have grown into pockets of kaolinite and ingested quantities of the clay mineral. Entire pockets have been incorporated into calcite in this manner. As shown in Figure 4, crystals of the same generation of calcite in a given geode assume scalenohedral habit where kaolinite is ingested and rhombohedral habit where kaolinite is lacking. This suggests that kaolinite predates the calcite and other ingesting minerals, and formed early in geode history. The ability of calcite and barite to form “sand crystals” by cementing sandstone is well known; in geodes is a parallel example of calcite ingesting large quantities of kaolinite and yet exerting the calcite crystal form. X-ray diffraction photographs of the scalenohedral crystals (Figure 4) show strong patterns of both calcite and kaolinite.

#### CRYSTALLOGRAPHY OF GEODE KAOLINITE

The writer examined numerous X-ray diffraction patterns of geode kaolinites and found them to be essentially identical to one another. Patterns were obtained with Straumanis-mount, 114.6 mm diameter, Debye-Scherrer powder cameras using Ni-filtered Cu radiation. Specimens to be irradiated were prepared by jamming the open end of glass capillary tubes into

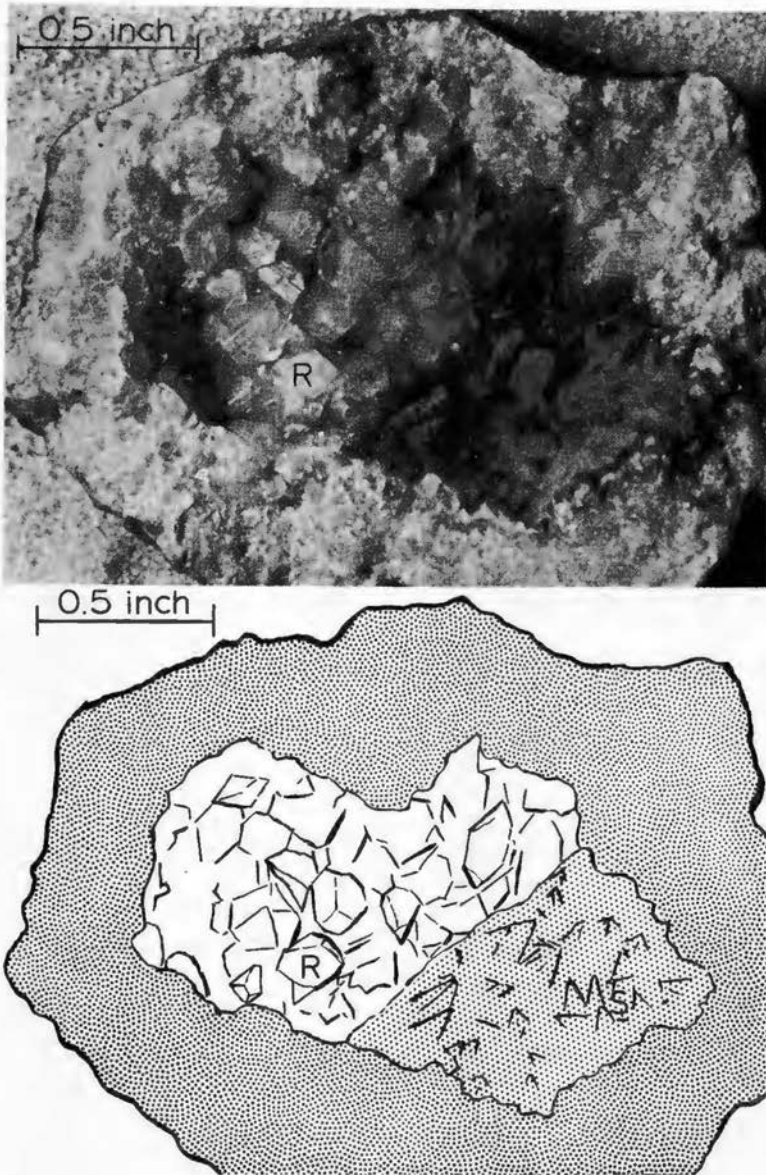


Figure 4. Geode specimen to show calcite-kaolinite relationships. Rhombohedral (R) calcite crystals contain no kaolinite, whereas scalenohedral (S) crystals are intergrown with a pocket of kaolinite, the dark powder.

pockets of naturally compacted kaolinite and withdrawing in the ends of the tubes cores of kaolinite with crystal orientation duce satisfactory powder patterns. Preferred orientation effects were not apparent on powder patterns, indicating random orientation of kaolinite crystals in geode cavities.

