Continental Rifting Under the Illinois Basin

and Mississippi River Embayment

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Abstract

A hypothesis is set forth herein that the Illinois basin and the Mississippi River embayment are troughs along a rift in the continental crust. The two troughs are separated by a small uplifted area in which erosion has exposed the underlying structure, that of the Illinois-Kentucky fluorspar district. In the district, the expression of the rift is a belt of narrow grabens separated by wider horsts, both in Carboniferous rocks. An analysis of the fracture mechanics shows that the faults bordering the grabens resulted from a decrease in the horizontal confining stress σ_3, which in turn resulted from extension of the continental crust. The extension is ascribed to a belt of dikes injected at depth from the mantle into the continental crust. Extension of the crust above the belt of dikes requires that the vertical dimension of the rocks decreases as the horizontal dimension increases in order to maintain constant volume. This lowers the surface, forming a trough into which sediments accumulate.

Affinities with the mantle are a large amount of fluorine, high temperature of deposition of the mineral deposits, sulfur isotope ratios nearly that of the meteoritic standard, and small dikes of peridotite.

Rifting was active throughout the carboniferous and is consistent with some features of stratigraphy and sedimentation in the Illinois basin. Many useful correlations of structure and sedimentation probably can be made. Some may have immediate economic value.

Introduction

In this paper the proposal is set forth that the Mississippi River embayment and the Illinois basin developed over a rift in the continental crust. The type of rifting proposed is caused by material from the mantle injected as a belt of dikes into the lower part of the crust. Addition of the dikes to the lower part of the crust caused the crust to extend in a horizontal direction normal to the strike of the dikes. The upper part of the crust reacted by faulting and sinking so that sediments collected in the trough. The dikes are not exposed; instead their presence is deduced from requirements set forth in an analysis of the mode of formation of an unusual group of grabens in the Illinois-Kentucky fluorspar district. An analysis of the faults and grabens was made in order to understand the fractures, small veins, and breccias under the bedded deposits of fluorspar in the Cave in Rock mining district. This writer, in studying the process of hydrothermal mineralization in the Cave in Rock district, has accumulated a body of geochemical, isotopic, thermal, and mineralogical data, all of which seems to be related to the structure under the deposits. The fractures bordering the grabens are similar to the fractures under the ore deposits. Hence an analysis of the mechanics of fracturing was undertaken. This analysis led to the requirement of extension of the crust. Extension of the crust leads to sinking of the surface and the consequent effect on sedimentation in continental basins.

