STATE OF ILLINOIS
DEPARTMENT OF REGISTRATION AND EDUCATION

RELATIONSHIP OF GRAVITY ANOMALIES TO A DRIFT-FILLED BEDROCK VALLEY SYSTEM IN NORTHERN ILLINOIS

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ABSTRACT

A reconnaissance survey with a gravity meter may aid in the analysis of drift-filled bedrock valley systems. The Troy Bedrock Valley in northern Illinois is cut into bedrock (bulk density ranges from 2.59 gm/cc to 2.82 gm/cc) and is filled with thick glacial deposits (bulk density ranges from 2.06 gm/cc to 2.54 gm/cc). A gravity survey was made of this area and the resultant residual gravity field is marked by negative, elongated lows following the major trends of the valley system.

Drift densities, as determined theoretically, will be highest for mixed grain-size material, such as till, and lowest for uniform grain-size material, such as sand or clay. Analyses of densities determined in the lab from drift cores approximate these results. Where the drift densities are high and bedrock density low, the gravity method cannot be utilized for bedrock valley mapping.

INTRODUCTION

Dissected bedrock topography in areas covered by thick glacial deposits results in a gravitational field strongly influenced by bedrock relief. In gravity studies designed to explore bedrock structures or reefs, this effect may be undesirable and may even negate the use of a detailed gravity survey. When the bedrock topography itself is to be studied, the large density contrasts between the unconsolidated glacial deposits and consolidated bedrock permit the gravity method to be used with some success. Using the gravity meter, Hall and Hajnal (1962) have been able to detect buried valleys cut into both drift and bedrock.

Knowledge of the trends of major buried bedrock valleys is desirable in ground-water studies. Thick and extensive sand and gravel deposits in these valleys would contribute to their importance as potential areas of ground-water development. This paper is a study of the bedrock geology, glacial deposits, and bedrock topography in the Troy Bedrock Valley system (Horberg, 1950) and the relationship of these to the gravity field in northern Illinois. The area (1,080 square miles) is located in Boone, DeKalb, LaSalle, Lee, McHenry, Ogle, and Winnebago Counties and was surveyed in 1961. It is 18 miles wide, borders Wisconsin on the north, and extends approximately 60 miles southward (fig. 1).

The gravity maps and their interpretation were prepared by McGinnis, and geologic studies and interpretations were made by Kempton. The analyses of drift densities were made by Heigold and McGinnis. A number of shallow seismic refraction and resistivity profiles also were made to determine further drift thickness and to identify lithology at a number of locations (McGinnis and Kempton, 1961).

A Worden gravity meter with a dial constant of .0912 milligal/div. was used in the survey. The instrument is temperature compensated, and although the drift rate was high, it was quite linear, increasing at the rate of .71 milligals per day throughout the time span of the field work. The meter has a range of 73 milligals, or 800 dial divisions without resetting and a recommended operating range of 977.358 to 983.569 gals. During the course of the survey, the instrument was reset numerous times with no variation in the rate of drift.

United States Geological Survey topographic sheets were used as base maps from which elevations were obtained. The value of gravity was read at the elevation locations accessible to an automobile. These locations included road intersections, section corners, bridges, and old landmarks. This network of stations resulted in a grid of about 1 mile. Base stations were reoccupied twice daily to provide a continuous drift curve during the course of the survey.

Bedrock topography (pl. 1a) was interpreted from available bedrock elevations obtained from field located well records and bedrock outcrops. The residual gravity map (pl. 1b) was derived with the aid of well records.

Acknowledgments

The geologic interpretations used in this paper are a result of research carried out at the Illinois Geological Survey and are adapted from portions of a doctoral dissertation submitted to the University of Illinois by Kempton. The split-spoon samples used to determine bulk drift densities were obtained from a subsurface sampling program undertaken as a part of a Water Resources Development and Management Study coordinated by the Northeastern Illinois Metropolitan Area Planning Commission.

GEOLOGY

The area in this study contains various topographic and geologic features. These include (1) a deep gorge cut into the bedrock, (2) bedrock uplands covered with thin drift that reflect the bedrock topography, (3) deep bedrock channels partially or completely filled with thick glacial deposits, (4) uplands covered with thick drift that conceal the bedrock surface features, and (5) the morainic ridges that lie over both bedrock uplands and bedrock valleys.

Paleozoic sedimentary rocks of Silurian, Ordovician, or Cambrian age crop out below the glacial deposits or, locally, at land surface where glacial deposits are absent. These rocks, approximately 3,845 feet thick (as indicated by the Edmond S. Wynn oil test in sec. 36, T. 41 N., R. 5 E.), rests on a basement

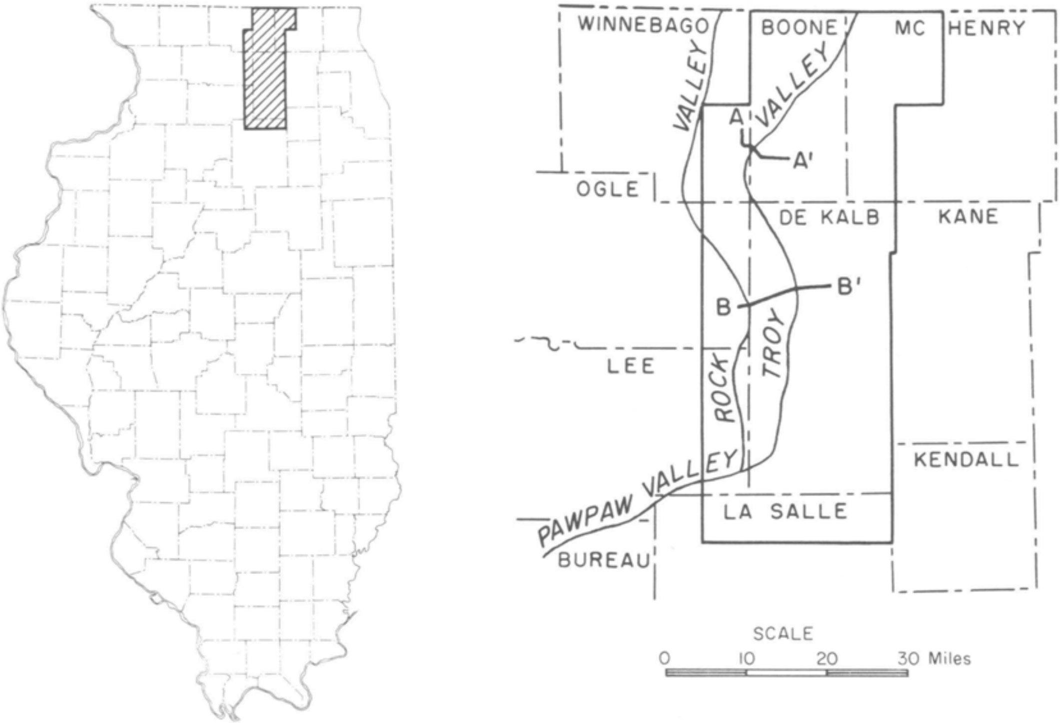

Fig. 1 - Location of area, major bedrock valleys and lines of cross section.

of crystalline rocks of Precambrian age. A summary of the essential features of the bedrock stratigraphy is shown on figure 2.

The oldest rocks at the bedrock surface, under thick drift cover, are the upper two Cambrian formations, Franconia and Trempealeau (fig. 3). Franconia outcrops are limited to areas just south of the Sandwich Fault Zone within the confines of the Ancient Rock, Troy, and Paw Paw bedrock valleys and their tributaries. The Trempealeau Dolomite is present directly below the drift in the area immediately south of the Sandwich Fault Zone, but it is centered primarily in and adjacent to the larger bedrock valleys.

All of the units of the Ordovician System (Templeton and Willman, 1963) that are mapped in northern Illinois crop out at the bedrock surface in one or more localities of the area covered (fig. 3). Rocks of the Prairie du Chien Group only crop out below the drift south of the Sandwich Fault Zone. The Prairie du Chien Group consists of three formations: the Oneota at the base, and the New Richmond and the Shakopee at the top (fig. 2). The Glenwood and St. Peter Formations, considered as a unit (Glenwood-St. Peter) for the purpose of this study, crop out at the bedrock surface both north and south of the Sandwich Fault Zone. The St. Peter is separated from the older rocks by a major unconformity. South of the fault zone, the lower portion of the St. Peter caps the uplands; but in T. 36 N., R. 2 E., LaSalle County, it dips steeply to the west-southwest and is overlain by the Galena-Platteville Dolomite. North of the fault zone, the St. Peter is present directly below the drift in the deeper portions of the Ancient Rock Valley and its tributaries and along segments of the Troy Valley and its tributaries.

System	Series	Group or Formation	Thickness (ft.)	Description
Silurian	Niagaran - Alexandrian	Undifferentiated	0 - 90+	Dolomite, light gray to white, silty at base, locally cherty
Ordovician	Cincinnatian	Maquoketa	0 - 175	Shale, gray to brown; locally dolomite, argillaceous
	Champlainian	Galena	0 - 369+	Dolomite, buff, cherty; shale partings at base
		Platteville		Dolomite and limestone cherty, sandy at base
		Glenwood	0 - 506+	Sandstone and dolomite with locally thick shale
		St. Peter		Sandstone, fine to coarse, locally cherty at base
	Canadian	Prairie du Chien Shakopee	0 - 84	Dolomite, sandy, cherty, oolitic
		New Richmond	0 - 95	Sandstone, some dolomite
		Oneota	0 - 210	Dolomite, white to pink, cherty, oolitic, sandy at base
Cambrian	Croixian	Trempealeau	0 - 190	Dolomite, white, pink to buff, finely crystalline; contains geodic quartz
		Franconia	80 - 130	Dolomite, sandstone, and shale, green to red, glauconitic

Fig. 2 - Stratigraphy of bedrock formations that directly underlie the glacial drift.