The atomic structure and chemical composition of kaolinite have been the subjects of much detailed study. A summary may be found in Brindley and Robinson (1946), although a brief review may be in order here.

Kaolinite has the composition  $2[Al_2Si_2O_5(OH)_4]$  per unit cell; numerous chemical analyses and crystallographic studies indicate that most kaolinites scarcely deviate from this ideal composition. Structurally, kaolinite belongs to the layer lattice silicates or phyllosilicates, which accounts for its platy morphology and fine grain size. Kaolinite crystals commonly are fine silt-size, as in Warsaw geodes, or coarse clay-size. The structural unit layer (Figure 5) is composed of a sheet of  $Si^{+4}$  in tetrahedral coordination with  $O^{-2}$ , joined with a sheet of  $Al^{+3}$  in octahedral coordination with  $O^{-2}$  and  $(OH)^-$ . Such structural unit layers are stacked one upon the other to form crystals. The  $a$  and  $b$

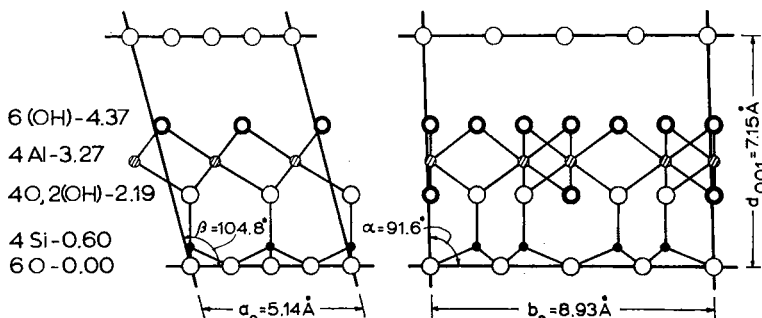


Figure 5. Projections of the unit layer of kaolinite along the  $b$  axis (left diagram) and the  $a$  axis (right diagram). Heights of planes of atoms above the basal oxygen plane are shown in Å, and the proportions of atoms are indicated. After Brindley and Robinson (1946) and Grim (1953).

crystallographic axes of the triclinic unit cell form a plane parallel to the sheets of atoms, with the  $c$  axis slightly inclined to the  $a$ - $b$  plane. The unit cell is only one structural layer thick, with the distance between layers,  $d_{001}$ , equal to 7.15 Å.

Kaolinites from a variety of occurrences deviate widely from perfect three-dimensional crystallinity. Some kaolinites are well crystallized or ordered as to stacking of structural unit layers, whereas others are poorly crystallized or disordered. All degrees of stacking from completely ordered to completely disordered have been observed. Brindley and Robinson (1946) discuss the nature of the disordered stacking. Well crystallized or ordered kaolinites will show diffraction maxima on X-ray powder photographs where the  $k$  index of  $hkl$  is not equal to a multiple of 3 ( $3n$ ), but poorly crystallized kaolinites will lack  $hkl$  diffraction bands where  $k \neq 3n$ .