Geologic features

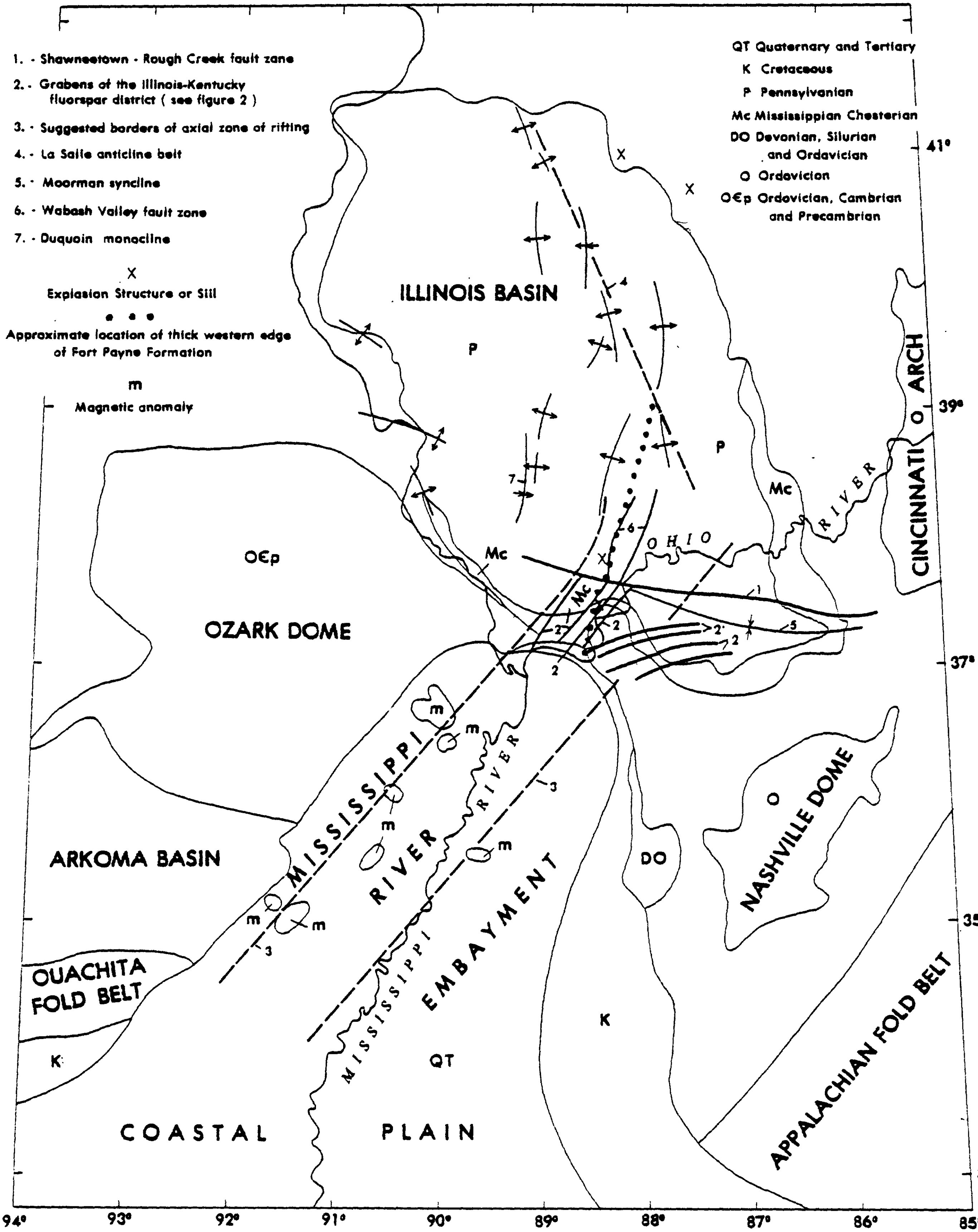
The Mississippi River embayment and the Illinois basin are separated by an area underlain by Carboniferous sediments which are greatly faulted (fig. 1). This area is the site of the

Figure 1.--NEAR HERE

Illinois-Kentucky fluorspar district. The fault structure of the district is considered by this writer to be indicative of structure underlying both the embayment and the Illinois basin, and is herein referred to as the Illinois-Kentucky rift zone.

The embayment extends northward from the Gulf coastal plain to southern Illinois, a distance of about 1,100 km, and widens southward. It contains unconsolidated sediments of Late Cretaceous to Tertiary age lying upon unidentified older sediments. Along the eastern, northern, and northwestern edges of the embayment, Paleozoic rocks dip gently into the embayment. Along the southwestern edge, folds of Pennsylvanian age of the Ouachita Mountains plunge under the embayment. Basic igneous intrusive bodies at shallow depth in the embayment are indicated by a row of magnetic anomalies (Hildenbrand and others, 1978).

Figure 1.—Generalized geologic index map showing structural features of the Illinois basin, Mississippi River embayment, and nearby areas. The index map is modified from King and Biekman (1974); sources of other data are referred to in the text.



1. - Shawneetown - Rough Creek fault zone
2. - Grabens of the Illinois-Kentucky fluorspar district (see figure 2)
3. - Suggested borders of axial zone of rifting
4. - La Salle anticline belt
5. - Moorman syncline
6. - Wabash Valley fault zone
7. - Duquoin monocline
X Explosion Structure or Sill
Approximate location of thick western edge of Fort Payne Formation
m Magnetic anomaly
QT Quaternary and Tertiary
K Cretaceous
P Pennsylvanian
Mc Mississippian Chesterian
DO Devonian, Silurian and Ordovician
O Ordovician
O€p Ordovician, Cambrian and Precambrian
ILLINOIS BASIN
OZARK DOME
ARKOMA BASIN
OUACHITA FOLD BELT
MISSISSIPPI RIVER EMBAYMENT
NASHVILLE DOME
CINCINNATI ARCH
APPALACHIAN FOLD BELT
COASTAL PLAIN
OHIO RIVER
MISSISSIPPI RIVER
41°
39°
37°
35°
33°
94°
93°
92°
91°
90°
89°
88°
87°
86°
85°