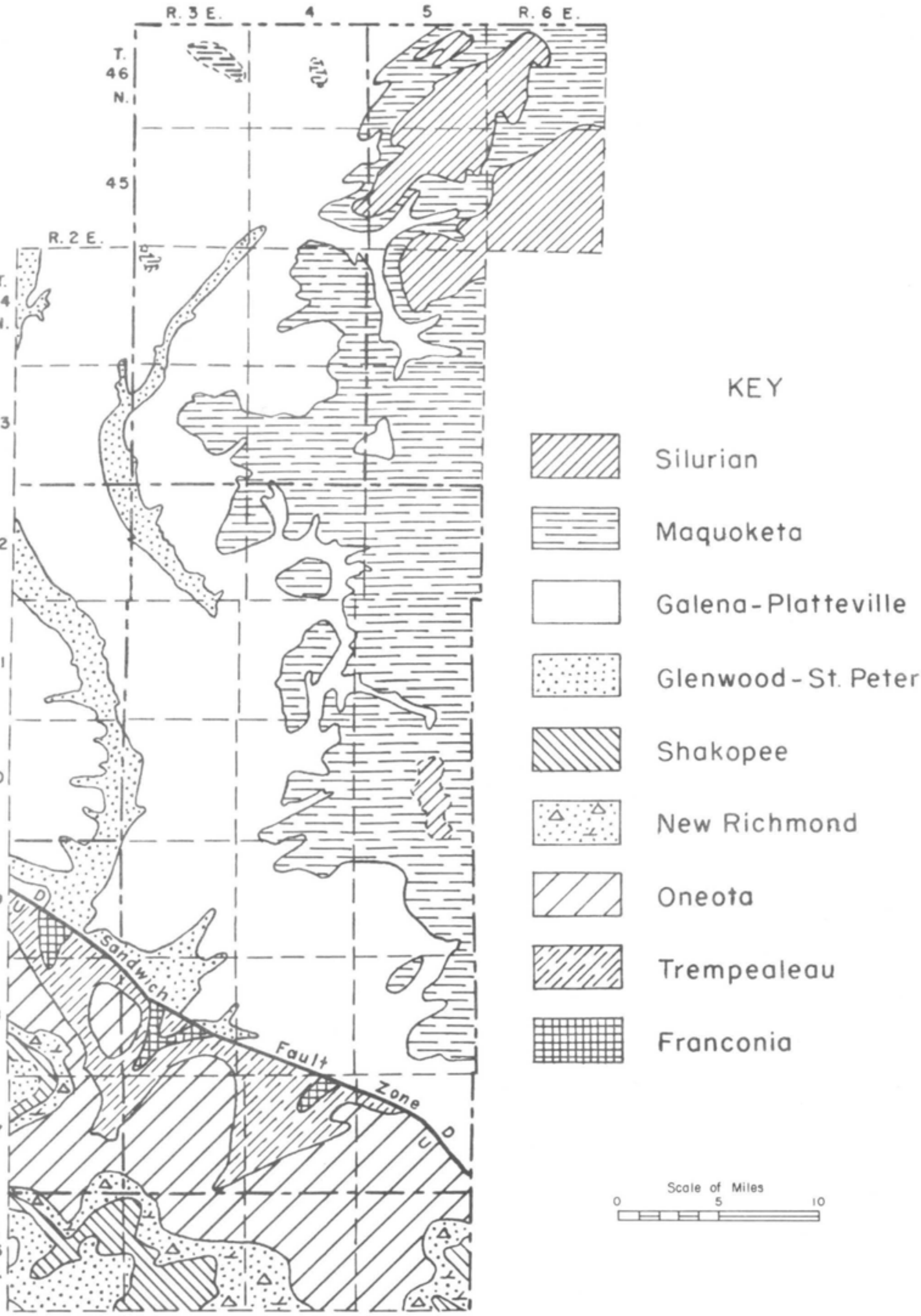

Fig. 3 - Areal geology of the bedrock surface.

A thick section of dolomite of the Galena and Platteville Groups, overlying the Glenwood Formation, is called here the Galena–Platteville Dolomite. It crops out below the drift over a large portion of the study area, north of the Sandwich Fault Zone, and is exposed at numerous localities within the outcrop (fig. 3). These rocks form the walls of the Ancient Rock and Troy Valleys, and some of their tributaries, and also the floor of segments of the Troy Valley.

The uppermost rocks of the Ordovician System in northern Illinois, the Maquoketa Formation, crop out extensively below the drift in eastern DeKalb, southeastern Boone, and western McHenry Counties. The Maquoketa generally consists of an upper green shale, a middle limestone or dolomite, and a lower brown shale.

Rocks of Silurian age occur only in McHenry County where as much as 100 feet of dolomite cap the bedrock uplands. It is likely that all of these rocks belong to the Alexandrian Series.

The major structural features of the area are shown in figure 4 and include the Sandwich Fault Zone, the Ashton Arch, a small portion of the LaSalle Anticline, and numerous minor structural features. The Ashton Arch (Willman and Templeton, 1951) crosses the area with its axis slightly south of the Sandwich Fault Zone. The dip on the southwest flank averages about 50 feet per mile. The crest of the LaSalle Anticline (Cady, 1920) lies just at the southwestern edge of the area in T. 36 N., R. 2 E., LaSalle County, and T. 37 N., R. 2 E., Lee County, as indicated by an average dip of approximately 100 feet per mile on top of the Trempealeau Dolomite.

The Sandwich Fault Zone crosses the area in T. 37 N., T. 38 N., and T. 39 N., DeKalb and Lee Counties (figs. 3 and 4). Based on an estimate of 550 feet of rock between the top of the Trempealeau and the base of the Platteville, an average throw of about 550 feet is indicated; in Tps. 37 and 38 N., R. 4 E., a throw of 750 feet or more is possible. Although more than one fault is present in a surface outcrop to the west in Ogle County, information is not sufficient to indicate faulting other than a zone of major displacement in the study area.

North of the fault zone, the major feature is a generally eastward regional dip of the rocks off the Wisconsin Arch, averaging 13 feet per mile. There are numerous minor structural features in the form of subordinate cross folds that trend generally east-west across the area (fig. 4).

The maximum relief developed on the surface of these bedrock formations, prior to or during the various glacial and interglacial stages, is about 530 feet from the upland in northwestern Boone County (910± feet) to the deepest part of the bedrock valley systems in southwestern DeKalb County (380± feet). Two large, generally parallel, bedrock valleys (Rock and Troy, fig. 1) and numerous smaller, deeply incised, tributary valleys characterize the drainage pattern developed on the bedrock.

The Troy Valley (Alden, 1904, and Horberg, 1950) enters Illinois from Wisconsin about 2 miles east of the Boone-McHenry County line in McHenry County (fig. 1). It is a moderately wide (average 2½ to 3 miles), deep (250-300 feet), generally V-shaped valley carved predominantly into or through the Galena-Platteville Dolomite throughout most of its course. The thalweg of the Troy Valley is less than a mile in width, and in the southern half of the valley it is usually less than one-half mile across (pl. 1).

The stratigraphic evidence in the study area indicates that Pleistocene continental ice sheets have advanced at least seven times into the Troy Valley