Other differences between the X-ray diffraction patterns of well

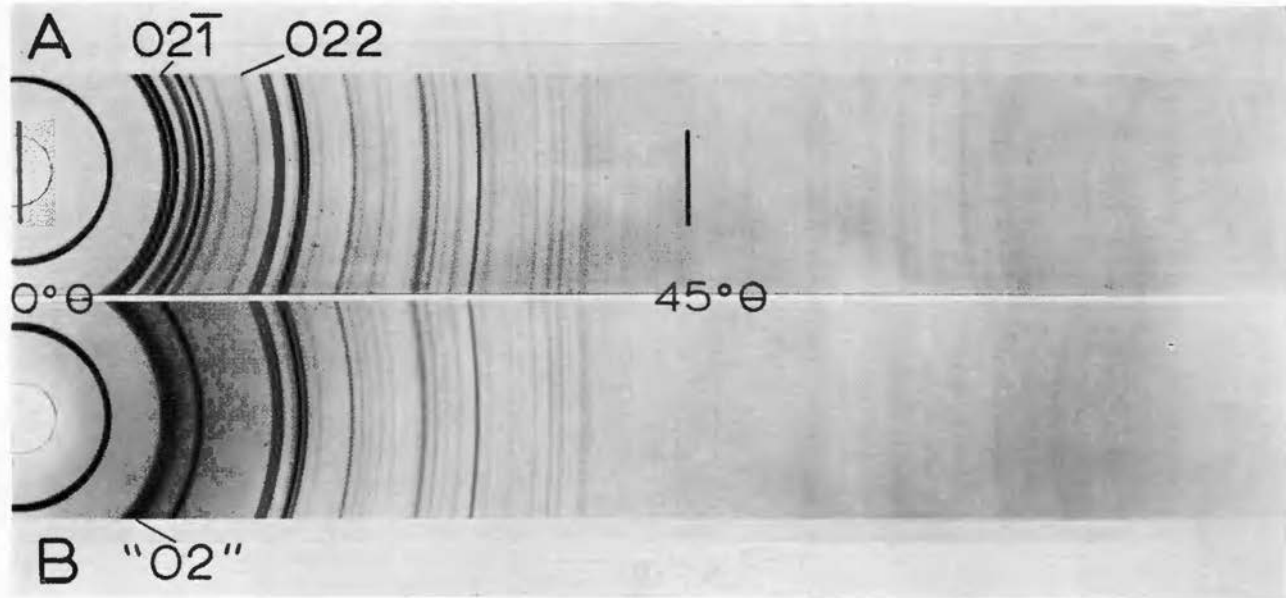


Figure 6. X-ray diffraction film patterns of kaolinite from geode cavity. Cuka, Ni filter, 114.6 mm Straumanismount camera. A. Unground material in random orientation. The  $02\bar{1}$  and  $022$  reflections are indicated as examples of those where  $k \neq 3n$  and which disappear upon grinding of the kaolinite. B. Kaolinite ground lightly by hand in agate mortar. The changes in diffraction effects, such as the appearance of the "02" band, are due to the regularity of stacking of kaolinite layers being disrupted by grinding.



ordered kaolinites versus disordered kaolinites have been given by Murray (1954) as: (1) sharpness of reflections, (2) number of reflections, (3) resolution of closely spaced reflections, (4)  $d$  spacings of the reflections.

The  $d$  values and intensities of reflections were measured for a number of geode kaolinites and compared with the observed values published by Brindley and Robinson (1946, p. 245-248) for the best crystallized kaolinite they could find. Such a comparison shows the geode kaolinite to be exceptionally well crystallized and superior to most mentioned in the literature. Of the 52 reflections listed by Brindley and Robinson (1946), from 001 at 7.15 Å to 006 at 1.188 Å, 49 were observed on the ordinary X-ray patterns of geode kaolinite (Figure 6A). Values of  $d$  and intensities for the reflections were comparable. Hayes (1961, p. 223-224) lists the  $d$  values, intensities, and  $hkl$  of the reflections in tabular form, so they need not be repeated here. Sharpness and resolution of lines also indicate a very high degree of ordered stacking, according to the previously mentioned criteria of Murray (1954). Notice back reflections in Figure 6A, another characteristic of well crystallized material. Workers in need of a pure, well ordered kaolinite may find that in Warsaw geodes suitable.

Some striking changes occur in the X-ray diffraction patterns of geode kaolinite after the mineral is ground lightly by hand in an agate mortar for one minute (Figure 6B). (1) Many reflections at low angle  $2\theta$  either disappear or merge together into broad and diffuse bands. (2) The background at low angles darkens. (3) Many weak reflections disappear. (4) The  $d$  spacings of all remaining reflections increase slightly with respect to the unground mineral. See Hayes (1961, p. 223-224) for tabular data.

Grinding of geode kaolinite produces a material similar in characteristics to disordered or poorly crystalline kaolinite. Random shifts of structural layers one over the other are induced parallel to the  $a$ - $b$  plane. The shifts are parallel to the  $b$  axial direction and are integral multiples of  $b_0/3$ . Such shifts in position can satisfy the requirement for  $(OH)^-$  to  $O^{2-}$  bonding between layers, but the random displacements destroy the repeat symmetry or crystallinity along the  $c$  axis; hence, the term poorly crystalline or disordered. All the reflections which disappear upon grinding, such as  $02\bar{1}$ ,  $022$ ,  $042$ , are of the type,  $k \neq 3n$ , where  $n$  is an integer (Figure 6). Most all of the remaining reflections can be assigned Miller indices where  $k = 3n$ . The broad, two-dimensional "02" band may result from the combined effects of random displacement of layers along the  $b$  axis and rotation of layers in the  $a$ - $b$  plane (Murray, 1954, p. 101).

The ground geode kaolinite is not as disordered as the *Puga D*

naturally disordered kaolinite described by Robertson, Brindley, and Mackenzie (1954), but it is probable that grinding time longer than one minute would increase disorder. The nature of disordered kaolinite and the effects of grinding upon kaolinite are documented in the literature, but it should be said here that workers investigating kaolinite crystallinity probably will alter that which they hope to measure by mechanically disturbing the crystals.