Fig. 1

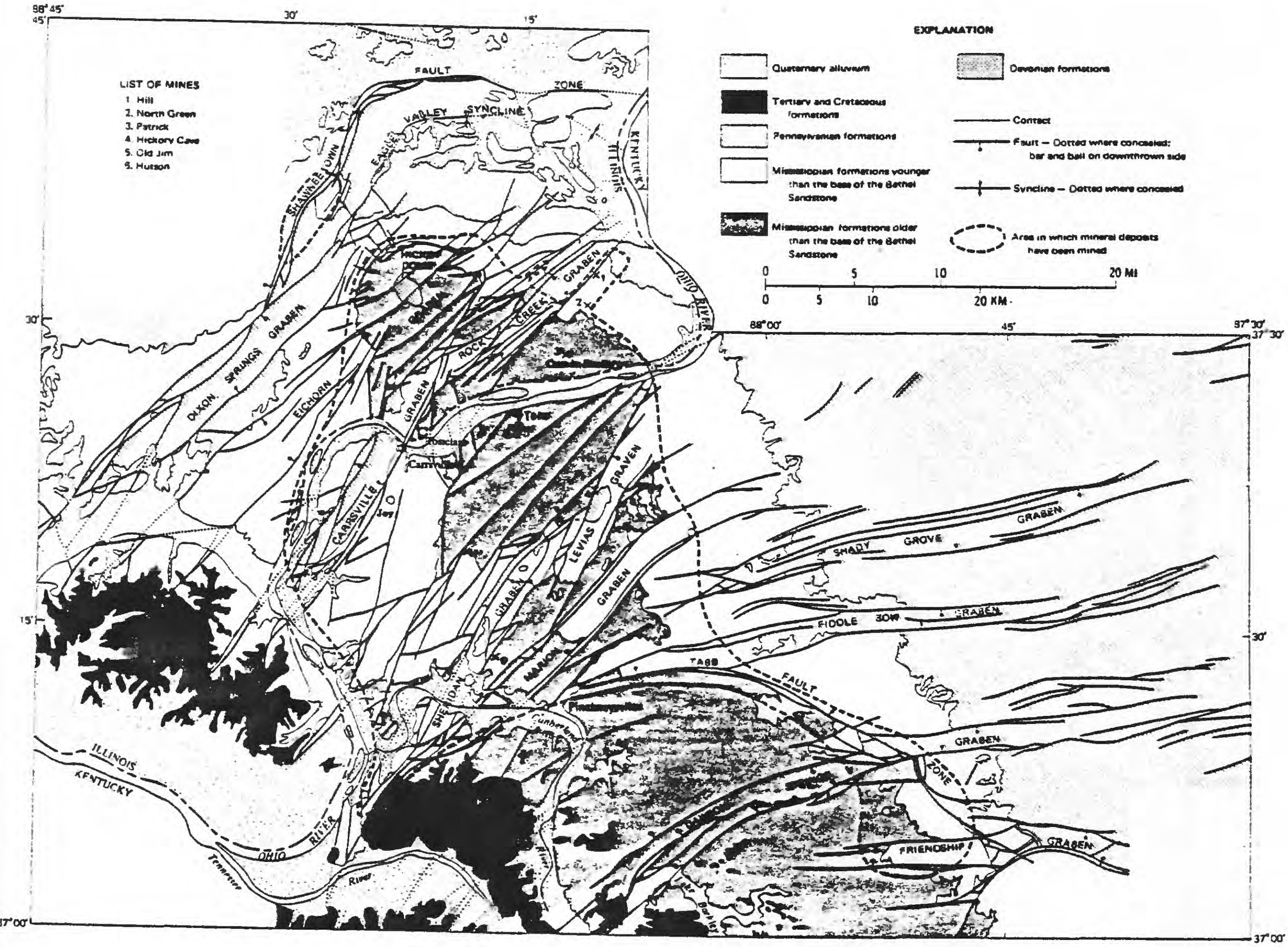
The Illinois basin extends northerly from western Kentucky about 600 km and is about 280 km wide. Its deepest part is near its southern end, near the fluorspar district. The basin lasted at least from Late Devonian to Permian time. Rocks from Devonian-Missippian age (Chattanooga shale) through Pennsylvanian age all thicken to the south in the basin. In one place in western Kentucky, Permian fossils were reported (Rice and others, 1979) and there the rocks are about 1,000 m thick. Elsewhere the younger rocks of the basin were eroded. The long axis of the basin is in the eastern part and is marked by a row of small anticlines bordered by narrow troughs, the La Salle anticlinal belt. All rocks bordering the basin are older than Pennsylvanian and dip very gently into the basin.