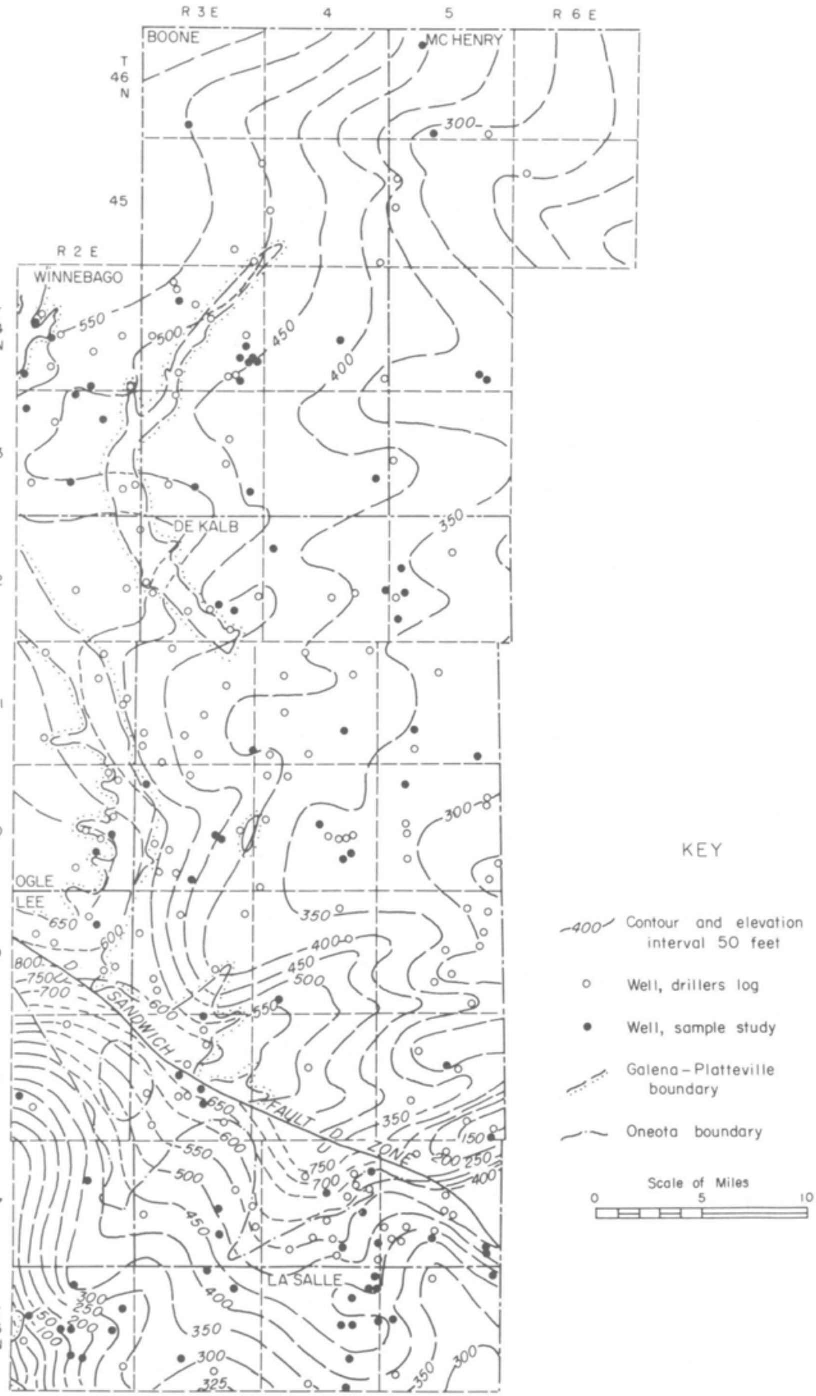

Fig. 4 - Structure, top of Glenwood-St. Peter, north and top of Trempealeau, south of Sandwich Fault Zone.

area, while at least one other moved close enough to allow the deposition of outwash sediments (Kempton, 1962). These glaciers partially or completely filled the Troy Valley and its deeper tributaries with deposits locally in excess of 550 feet thick and averaging about 400 feet thick. Over the bedrock uplands, glacial deposits average about 150 feet thick and reach a local maximum of over 300 feet thick.

Not only the thickest but the most complete sequence of glacial deposits occurring in the study area is found in the Troy Valley (fig. 5). This sequence consists of a basal sand or sand and gravel overlain by two or three distinct till units (Altonian or Pre-Wisconsinan), a locally thick sand and gravel, and by another two or three till units (Altonian and Woodfordian) with some interbedded glacial-lacustrine sediments, and sand and gravel. There are also extensive areas of surficial sand and gravel.

A sufficient density contrast between drift and bedrock must exist everywhere for a successful gravity survey. Bedrock densities of the rocks to be found in this area have been determined in previous studies (Krey and Lamar, 1925; Lamar 1927) and are always greater than 2.59 gm/cc, ranging to 2.82 gm/cc. A detailed study of the densities of till follows.

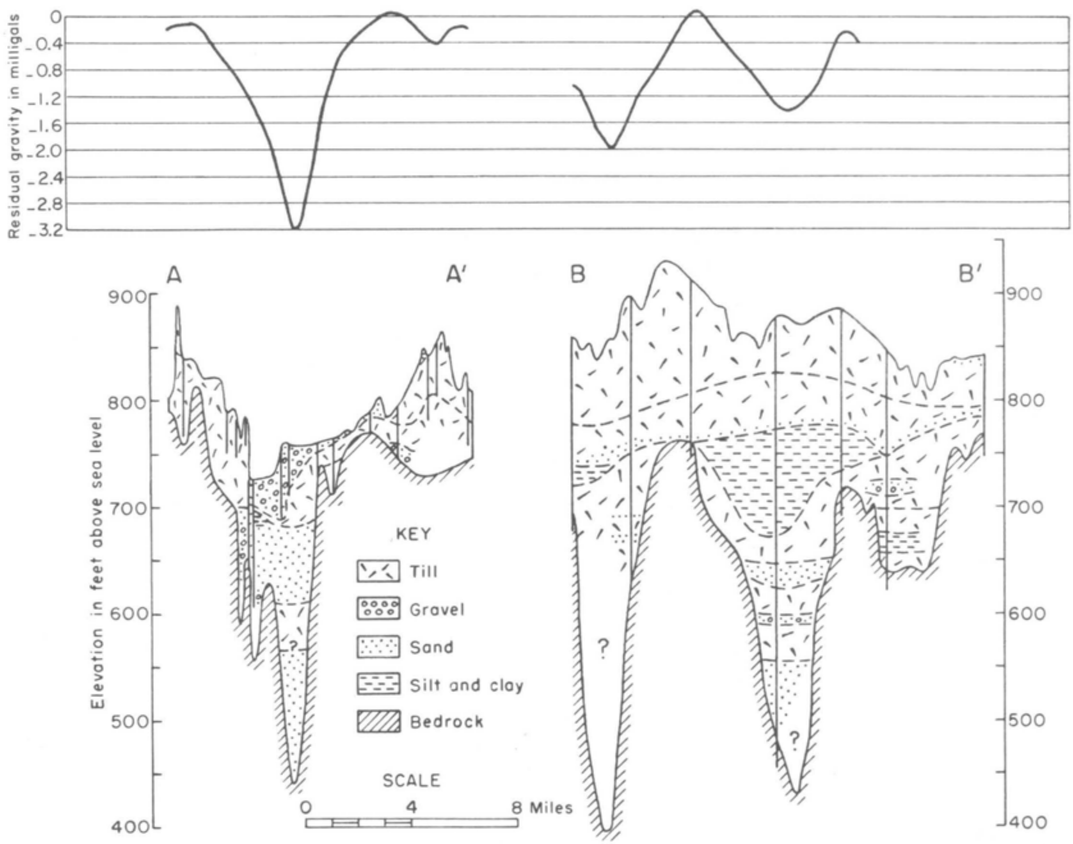

Fig. 5 - Cross section of glacial deposits and associated gravity anomalies in Troy and Rock Valleys.

DRIFT DENSITIES

Observed Densities

Densities of drift cores in the Troy Valley area were determined from approximately twenty borings, ranging in depth from 100 to over 400 feet. The cores were obtained with a split-spoon core sampler. These samples were preserved in plastic bags; thus, all of the pore water loss, if any, was due to the influence of gravity.

As a first step in the analysis of the drift densities, the selected samples, whose original environmental depths ranged from 0 to 473 feet, were examined to see if there was any direct correlation between sample density and depth. Of particular interest were those samples that contained high clay-silt and low sand-gravel percentages, since these materials would be more susceptible to compaction (a result of their often tabular shaped particles and random orientation at the time of deposition). Plots of sample density versus depth for drift with high clay-silt percentages show a completely random distribution, with no clear cut increase of density with depth. The only conclusion to be drawn is that the sample depth has no appreciable correlation with sample density.

Next, sample density versus the combined percentages of sand and gravel was plotted (fig. 6). One hundred and twenty-three of the samples tested contained a percentage of sand and gravel ranging from 0 to 68 percent. In an attempt to give meaning to the plotted data, a regression line was determined for these 123 points (fig. 6), using the standard least squares method (Brunk, 1960, p. 199-200).

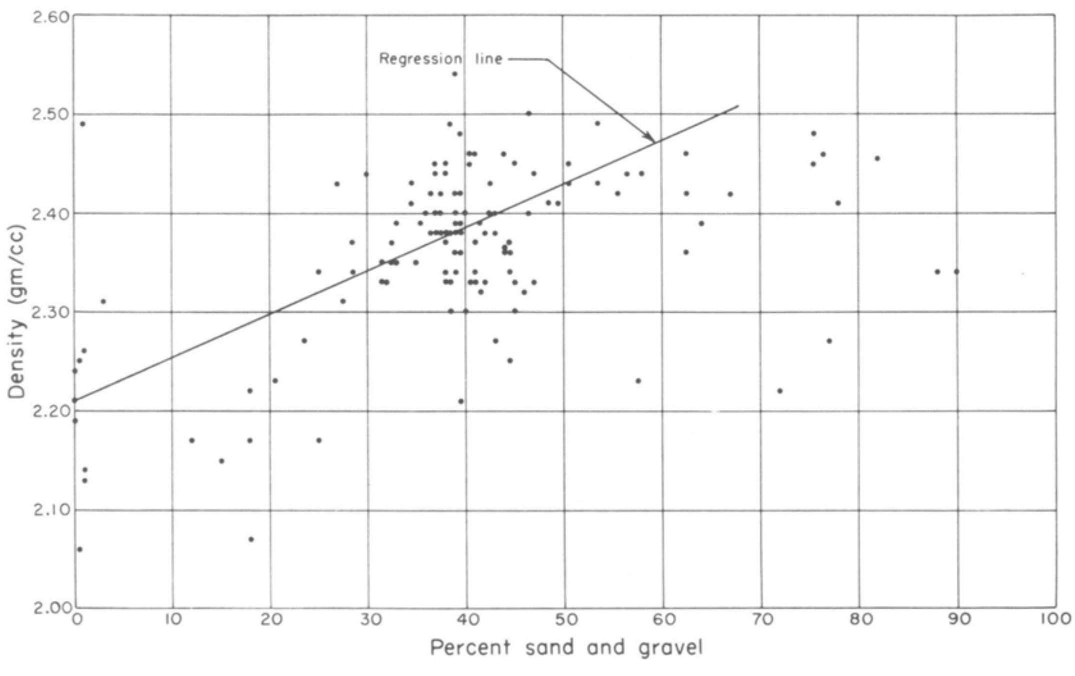

Fig. 6 - Observed drift densities versus percent of sand and gravel.