#### GEODES AND CONCRETIONS

The nature and origin of Warsaw geodes and calcareous concretions are discussed by Hayes (in press). A few points relevant to geode kaolinite need be mentioned here.

Warsaw geodes formed from early diagenetic calcareous concretions by a complex process involving recrystallization, replacement, solution, and precipitation. Geode cavities resulted from acid dissolution of the calcite cores of concretions. Some geodes still contain remnants of original concretion calcite, but none of that kind contains kaolinite. The concretions incorporate small amounts of detrital silicates, including clay minerals; concretions from the upper mudstone unit of the lower Warsaw contain more non-carbonate material than do concretions in the lower unit of the lower Warsaw (Figure 1). Clay minerals in the concretions are the same as those in the mudstone, namely, illite, chlorite, and randomly interstratified illite-montmorillonite (Hayes, 1963). Kaolinite is significantly absent from both geode-bearing mudstones and concretions.

#### ORIGIN OF GEODE KAOLINITE

Facts pertaining to geode kaolinite origin are now summarized. (1) Kaolinite occurs only in geodes from the upper mudstone unit of the lower Warsaw. (2) Concretions from which kaolinite-bearing geodes developed contained substantial amounts of silicate minerals, including clay minerals, but no kaolinite. (3) Kaolinite is absent from those geodes which still have remnants of original concretion calcite. (4) No other clay minerals, such as those found in concretions, occur with kaolinite in geodes. (5) The ingestion of kaolinite by calcite euhedra (Figure 4) and other minerals indicates that kaolinite genesis predates growth of all other minerals in geode cavities and was very early in geode history.

The writer believes that the kaolinite formed by some mechanism from the clay minerals left as residue after dissolution of the calcite concretion core. The same acid solutions which dissolved the calcite to create a cavity also reacted with the silicate residues to form kaolinite. It has been stated many times in the literature that kaolinite is the product of acidic environments at temperatures less than 300°C; for example, see Noll (1936).

Kaolinite formation did not begin until virtually all the original calcite was dissolved. This is because the pore fluid, likely groundwater, could not build up concentrations of hydrogen ions (achieve low pH) as long as calcite remained in the system. Grim (1953, p. 343) states: "In the weathering of calcareous sediments, there is substantially no alteration of the silicates until the carbonate is completely broken down and the calcium removed from the environment." The absence of kaolinite, a silicate alteration product, from geodes still containing original concretion calcite is significant evidence.

The suggestion of Tarr and Keller (1937) that the geode kaolinite precipitated from solution is probably an oversimplification, for such a process could occur in any geode cavity regardless of the absence or presence of residual clay material. The present writer's evidence strongly points to a genetic relation between the residual clay minerals and kaolinite. The exact mechanism for converting illite, chlorite and perhaps quartz to kaolinite is problematical. In any event, after complete removal of concretion calcite, the rock pore fluid must have remained quite acidic. Silica is relatively insoluble in acidic solutions and mildly alkaline solutions, but becomes markedly soluble above pH 9 (Krauskopf, 1959). Alumina is practically insoluble between pH 4 and pH 9.5, but increases abruptly in solubility in strongly acid and alkaline systems (Correns, 1949, p. 210). At pH 4, and below, relatively little silica and much alumina are mobile. The silicon to aluminum ratio of kaolinite is 1:1; for illite and chlorite, Si:Al is greater but variable. Thus, to produce kaolinite from illite and chlorite would require relative enrichment of alumina, either by addition of alumina or removal of silica. Also, many species of metallic cation would need to be placed into solution and perhaps removed from the geode cavity.

In a moderately strong acid system, alumina is much more mobile than silica. Alumina must have been brought by acid solution into the geode cavity where it reacted with material derived from illite and chlorite, to form kaolinite. The amount of kaolinite in most geodes is as great or greater than the amount of insoluble residue remaining after laboratory dissolution of calcite concretion cores.

The relative importance of colloidal and ionic effects in geode kaolinite formation is unknown. It can be said that the illite and chlorite must have been reduced to an almost amorphous state by the acid solutions removing cations from their structures. Kaolinite then probably formed by constructional crystallization processes from the interaction of ionic and amorphous matter. The spatial distribution and orientation of silicon-oxygen tetrahedral sheets in illite and chlorite differs from that of kaolinite

to the point where the former could not pass to the latter simply by the rearrangement of cations within a fixed anionic framework. The high degree of crystal perfection, the relatively large crystal size, vermicular aggregates, and random orientation of kaolinite all suggest that the constituents of kaolinite were in the mobile ionic and colloidal states, rather than in some solid state.