The Illinois-Kentucky fluorspar district (fig. 2) is in an area chiefly of Mississippian sediments which are exposed for about 50 km between the younger rocks of the embayment and of the Illinois basin. The structure, stratigraphy, and geologic setting of the district were summarized by Pinckney (1976). The district contains two major structural features: (1) a group of three roughly alined domes near the towns of Pinckneyville, Tolu, and Hicks that trend northwesterly, forming an arch; and (2) a group of grabens that trend northeasterly or easterly across the arch. Hicks dome contains explosion breccias (Brown, Emery, and Meyer, 1954), some of which are extensively mineralized, containing much fluorite. The district and surrounding region contain many peridotite dikes and diatremes (Clegg and Bradbury, 1956; Weller, 1921). The dikes strike northwesterly, approximately along the crest of the arch.

Figure 2.--NEAR HERE

The main sequence of events is approximately as follows: (1) formation of grabens, followed by (2) raise of the arch causing arching of the Carrsville-Rock Creek graben-complex, with associated explosion of Hicks dome, and intrusion of small dikes and diatremes into Mississippian and Pennsylvanian rocks, and finally (3) deposition of ore. The small dikes were dated as Early Permian (Zartman and others, 1967). Faulting related to the grabens may have started as early as Late Devonian or very early Mississippian time and continued into Permian time. The tectonics of earlier stages of the basin is not discussed in this paper.

Figure 2.—Geologic map of the Illinois-Kentucky fluorspar region, modified from Pinckney (1976).



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ig. 2

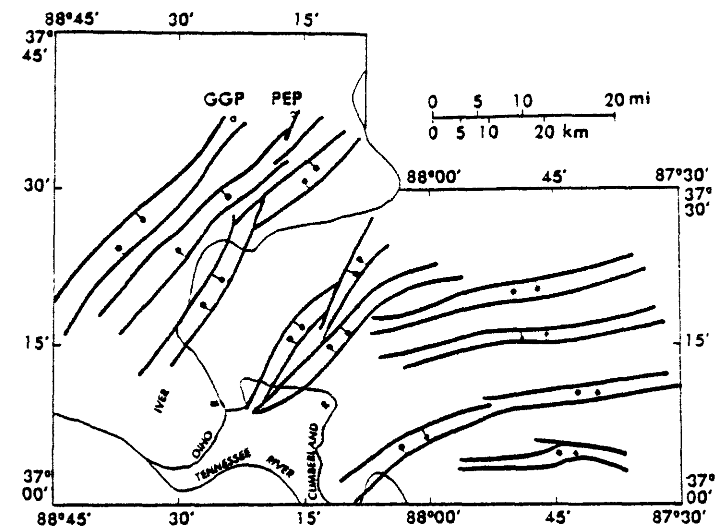
Bottom

Eleven of the grabens are shown on figure 2, and the belt of grabens is known to extend farther to the southeast beyond the map area and possibly to the northwest. The grabens are narrow, only about 3 to 4 km wide. They are bordered by steeply dipping faults and nontilted horsts. Vertical displacements commonly are from 200 to 400 m. The untilted horsts appear to be structurally undisturbed except for faults of small displacement. The grabens contain many faults that dip about 45 degrees toward the bordering faults. A simplified pattern of the grabens is shown in figure 3 which shows only the bordering fault systems. The bordering faults are normal faults, most of which dip toward the grabens, but a few dip toward the horsts and are therefore reverse faults. Nevertheless, the relative movement is graben-side down.