The regression line possesses an ordinate intercept of 2.21 gm/cc and a slope of .0039 gm/cc/percent, with a significant correlation coefficient of .62.

Theoretical Densities

In order to obtain an idea from a theoretical viewpoint of the drift densities as a function of percentage of sand and gravel, the systematic packing of spheres and its influence on porosity were considered. Large spheres in the following discussion are assumed to play the role of sand and gravel; small spheres are assumed to take the place of silt and clay.

In general, six cases of simple, systematic packing of spheres are recognized (Graton and Fraser, 1935, p. 785). These six cases represent various energy levels at which spheres can be packed geometrically. The highest energy level corresponds to Graton and Fraser's Case 1 (cubic packing). In this type of packing, the unit cell (the smallest portion of a body that gives complete representation of the manner of packing and distribution of the voids throughout that body) is composed of eight spheres with the two sphere rows of each square four sphere layer intersecting at 90°, and the spheres in the second layer placed vertically over those in the first layer. The lowest energy level corresponds to Graton and Fraser's Case 6 (rhombohedral packing). In this type of packing, the unit cell of eight spheres has the two sphere rows of each rhombic four sphere layer intersecting at 60°, and the spheres in this second layer are offset with respect to those of the first layer in a direction bisecting the angle between the two sets of rows and by a distance of $2R/\sqrt{3}$ where R is the radius of a sphere.

In the determination of the theoretical densities, it is assumed that the material is composed of two groups of spherical particles. The group of spheres with the larger diameters represent the gravel and sand; the group of spheres with the smaller diameters represent the silt and clay. In reality, tabular-shaped particles are a more accurate representation for the clay. However, as a first approximation in the determination of drift densities, the assumption of spherical particles must suffice.

In the lowest energy level situation, a complete unit cell composed of eight larger spheres has a porosity of 25.95 percent (Graton and Fraser, 1935, p. 805). In order to compute the maximum saturated density for the lowest energy level till, the void spaces in a complete unit cell are assumed to be filled with very small, but finite, diameter spheres, which also take on the lowest energy level. This results in a porosity of 25.95 percent in the total space allowed the smaller spheres. This is not a true physical representation of drift materials since all the empty space left by the larger spheres is not equally accessible to the smaller spheres. However, the assumption is valid for minute smaller spheres. A density of 2.67 gm/cc is assigned both sizes of particles. Grim (1953, p. 312) gives a wide range of densities for clay particles, with the most representative figure near the 2.67 value. It also is assumed that sand and gravel particles have densities near that of clays. The maximum saturated density for such a material is 2.56 gm/cc. The corresponding percentage by weight of large grained material of the solids is 79 percent.

To compute theoretical maximum saturated densities for material with less than 79 percent large spheres, the scheme is as follows. One large sphere is removed from the original unit cell at the lowest energy level, and the void is filled with the maximum amount of lowest energy level small grained material. The remnant voids are filled with water, and the density and the percentage of large spheres are computed again. Repetition of this process provides maximum saturated densities of material for nine different percentages of large spheres. The situation where all eight larger spheres are removed corresponds to zero percent sand and gravel.

The equation describing the above process is

$$\rho\max = \left\{\overline{1-P_L-\frac{(8-n)}{8}(1-P_L)} + (1-P_L)\left[1- \overline{1-P_L-\frac{(8-n)}{8}(1-P_L)}\right]\right\}\rho_S$$
$$+\left\{1-\left(\overline{1-P_L-\frac{(8-n)}{8}(1-P_L)} + (1-P_L)\left[1- \overline{1-P_L-\frac{(8-n)}{8}(1-P_L)}\right]\right)\right\}\rho_W \qquad (1)$$

where

$\rho\max$ = maximum density of saturated material

P_L = porosity associated with lowest energy level unit cell (25.95%)

ρ_S = density of solid material = 2.67 gm/cc

ρ_W = density of water = 1.00 gm/cc

n = number of large spheres in unit cell at lowest energy level.

This is a linear equation in n.

In order to fill the gaps in the curve of maximum saturated density versus percentage of large spheres between 0 and 79 percent, the reasoning is as follows. Percentage of large spheres is a linear function of the number of large spheres in the unit cells, and the maximum saturated density is also a linear function of the number of spheres in the unit cells. Thus, maximum saturated density is a linear function of the percentage of large spheres. This relationship is shown in figure 7. The curve of density versus percentage of large spheres is also shown for a dry sample.

For percentages of large spheres in the range 79 to 100 percent, the following scheme is employed. A fraction of the remnant void left by the larger spheres, making up a complete unit cell not exceeding $1-P_L$ (=.7406), is considered filled with smaller spheres (silt and clay) at lowest energy level. As this fraction decreases to zero, the combined percentage of large spheres increases from 79 to 100 percent. The corresponding maximum saturated densities can be readily computed. The equation describing this process is

$$\rho\max = \left[1-P_L + P_L a\right]\rho_S + \left\{1-\left[1-P_L + P_L a\right]\right\}\rho_W \quad a \leq 1-P_L = .7405 \qquad (2)$$

where $\rho\max$, P_L, ρ_S, ρ_W are the same as in equation (1), and a is a fraction of the remnant void left by larger spheres making up a complete lowest energy level unit cell filled by smaller spheres at lowest energy level. This is a linear equation in a.

By reasoning similar to that described for the part of the curve between 0 and 79 percent large spheres, the curve of maximum saturated density versus percentage of large spheres between 79 and 100 percent is again linear. This relationship is also shown in figure 7.

A similar curve is drawn (fig. 7) for the maximum dry density of lowest energy level material versus percentage of large spheres by eliminating the water factors in equations (1) and (2).

For the highest energy level material, the above described process is repeated. The porosity of a complete large sphere unit cell at highest energy level is 47.64 percent (Graton and Fraser, 1935, p. 805). The corresponding equations are

$$\rho\,\mathrm{max} = \left\{\overline{1-P_H - \frac{(8-n)}{8}(1-P_H)} + (1-P_H)\left[\overline{1-1-P_H - \frac{(8-n)}{8}(1-P_H)}\right]\right\}\rho_S + \left\{1-\left(\overline{1-P_H - \frac{(8-n)}{8}(1-P_H)} + (1-P_H)\left[1-\overline{1-P_H - \frac{(8-n)}{8}(1-P_H)}\right]\right)\right\}\rho_W \qquad (3)$$

where ρmax, ρ_S, ρ_W, n are the same as in equation (1), and P_H is porosity associated with highest energy level unit cell (47.64% or .4764).

$$\rho\,\mathrm{max} = \left[1-P_H + P_H\beta\right]\rho_S + \left\{1-\left[1-P_H + P_H\beta\right]\right\}\rho_W \qquad \beta \leq 1-P_H = .5236 \qquad (4)$$

where ρmax, ρ_S, ρ_W, are the same as in equation (1), P_H is the same as in equation (3), and β is fraction of remnant void left by larger spheres, making up a complete highest energy level unit cell filled by smaller spheres at highest energy level.

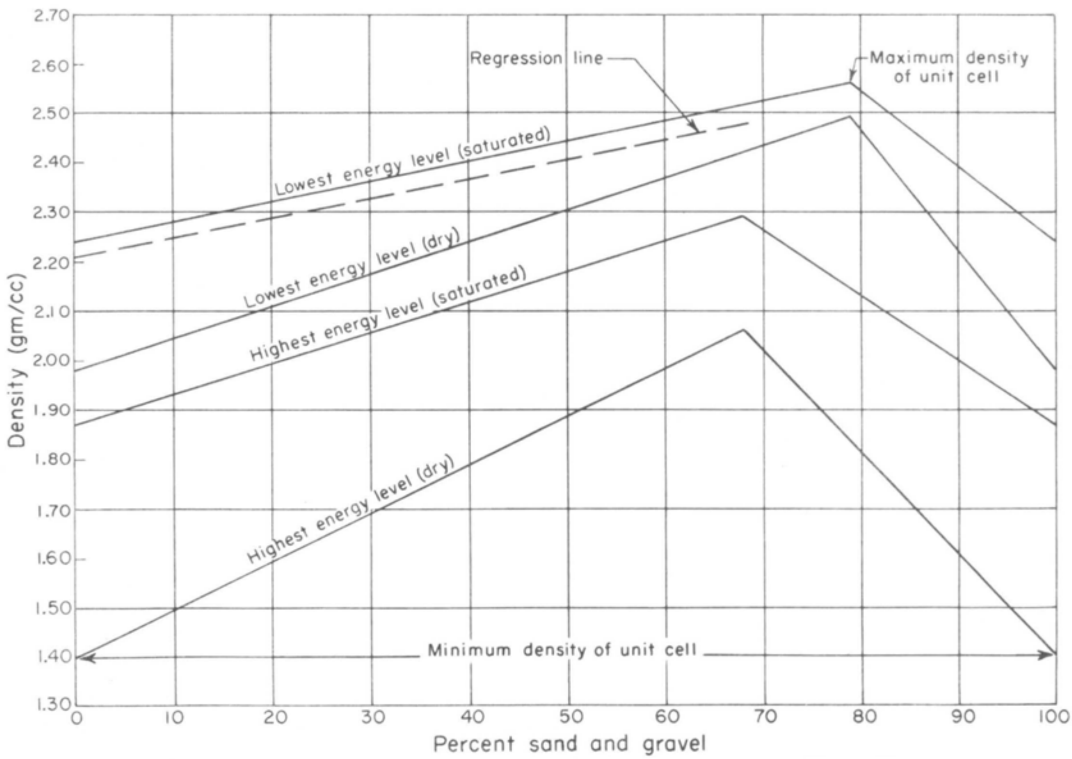

Fig. 7 - Theoretical drift densities versus percent of sand and gravel.