In summary, kaolinite formed under acid conditions from illite and chlorite left as residue after dissolution of the calcite concretion core. Alumina was probably added to the system. Kaolinite formed by constructional crystallization from amorphous colloidal and ionic matter. Kaolinite formed during the same period of acid weathering which dissolved the concretion cores to produce geode cavities; as such, it predates all minerals which grew as euhedra in the geode cavities.

#### CONCLUSIONS

- (1) Kaolinite found in geodes of the upper mudstone unit of the lower Warsaw is exceptionally well crystallized with ordered stacking of structural layers, according to X-ray diffraction criteria.
- (2) Kaolinite was the first mineral to grow in geode cavities, and did so early in geode history.
- (3) Kaolinite formed in an acid environment from the weathered silicate minerals left behind after acid dissolution of calcareous concretions, such concretions subsequently being transformed into geodes.

#### ACKNOWLEDGMENTS

S. W. Bailey and the late S. A. Tyler, University of Wisconsin, are to be thanked for guidance and encouragement throughout the course of this investigation.

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## A Geophysical Reconnaissance Method for the Location of Buried Valleys

LYLE V. A. SENDLEIN<sup>1</sup>

The purpose of this study was to investigate the feasibility of using the constant depth electrical resistivity method to locate the boundaries of Pleistocene or pre-Pleistocene valleys buried by glacial drift. Instrument stations were located on one-mile centers in order that a large area could be covered in a short period of time. This investigation included a total of 42 stations.

The area investigated is located in the Townships of Yellow Spring Creek (T 72 N, R 3 W) and Washington (T 72 N, R 4 W) in Des Moines County, Iowa. Figure 1 shows the location.

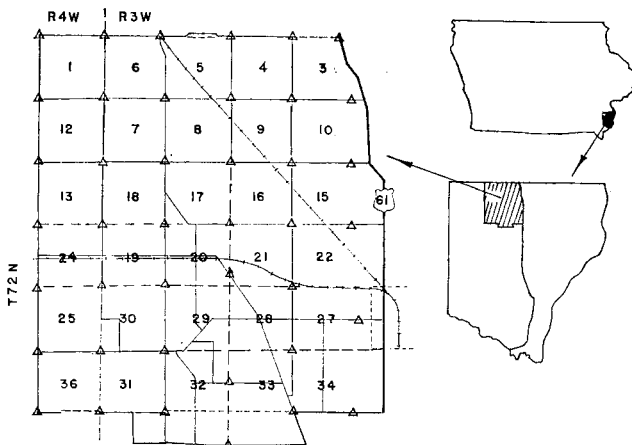


Figure 1. Location of study area.

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## BEDROCK TOPOGRAPHY

A bedrock topographic map of southeastern Iowa was compiled by T. M. Beveridge in 1947<sup>1</sup>. A more recent study of the bedrock topography of Des Moines County was carried out by Mr. R. Campbell, of the Iowa Geological Survey and the author in the summer of 1962. Figure 2 shows the bedrock topography of the study area as obtained from water well data. These data included drillers logs, sample logs, and house to house inventory of wells in critical areas.

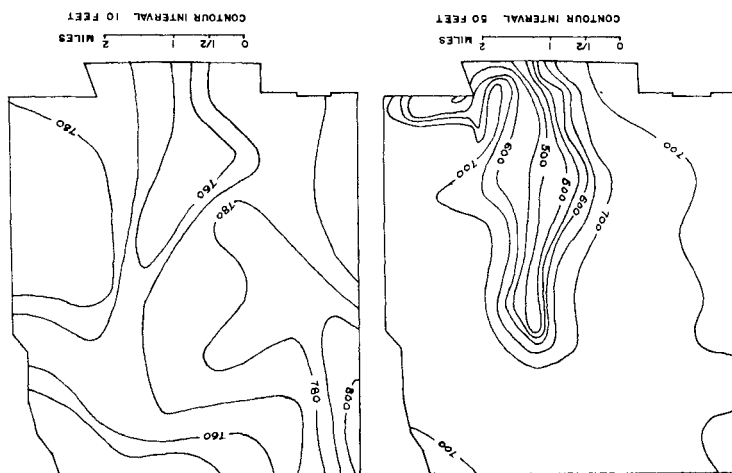


Figure 2. Bedrock topography.  
Figure 3. Surface topography.

Beveridge's map shows two major buried valleys in southeastern Iowa. The smaller one, named the Washington Channel, is located about 15 miles south of the other, the Poweshiek Channel. Both valleys trend southeast to northwest and merge with the present Mississippi River Valley. Beveridge's map was compiled on the data available in 1947. Since that time, many more data have been made available by well drillers, and therefore more details have become apparent. Prior to inundation by the drift, the relief along the stream valleys approaches 200 feet, and the topography must have been similar to the bluff topography observed along the Mississippi River today. Many small, steep-walled tributaries were probably present. The study presented here is of one of the northward extending tributaries of the Washington Channel.