Figure 3.--NEAR HERE

Figure 3.--Pattern of faults bordering narrow grabens. GGP and PEP are measured sections of Morrow (1979). These sections of the Lower Pennsylvanian Pounds Sandstone Member of the Caseyville Formation are in or adjacent to grabens and are thicker than Pounds sandstone in the intervening horst. Ball and bar indicate down-dropped side of faults.

Top



88°45'
30'
15'
37° 45'
GGP
PEP
0 5 10 20 mi
0 5 10 20 km
88°00'
45'
87°30'
30'
37° 30'
15'
IVER
OHIO
TENNESSEE
RIVER
CUMBERLAND
37° 00'

Fig. 3

The problem

The number of grabens and their displacements relative to the horsts are both of sufficient magnitude so that at least four things must be accounted for. These things are (1) the volume of the rocks in each fault block, (2) the dip of the faults, (3) the dip of the beds in the horst blocks, and (4) the elevation of the surface of the earth relative to sea level. In order to conserve mass, the volumes of the fault blocks must be the same before and after faulting. No reason existed to significantly change the density of the rocks; hence volume was constant.

From theoretical and experimental analysis, Hubbert (1951) showed that normal faults should dip about 60 to 65 degrees. Many normal faults do have such moderate dips. However, if the faults of the district dip generally 60 to 65 degrees toward the grabens, a problem arises as to how the narrow grabens sank relative to their bordering horsts, as shown in figure 4. If the faults dip very steeply, or are vertical, they can extend to great depths, even to a zone of compensation in the mantle. However, theoretical problems are encountered in accounting for both the steep dips and the sinking of the grabens (see fig. 4). The bordering rocks of many of the large grabens in continents are tilted back from the graben. The beds in the horsts in the district are level or nearly so. These problems can be more clearly understood by examining a well-known rift zone to determine the essential properties of rifting and then finding mechanics that produce the combination of features found in the Illinois-Kentucky rift zone.

Figure 4.--NEAR HERE

Figure 4.--Fault-block problem. In (<u>A</u>) faults bordering block S dip 60 degrees in accordance with the "theoretical" dip of about 60-65 degrees. Block S extends to a depth of only 2.6 km. It cannot sink to become a graben by movements on faults a or b as long as the walls of a and b are in contact. If the walls of a or b did move apart, the width of the shallow grabens so formed would vary with depth of exposure. This is inconsistent with the slight variation of width of the grabens in figure 3.

In (<u>B</u>) the moderately dipping faults of (<u>A</u>) are shown to scale in a crust 30 km thick for perspective. Vertical faults c and d can extend to great depth, entirely through the crust, or deeper, separating a narrow block, Q, of uniform width from the rest of the crust. However, block Q will not "sink" due to faulting alone and become a graben unless its mass is increased or the masses of adjacent blocks are decreased. No reason has been found for such differential changes of mass.

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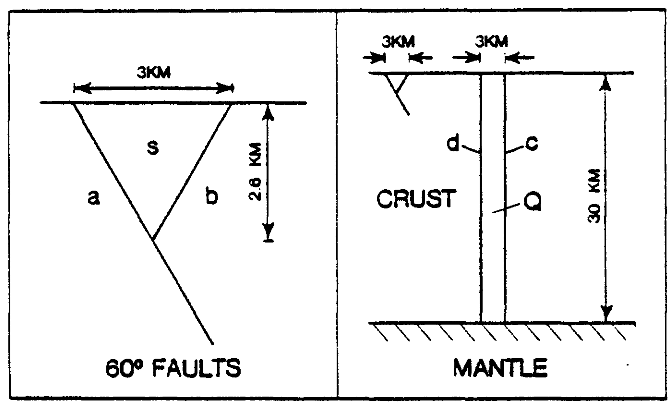