Both of the curves for saturated and dry sample density versus percentage of large spheres for the highest energy level materials (lowest density) are shown in figure 7.

Significance of the Regression Line With Respect to Theoretical Density Curves

As seen in figure 7, the regression line lies between the curves of theoretically saturated and dry density versus percentage of large spheres for the lowest energy situation. The slope of the regression line is slightly less than that of the curve of theoretically saturated material.

It is feasible that a till will assume the lowest energy level rather than the highest energy level, although in reality an intermediate state is probable. This, together with the fact that the samples examined in the laboratory were still highly saturated, might explain the relative position of the regression line.

Till samples with lower percentages of sand and gravel (higher percentages of silt and clay) would have a higher specific retention of water under the influence of gravity than those with higher percentages of sand and gravel. Specific retention is defined as the volume occupied by retained water divided by the gross volume of the rock (Todd, 1959, p. 23). Specific retention here is an assumption since it implies that the till sample has yielded all the water it possibly can. A more accurate term would perhaps be gravity retention (Rasmussen and Andreasen, 1959, p. 83), which includes a time factor. Since the samples considered had been allowed to drain under the influence of gravity for a period of several months, specific retention was considered adequate. This implies that till samples with lower percentages of sand and gravel tend to remain closer to the completely saturated state than the samples with higher percentages of sand and gravel. Simple calculations, involving the porosities associated with the lowest energy level situation, show that a theoretically saturated lowest energy level till with zero percent large spheres would have to yield 12 percent of its water in order to have its density coincide with the corresponding value given by the regression line. According to Tolman (1937, p. 113), a rock that yields 12 percent of its water in this way has an average grain size of approximately .02 mm diameter. This grain size would fall into the category of a fine silt, according to the laboratory criteria by which percentages of gravel, sand, silt, and clay were determined:

Gravel	=	>2.00 mm to 2.00 mm
Sand	=	2.00 mm to .062 mm
Silt	=	.062 mm to .0039 mm
Clay	=	.0039 mm to <.0039 mm

This is a consistent result.

A theoretically saturated lowest energy level sample with 68 percent large spheres would have to yield 44 percent of its water to have its density coincide with the corresponding value given by the regression line. A rock yielding 44 percent of its water in this way has an average grain size of approximately .07 mm diameter (Tolman, 1937, p. 113). This grain size would fall into the category of a fine sand, according to the above mentioned criteria. This too is a consistent result.

As previously mentioned, data determining the regression line were obtained from 123 samples with sand and gravel percentages, less than or equal to 68 percent. This particular percentage was chosen on the basis of the curve of theoretical density versus percentage of large spheres for the highest energy level case. For this energy level, the greatest densities possible (dry and saturated material) correspond to 68 percent large spheres.

While other data for samples with percentages of sand and gravel higher than 68 percent were plotted, this data is deemed too sparse for any conclusion to be drawn. Samples of this nature are difficult to handle in the laboratory. Many sample densities at low sand and gravel percent fall above the theoretically derived maximum density curve. These values may be accounted for if we remember that silt and clay particles are far from spherical. A well oriented clay, with c-axis vertical, may be arranged so that porosity is very low, which would result in a density higher than the maximum density of a cell composed of spheres.

Making use of the above information and the discussion on geology, we can see that the most prevalent type of drift material covering the Ancient Rock and Troy bedrock valley region is high density and compact glacial till. The average density of this till as determined in the laboratory is 2.36 gm/cc. In the IBM program, to be discussed later, Bouguer densities on either side of this value were 2.20 and 2.40 gm/cc; therefore, the Bouguer anomaly maps are drawn assuming a constant surficial density of 2.40 gm/cc.

GRAVITY DATA REDUCTION AND THEORY

Corrections applied to the field gravity observations include those for variations in latitude, meter drift, free air, and Bouguer. The largest element of error is involved in the free air correction since elevations, as read from the topographic maps, have a maximum error of ±1 foot. Thus, the maximum error due to elevation inaccuracies is ±0.095 milligals. The latitude correction was obtained from the 1930 International Formula. Station locations with respect to latitude are accurate to within 0.1 minute, thus errors of ±0.075 milligals may result from location inaccuracies. Minor uncorrected variations in the daily drift rate, either tidal or instrumental, may contribute a ±0.02 milligal error. Maximum relative error to be expected is ±0.19 milligals. The survey is tied to the gravity control network in North America at Janesville-Beloit, Wisconsin Station (WA175) reported by Behrendt and Woollard (1961).

Free Air and Bouguer anomalies were calculated on an IBM 650 computer at the University of Illinois. Bouguer values were obtained for each of eight different surface densities, which are 1.77, 2.00, 2.20, 2.40, 2.50, 2.67, 2.80, and 2.90 gm/cc. Bedrock densities range from about 2.59 gm/cc for Cambrian sandstones and shales to 2.82 gm/cc for Silurian and Ordovician dolomites (Krey and Lamar, 1925; Lamar, 1927). Besides the measurements made in the laboratory, 2.40 gm/cc for drift densities was also obtained in bedrock upland areas using the method described by Nettleton (1940, p. 57). In bedrock valley areas Nettleton's method was again used and drift densities were found as low as 2.02 gm/cc. Terrain corrections were not required in the survey because of low rolling topographic relief. The Bouguer maps resulting from these corrections are shown in figures 8 and 9.

Since considerable well control data is available for most of the area, it was decided to utilize this control in the construction of regional gravity maps (figs. 10 and 11). Bedrock outcrops have been mapped in some areas, but since well control was also available in these areas only well data was used. This

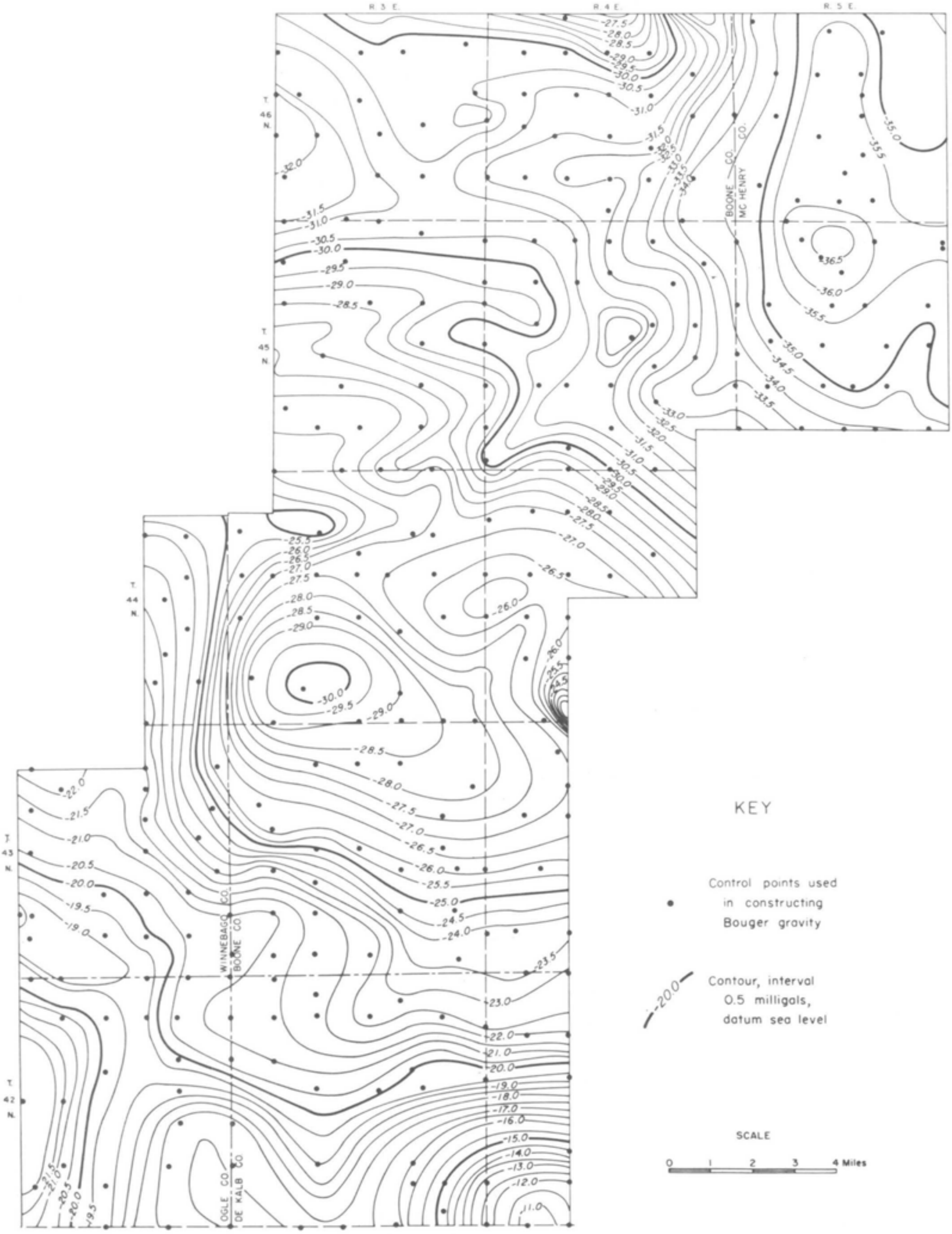

Fig. 8 - Bouguer gravity map of north half of area.