## SURFACE TOPOGRAPHY

In attempting to solve practically any geological problem in Iowa, the investigator is confronted with the non-existence of base maps. This study was no exception. It was necessary to make a generalized topographic map of the study area. Mr. R.

Campbell obtained county road surveys from the Des Moines County Highway Engineers Office in Burlington, Iowa. From these surveys, elevations of many road intersections were obtained. Through use of an engineers level or a hand level, elevations could be carried from these points to any desired location.

The elevations for this study were located by hand level or estimated from the base map. The resultant map is shown in figure 3. This is a very low-order survey, but sufficient for reconnaissance level work.

#### BEDROCK GEOLOGY

The most abundant outcropping rock observed in the county is Mississippian in age. Limited outcrops of Pennsylvanian and Devonian (Maple Mill) are also present. The drift ranges in thickness between 60 and 120 feet on the buried upland surfaces, and 200 to 300 feet over the buried valleys. Mississippian limestones are beneath the drift on the buried upland surfaces and the upper portion of the buried valley walls, whereas Devonian shales are beneath the drift in the valley bottoms. The limestone average thickness is 100 feet, and the Devonian shale is approximately 300 feet thick when not cut by a buried valley. A typical cross section across a buried valley is shown in figure 4.

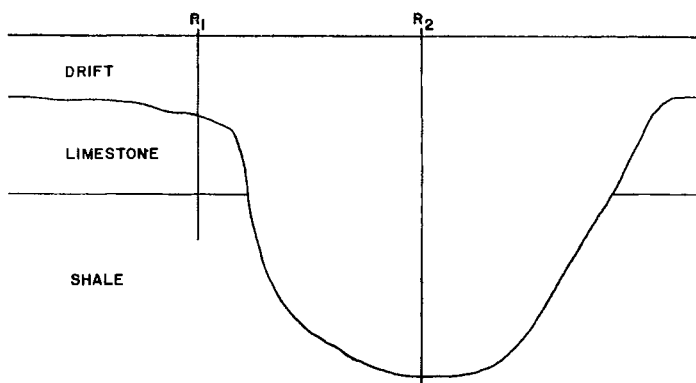


Figure 4. Cross-section of a buried valley.

#### GEOPHYSICAL METHOD

The electrical resistivity method has been described by several authors (2, 5, & 6), and will not be discussed here in detail. Figure 5 shows the distribution of the current and equipotential surfaces in a homogeneous model and figure 6 shows the current distribution in a nonhomogeneous model. The bottom layer in figure 6 has a smaller resistivity, thus a higher conductivity and, therefore, the current lines will be distorted to the path of least

resistance. By moving the electrodes ( $C_1$ ,  $C_2$ ,  $P_1$  and  $P_2$ ) in a systematic manner, the depth to the interface can be determined. When the electrodes are spaced close together (i.e. - "a" is small) the apparent resistivity  $\rho$  approaches  $\rho_1$ , and when the electrodes are widely spaced, (i.e. - "a" is large with respect to the depth to the interface) the apparent resistivity  $\rho$  approached  $\rho_2$ .

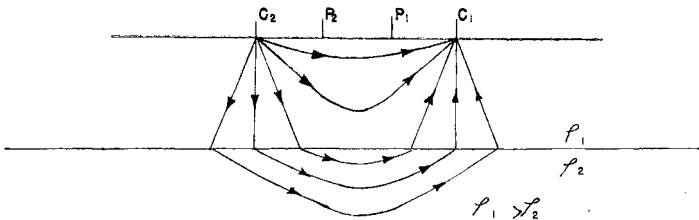
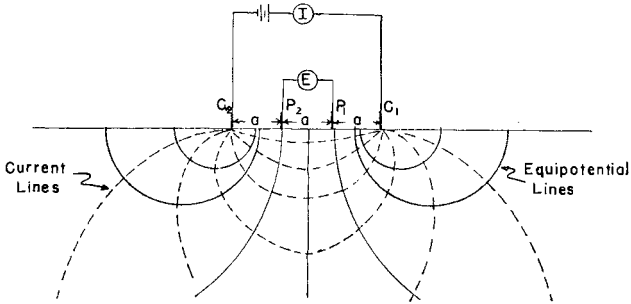


Figure 5. Current flow in a homogenous model.  
 Figure 6. Current flow in a non-homogenous model.