A B

Fig 4

Rhine graben system

The Rhine graben in Europe (illustrated in fig. 5) is the most thoroughly studied of all of the continental rifts of the Earth. Concise descriptions of it are given by Illies (1969) and by Mueller (1969). The Rhine graben is 300 km long and 38 km wide. The bordering rocks are tilted back 1-3 degrees. Subsidence started in middle Eocene time and is continuing today at the rate of 0.5 mm per year. Vertical displacement is 4.4 km. Faults along the borders dip between 55 and 80 degrees, but drilling shows that the predominant dip of the master faults is between 60 and 65 degrees. The graben block is broken by many antithetic faults, and when reconstructed to its original width, it is 4.8 km too narrow to fill the space between the bordering blocks. The horizontal extension therefore is about 15 percent of the original width of the top of the graben. The bordering faults, projected downward, intersect at a depth of 30.6 km, which is the depth of the bottom of the continental crust in central Europe away from the rift. Seismic data indicate the presence of a broad and low lens, or pillow, of light mantle material at the base of the crust, presumably intruded material, centered beneath the graben. Heat flow in the graben is high. The graben contains basic-alkalic igneous rocks. Illies (1969) suggested that the horizontal extension is due to the crust sliding off the top of the lens.

The Illinois-Kentucky grabens differ from the Rhine graben in three essential features. They are numerous and narrow within a wide belt rather than being a single wide graben, their bordering faults dip steeply (fig. 9) rather than moderately, and neither the horsts nor the borders of the belt are tilted back. For both types of graben, however, horizontal extension of the crust is required to make the bordering fractures and to cause widening of the trough.

Figure 5.—NEAR HERE

The main differences between the Rhine rift zone and the Illinois-Kentucky fault zone can be accounted for by the shape of the grabens. The Rhine graben sank partly because, as the sides moved apart, a greater part of the mass of the graben was supported at the bottom of the crust by the small area near the intersection of bordering faults. The vertically-directed downward stress from the graben therefore exceeded the vertically-directed upward stress of the mantle, upsetting the stress equilibrium that existed, and the graben sank. Similarly, under the borders the loss of mass from the top of the crust, caused by sinking of the graben, resulted in a decrease of vertically directed downward stress under the borders so that upward stress of the mantle caused the borders to rise. Also, as the borders moved apart, the graben widened by breaking up. In order to maintain constant volume it shortened vertically and the top sank more than the bottom. The borders also probably rose generally over a broad area as the pillow thickened, adding to the isostatic rise. Figure 6 shows that a similar vertical shortening of an Illinois-Kentucky graben would occur, in order to maintain constant volume, if the bordering rocks moved apart. Movement of the bordering rocks in a horizontally outward direction must be proved. If proved, the fault dips then give the shape of the grabens and the nature of the borders. A proof is given in the appendix by analyzing the mechanics of the fracture systems. This analysis shows that the confining stress σ_3 decreased during faulting.

Figure 5.--Mechanics of a wide graben of which the Rhine graben is an example. _A_ and _B_ are from Illies (1969, fig. 5). In _A_, the crust is slightly arched by intrusion of material (2) from the mantle (1); fractures dipping about 60 degrees cut through the lower crust (3) and upper crust (4); subsidence started in middle Eocene time. In _B_, representing Holocene time, the crust is well arched, the sedimentary cover (8) is stripped from the crystalline basement in the Black Forest (6) and Vosges (7), the large graben is breaking up, and sediments (5) are collecting in the wide trough.

Figure 5 _C_, _D_, and _E_ show the stresses in a simple manner; the arrows indicate the direction of the active stress. In _C_ the crust was stretched beyond its elastic limit by the intrusion of (1) forming the border fractures (9). In _D_ the crustal segments moved apart allowing the fracture-bounded block (10) to sink into the mantle because of increased pressure at its lower tip. Because of loss of mass (12) from the crustal segments, the overburden-load of the crustal segment was reduced, but the pressure in the "plastic" mantle in _E_ was not. The ends (11) of the crustal segments rose and became eroded, forming the bordering highlands (13) of (6) and (7) in _B_.