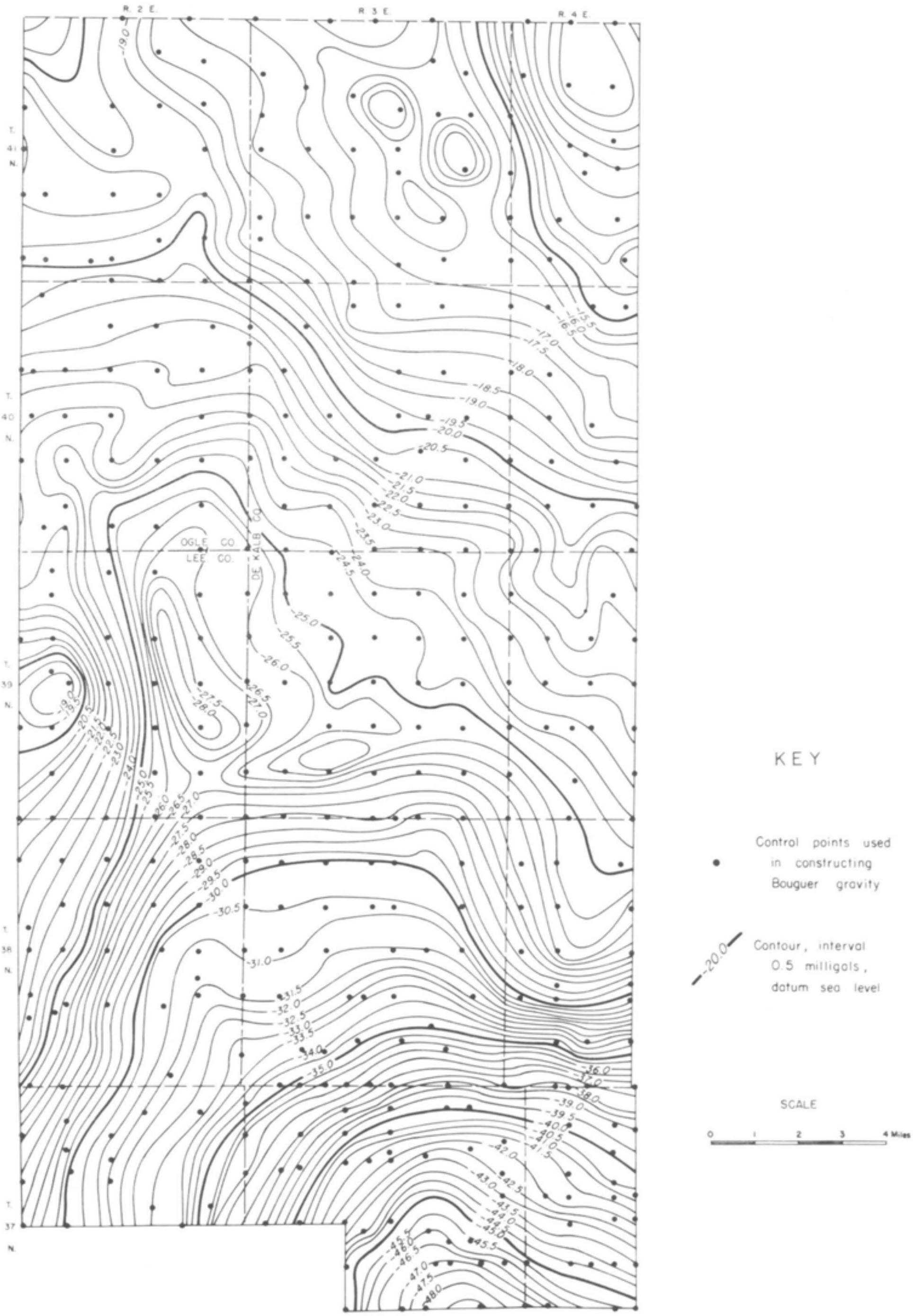

Fig. 9 - Bouguer gravity map of south half of area.

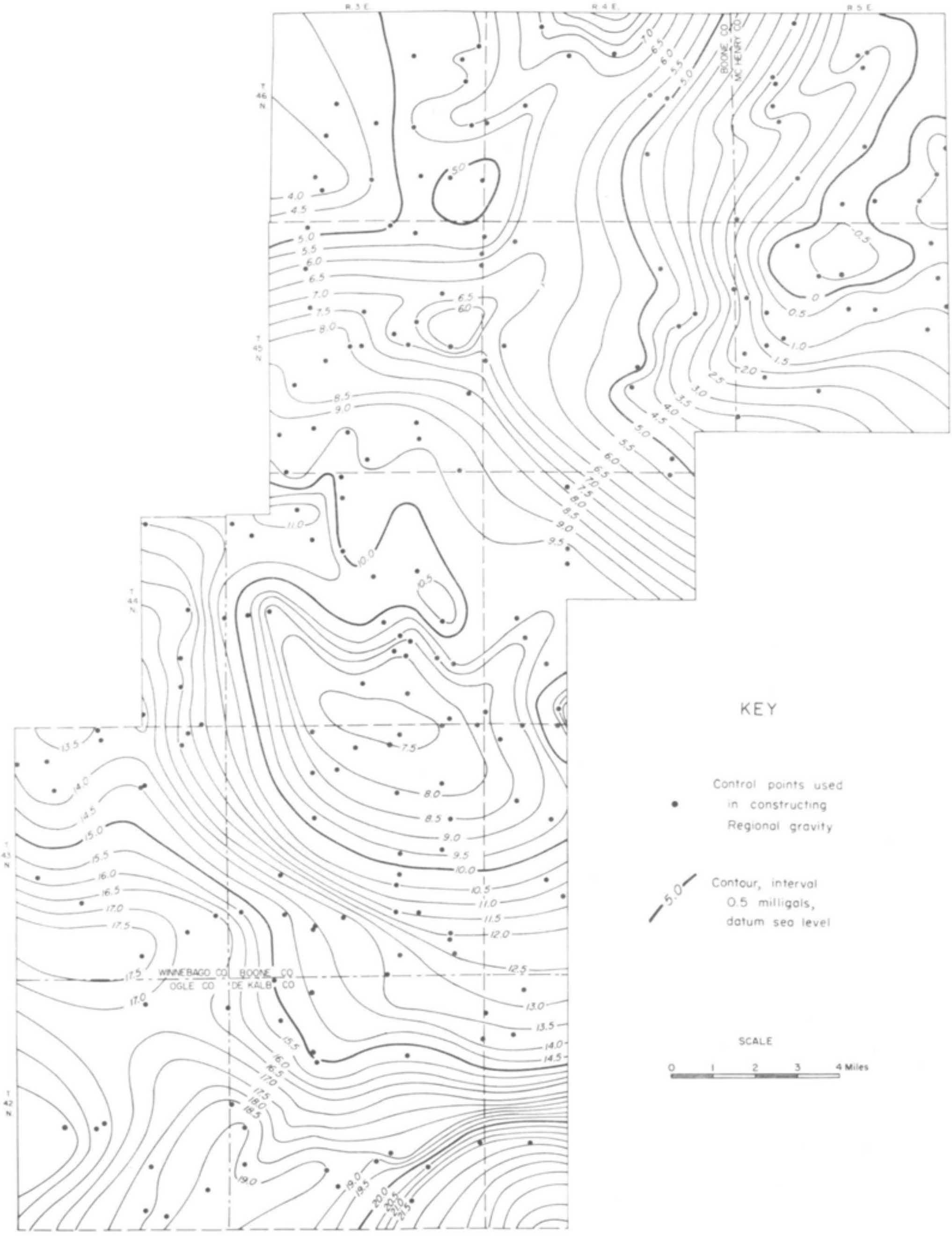

Fig. 10 - Regional gravity map of north half of area.