The Wenner<sup>6</sup> configuration of electrodes was employed. Four metal electrodes are placed in the ground, as shown in figure 5. The distance between each electrode is equal to the distance "a". Using this method, the following equation is applicable for the calculation of the apparent resistivity:  $\rho = 2 \pi a \frac{E}{I}$

- where  $\rho$  = apparent resistivity
- $a$  = electrode spacing
- $E$  = potential between  $P_1$  and  $P_2$
- $I$  = current flowing between  $C_1$  and  $C_2$

The instrument used in the survey was constructed by Carl A. Bays and Associates of Urbana, Illinois. It is a Gish-Rooney type<sup>3</sup> which has been modified so that the apparent resistivity ( $\rho$ ) can be read directly for each electrode spacing ( $a$ ). The power source is a 12-volt storage battery which, in this case,



was part of the standard equipment of the vehicle used for transportation of the survey party.

There are two standard field methods which can be employed in an electrical resistivity survey. One requires that the electrodes be separated a constant increment after each reading so that a vertical distribution of the resistivity can be measured at each station. The distance "a" is plotted vs. the apparent resistivity  $\rho$ , with the assumption that "a" approximates the depth of penetration. Vertical discontinuities can be detected in this manner. The second method requires that a fixed "a" distance be selected and only one reading made at each instrument station. This is called the constant depth method, and was the technique employed in this study.

The optimum field procedure is realized when one instrument operator and four helpers, one helper for each electrode, make up the field party. At each station the instrument is placed on the ground, connected to the power source, and then connected to the electrode leads. These, in turn, are connected to the electrodes which have been placed in the ground with the proper "a" spacing. A signal is given to stand clear of the electrodes and the power is turned on. The correct amperage is selected, the galvanometer is zeroed, and the potential observed and recorded. The power is shut off and the crew signalled to roll up the cable and prepare to move to the next station. For a well trained crew, this will require approximately 5 minutes per station, including the move to the next instrument site 1 mile away.

The constant depth principle could be satisfactorily employed here for two reasons. 1) The difference between the resistivities of the unconsolidated drift and the limestone is appreciable. 2) The relationship of the limestone to the drift and shale. In all cases the absence of the limestone indicates a buried valley.

If a depth was chosen so that the potential being measured was well into the limestone, a certain range of apparent resistivity values could be expected. From previous work in this area typical resistivity values were known. Two continuous depth electrical resistivity profiles, which are considered to be typical for each bedrock condition, are shown in figure 7.

The continuous depth resistivity profile which encounters limestone at 62 feet in figure 7 is typical of a section along  $R_1$  in figure 4. Likewise the resistivity profile which does not encounter limestone is typical of profile  $R_2$  in figure 4.

An "a" value of 120 feet was selected for the constant depth method, and the breaking point between limestone and the absence of limestone was taken to be 20,000-OHM/CMS. This value was chosen from experience gained from 150 continuous

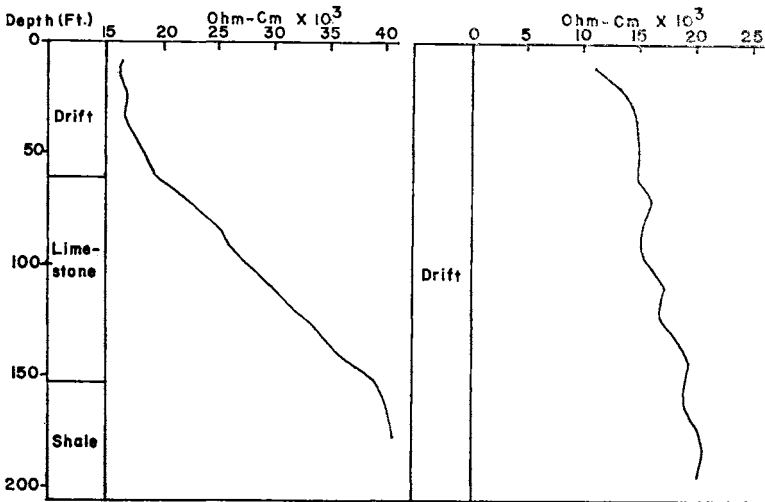


Figure 7. Continuous depth electrical resistivity profile.

depth profile stations occupied in the area around Mediapolis, Iowa, during the same field season.

### RESULTS

The results are presented in figure 8, superimposed over the bedrock topography map obtained from well data. The apparent resistivity readings have been plotted on the individual vertical axes with the horizontal axes representing 20,000-OHM/CM. The reading above this horizontal axis represent limestone at 120 feet depth, and those below the horizontal axis represent drift as the material present at this depth.