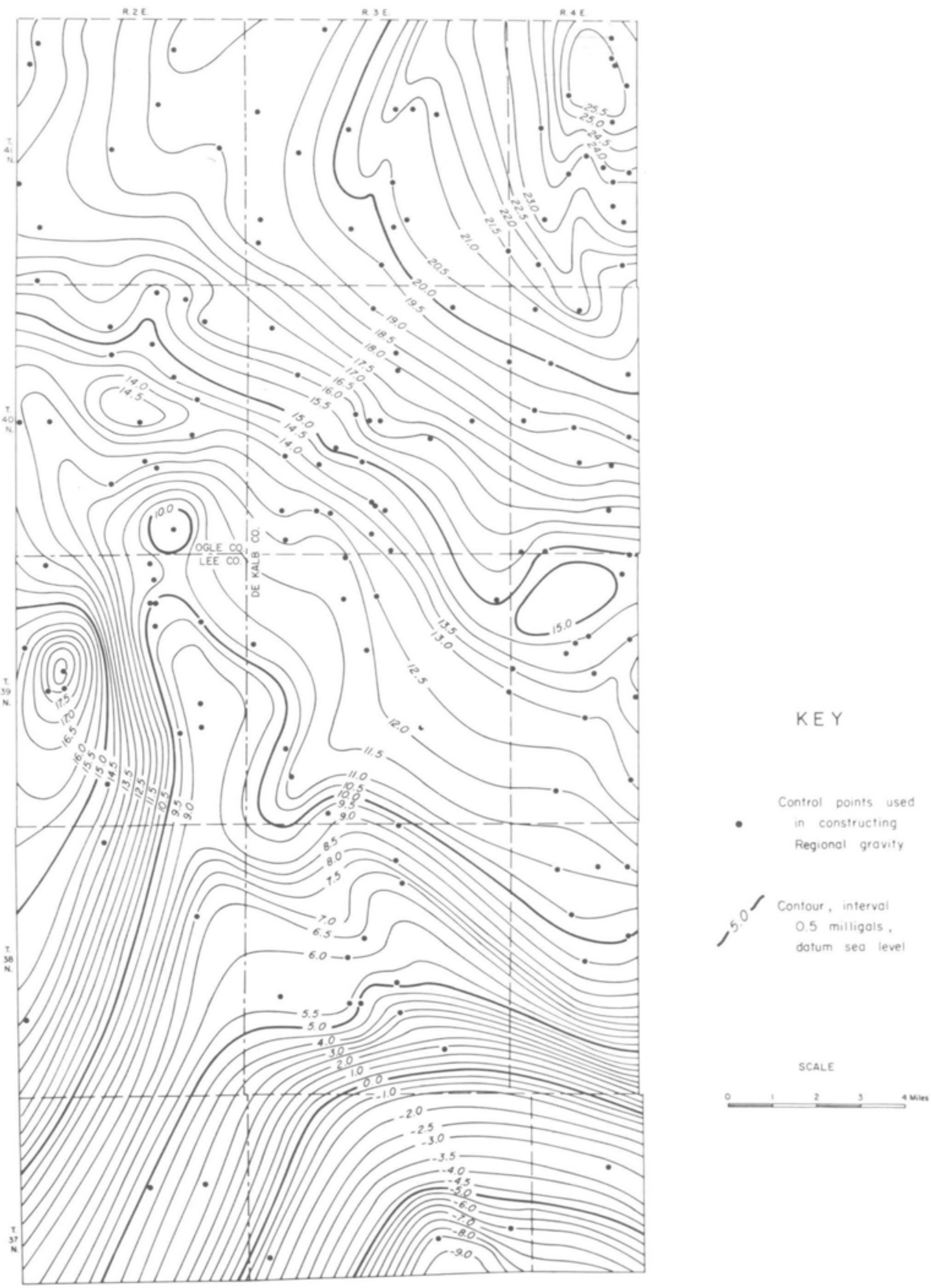

Fig. 11 - Regional gravity map of south half of area.

simplified construction of the regional. This distribution of drift and bedrock densities are so arranged that the density contrast is usually high. Most of the control wells are located on the bedrock uplands where the drift-bedrock density contrast is for the most part 0.4 gm/cc (2.80-2.40 gm/cc). According to Nettleton (1940, p. 114), the gravitational effect of a slab of rock having a thickness t, and density σ, is $g = 2\pi\gamma\sigma t$, where γ is the gravitational constant. Therefore, the thickness of material of density contrast σ, required to have a gravity effect of 1 milligal is about $80/\sigma$ feet (Nettleton, 1940, p. 115).

The procedure for determining the regional gravity value is as follows. A gravity base station is selected where the bedrock elevation is known and the Bouguer gravity is equated to elevation. For this survey the base station was chosen at a well in sec. 1, T. 46 N., R. 5 E. Bedrock elevation is 802 feet and Bouguer gravity is -36.0 milligals. At a field station (another bedrock well) elevation and gravity are again recorded. The difference between Bouguer gravity values at the base station and field station is called Δg and differences in bedrock elevation are ΔE. The theoretical gravity value at the field station due to bedrock relief is $\Delta E x$ (1/200), where 1/200 is the factor determined by assuming the density contrast to be 0.4 gm/cc. In other words, about 200 feet of bedrock relief should produce a gravity anomaly of 1 milligal. Thus, the actual bedrock relief is divided by 200 to determine the magnitude of the theoretical anomaly that should be associated with it. The value of the regional at the field station is, therefore, $\Delta g - \Delta E x$ (1/200). This method is repeated at all bedrock well locations and the regional is then contoured.

With the regional map contoured, construction of the residual is quite simple. At every gravity station on the Bouguer maps (figs. 8 and 9) Δg is known (field reading-base reading). The regional value obtained from the smoothed regional maps (figures 10 and 11) is then subtracted from the Bouguer gravity at each gravity station. The remaining value is the residual due to bedrock relief and is either a positive, negative, or zero number depending mainly on whether bedrock at the field station is above, below, or at the same elevation as the base station. A contour map of the residual values is shown on plate 1b.

REGIONAL GRAVITY AND GEOLOGIC STRUCTURE

Although the main interest of this paper is with bedrock valleys and residual anomalies, relationships between structure (fig. 4) and regional gravity (figs. 10 and 11) are apparent. The usual association of gravity highs with structural highs and gravity lows with structural lows is not observed in part of the region surveyed. North of the line of townships, T. 42 N., a direct relationship does exist between structure and the gravity field. For example, in the discussion on geology it was stated that the regional dip in the area was 13 feet per mile to the east and correlates with a regional gravity gradient of 0.5 mg/mile, decreasing to the east as determined from a recent gravity survey that extends the present survey to Lake Michigan.

South of the line of townships, T. 42 N., an inverse relationship exists. Principle examples of this inverse relationship are:

1) a gravity high centered in T. 41 N., R. 4 E., over a structural low;

2) a gravity low in T. 37 N., R. 3 E., is developed over a large structural high south of the Sandwich Fault Zone;

3) a gravity high in Tps. 38 and 39 N., R. 2 E., over a structural low.

An explanation for this reversal in relationships may be illustrated by a discussion of example 2 above. In the area of the Sandwich Fault Zone shown on the southern part of the structural map in figure 4, the gravity regional (fig. 11) is unusual. Although the displacement of the fault ranges from 550 feet to a maximum of 750 feet, with the upthrown side to the south, it is marked to the southeast by a 19 milligal gravity low. The gravity low encloses a structural high from which a minimum of 550 feet of sediment, mostly dolomite, has been removed relative to the area immediately north of the Sandwich Fault. The usual concept of denser rocks replacing those less dense on the upthrown side of a fault is not the case in this situation. With uplift on the south and removal of the dolomites, the most dense rocks in the sedimentary section are absent, placing lower density Cambrian sandstones and shales against higher density Ordovician dolomites. This situation could offset any deep effects caused by the uplift of high density Precambrian rocks to the south. However, even with this reversal of density values near the top of the sedimentary section, the anomaly is still much too great to be caused by the evident geologic conditions. A slab of rock 550 feet thick and density contrast 0.20 gm/cc would result in an anomaly of only -1.4 milligals. A possibility that will merit further consideration, after the areal extent of the anomaly has been determined by further surveying, is that part of the earth's crust south of the fault is not in isostatic equilibrium. It is interesting to note that the gravity effect of a slab of rock 550 feet thick, having a density of 2.67 gm/cc, is 18.8 milligals. If this slab were added to the upthrown side of the fault, in other words, replacing the material that has been removed, the gravity anomaly would be removed. Factors lending support to the hypothesis that complete isostatic equilibrium does not exist are (1) the existence of the gravity low; (2) the fact that the region has undergone some adjustment in the past as evidenced by the extensive bedrock faulting; and (3) earthquakes, although of rare occurrence, have been reported in the vicinity of the Sandwich Fault since 1909 (Udden, 1912).

RESIDUAL GRAVITY AND BEDROCK VALLEYS

The residual gravity values (pl. 1b) are due to variations in bedrock elevation and variations in the densities of the drift and bedrock. Residual values range from a high of +0.8 milligals (generally about +0.6) in the bedrock uplands to a low of -3.8 milligals (generally about -1.4) in the valleys. If the density contrast between drift and bedrock were constant and equal to 0.4 gm/cc, the spread of 4.6 milligals would be a function only of bedrock relief (1 milligal = 200 feet) and would indicate a total relief of 920 feet. Since maximum bedrock relief is only about 530 feet, reasons for part of the anomaly must be explained by variations in density contrast. The gravity value of the contour line that can be drawn throughout the extent of the valley is -1.4 milligals. Since the 0.0 line has been equated to 802 feet, the -1.4 value represents an elevation of 522 feet (800 feet + residual anomaly value x200) and compares favorably with the minimum elevation of 550 feet in the northern portion of the area. The elevation corresponding to the +0.6 value is 922 feet and also agrees well with the upland elevations.

As is the case in many glaciated areas, the greatest thicknesses of sand and gravel are found in the bedrock valleys. Figure 5 shows two cross-sectional views of the sequence of materials in the Troy Valley and their associated residual anomalies. A comparison of the cross sections of bedrock topography and glacial deposits in the two profiles mentioned above, with their residual gravity profiles, indicates that the anomalies are largest where the thickness of sand and gravel is greatest. Knowing the true bedrock relief, the bedrock density, and the associated residual anomaly, the drift density can be derived. Using these values and assuming a bedrock density of 2.80 gm/cc, the density of the drift in the cross section A-A' (fig. 5) is 2.02 gm/cc and contains about 200 feet of sand and gravel; whereas the densities in the Ancient Rock and Troy Valleys in cross section B-B' are 2.34 and 2.45 gm/cc respectively. The sand and gravel thickness in the Ancient Rock Valley (B-B') is unknown, however, that in the Troy Valley is roughly 100 feet. In some instances the lateral displacement of minima may be caused by either lack of adequate well control, gravity stations, or both. It also must be noted that extreme lows on the residual map (pl. 1b) and a low on the regional map, (fig. 11) in T. 39 N., R. 2 E., cutting northwest across the Sandwich Fault and joining the 19 milligal low mentioned previously, are associated also with areas in which the high density Galena-Platteville Dolomite has been cut completely out by the entrenched Ancient Rock and Troy Valleys.

Where the density contrast differs from 0.4 gm/cc, errors in depth estimation, using the residual map, will result. This density contrast was chosen so that variations from it will usually be greater then 0.4 gm/cc and will emphasize anomalies in the bedrock valleys, since it is here that the density contrast is probably large. Thus, large and elongate gravity lows on the residual map not only represent bedrock valleys but also areas of high density contrast. The portion of the residual anomaly not controlled by bedrock topography is very minor in the uplands, since here bedrock and drift densities are more or less constant.

The combined effect of diminishing well control and the distortion of the gravity field near the Sandwich Fault Zone produce variations in the residual map that may not agree with geologic conditions in the southern part of the survey. The bedrock density and therefore the density contrast may also diminish to the south across the fault where Cambrian sandstones and shales lie beneath the drift materials. More precise surveying would be required in this area to provide a more realistic bedrock picture.

CONCLUSIONS

Geophysical mapping of bedrock valleys filled with glacial deposits may be done in a number of ways. The seismic refraction method is the most exact (McGinnis and Kempton, 1961), but it is also the most expensive and time consuming. Resistivity exploration has also been used (McGinnis and Kempton, 1961), but these results are usually ambiguous. The gravity survey has proven useful, especially where some bedrock well control is available, in outlining the trends of a large (Troy) bedrock valley. The method is extremely rapid and relatively inexpensive when elevation control, accurate to within ± one foot, is available.

Knowledge of the density of bedrock formations is required, and this density must differ from drift densities. In the Troy Bedrock Valley area, bedrock consists mainly of high density (2.80 gm/cc) dolomite of Silurian and Ordovician age. In the southern portion of the area, Cambrian sandstones and shales of lower density (2.60) cause the anomalies to approach the limits of error. Thus, valley trends cannot be assured in areas of low bedrock density.

Drift densities may vary over a broad range (2.06 gm/cc to 2.54 gm/cc). Uniform particle size is the main criteria for low density till, while tills having mixed grain sizes, with the larger grain size comprising 79 percent of the till sample, form the highest density material.

Removal of the gravitational effects caused by bedrock valleys enables one to obtain a clearer look at the effects of geologic structure on the gravitational field. The gravity regional is directly related to geologic structure in the northern half of the area, while in the southern half gravity highs are related to structural lows. Reasons for this inversion are not entirely clear but are probably mainly due to the thinning of high density dolomites over the highs.

The residual gravitational field is dominated by long narrow lows, coinciding with the major deeps of the Ancient Rock and Troy Bedrock Valleys. Numerous minor tributaries are not evident on the residual map (pl. 1b) although some of the larger tributaries are indicated by gravity lows. A direct relation between the residual values and bedrock elevations is not evident owing to variations in drift and bedrock densities.

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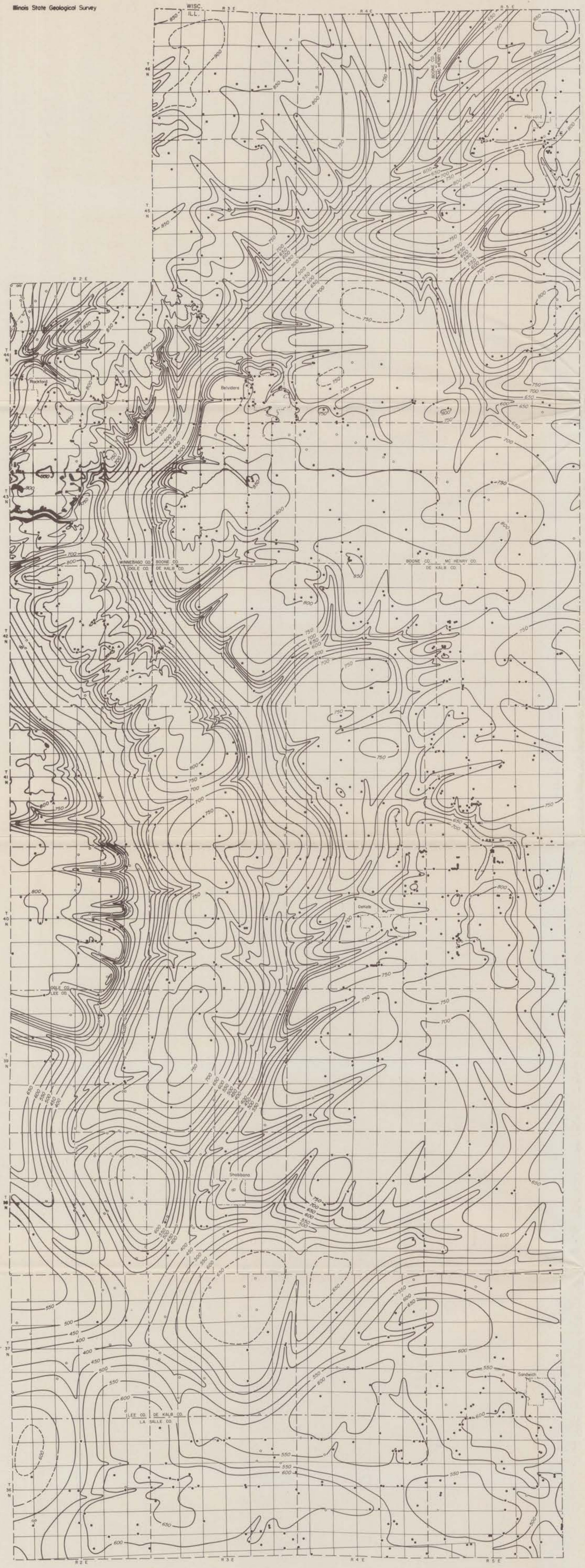
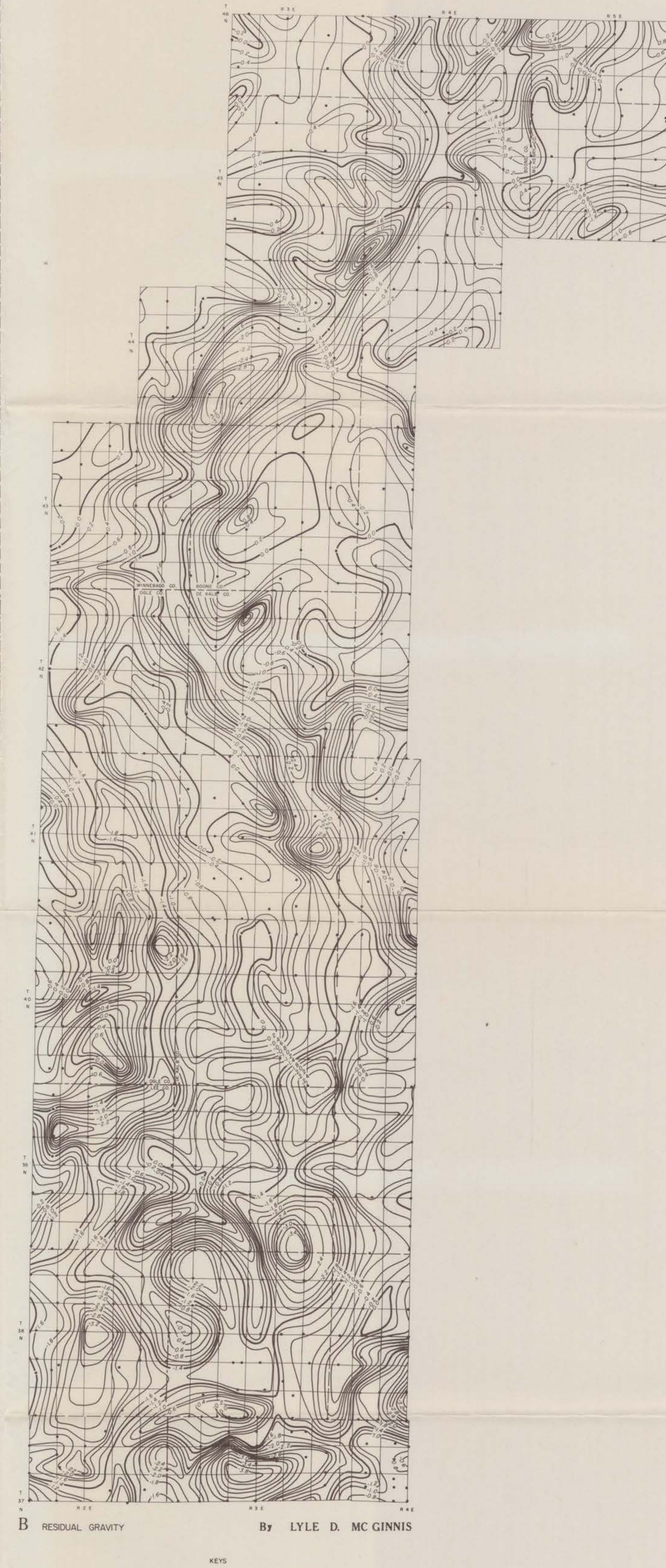


A BEDROCK SURFACE TOPOGRAPHY

By JOHN P. KEMPTON



B RESIDUAL GRAVITY

By LYLE D. MC GINNIS

KEYS

A

Well Control

- • Well or boring penetrating bedrock
- ○ Well or boring completed above bedrock
- 700 Contour, interval 50 feet; datum, sea level
- ▲ Bedrock exposure

B

- • Gravity station
- 1.4 Contour, interval 0.2 milligals

SCALE

0 1 2 3 4 Miles

RELATIONSHIP OF GRAVITY ANOMALIES TO A DRIFT-FILLED BEDROCK VALLEY SYSTEM IN NORTHERN ILLINOIS

1962