A study of this figure indicates that a correlation seems to exist. The readings obtained directly over the buried valley all indicate the absence of the limestone, and the majority of those over the buried upland surface indicate limestone as the bedrock material. However, there are a few exceptions which should be noted. These are the stations with coordinates P4-B, P4-E, P4-F, P2-C, P2-D, P1-B, and P1-C. They are anomolous and will be discussed later.

### DISCUSSION

The data presented shows that this method used as a reconnaissance tool can give valid information which could be followed up by a more detailed study for the location of actual valley boundaries.

The resistivity data suggest that this valley probably extends to the north and west of the presently mapped valley. This would account for the low resistivity readings at P2-C, P2-D,

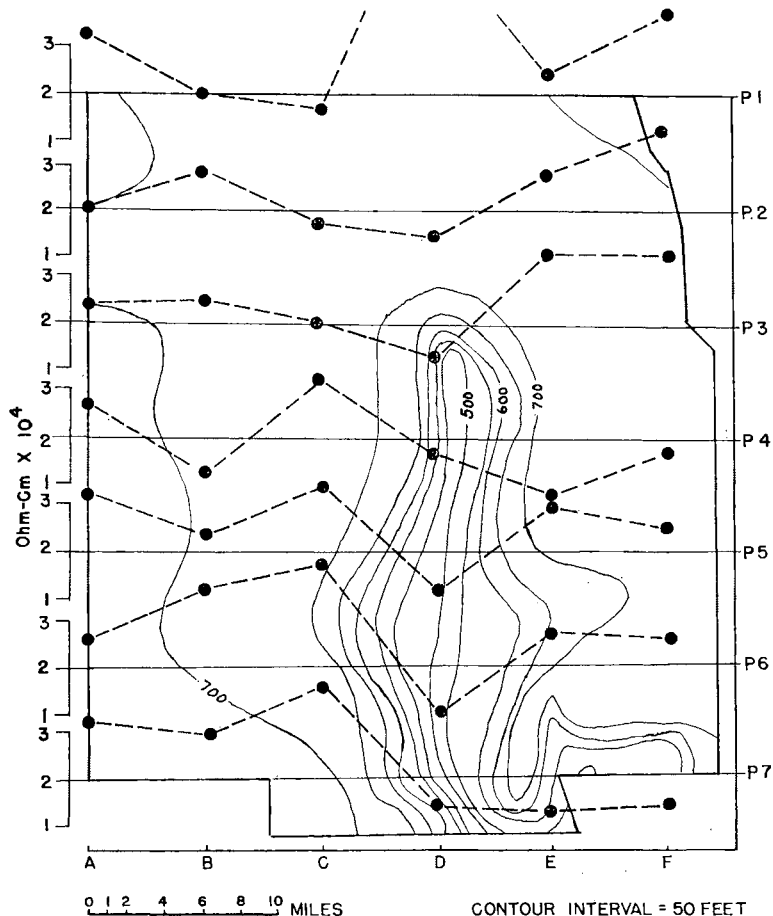


Figure 8. Constant depth cross-sections superimposed on the bedrock topography map.

P1-B and P1-C. There are no water wells in this region to prove or disprove this possibility.

The other anomolus readings occur along profile 4 at stations B, E, and F. A possible explanation is that the limestone may be present but very thin, less than 30 feet, which could indicate tributary valleys. At this point, it might be worth pointing out that the potentials were measured at a constant depth below the earth's surface and not a constant elevation. Therefore, if the elevation varies widely, the depth of measuring will also vary. The constant depth of 120 feet that was used actually falls approximately 30 feet above the contact of the base of the limestone and shale. So, if the limestone thins to less than 30 feet, or if the surface elevation increases and the limestone surface does not, then a low reading will be obtained. These

facts must be kept in mind when interpretations are being made.

The method of investigation is only good when it can be used in conjunction with other known geological information of the area, such as the character and sequence of beds and the geologic structure. Some limitations of this technique for the location of buried valleys are as follows: 1) This method is based on the appreciable difference between the electrical properties of limestone and shale. Therefore, if the proper lithologic sequence is not present, the accuracy of the method decreases and may not work at all. For example, if the limestone was not present and the buried upland surface were shale, the data would have been harder to interpret. The electrical properties of shale and the drift are very similar. 2) If the topographic relief were great, this technique would have to be altered to a constant elevation method. This would not be difficult, but it would require a knowledge of the elevation of the station prior to occupancy, so that the correct depth could be selected at each instrument station. 3) It is essential that some drill hole information be available for the determination of typical resistivity values.

This geophysical technique can be a valuable tool to the geologist when it is applied to problems within its realm of application.

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