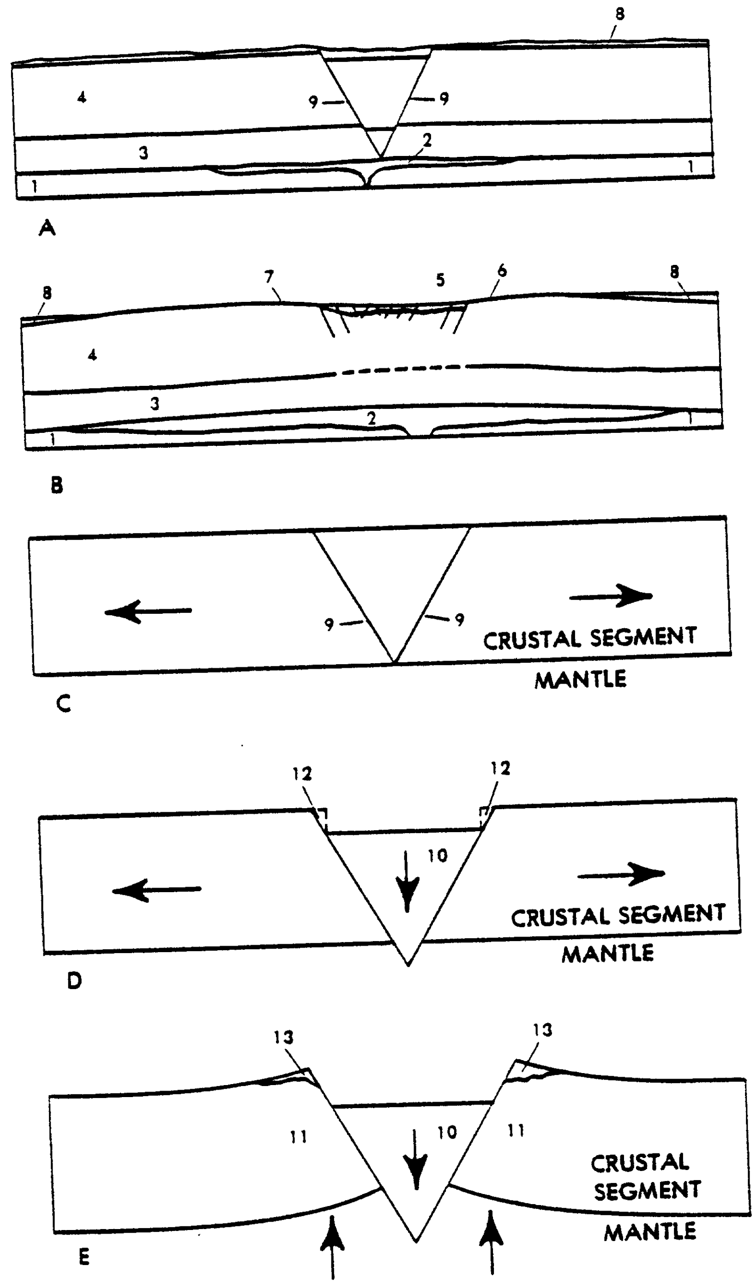
8
4
9
9
3
2
1
1
A
8
7
5
6
8
4
3
2
1
1
B
9
9
CRUSTAL SEGMENT
MANTLE
C
12
12
10
CRUSTAL SEGMENT
MANTLE
D
13
13
11
10
11
CRUSTAL
SEGMENT
MANTLE
E

Fig 5

Horizontal extension

A decrease in $\bar{\sigma}_3$ means extension, that is lengthening, of the Earth's crust in the direction of the $\bar{\sigma}_3$ axis. This is required by Hooke's law. The amount of extension required to fracture all of the rocks is very small. However, the process that caused the extension was not necessarily limited by the small amount associated with fracturing. The decrease in σ_3 shows the direction of movement but not necessarily the maximum amount that occurred.

The mechanism of extension proposed herein is the injection of dikes of material from the mantle into the crust. If the dikes were injected along a belt, as dikes high in the crust are frequently found to be, they would cause considerable extension of the crust. The direction of extension would be perpendicular to the strike of the dikes. Figure 6 shows, schematically, what can be expected from the intrusion of dikes at depth. In figure 6*A*, a dike or a group of dikes totaling 150 meters in width are injected upward to 5 km below the surface. The resultant "opening" of the rocks above the dikes is 150 meters, and vertical fractures extend upward to the surface. The narrow prism of rock separated from the adjacent blocks is unstable and it breaks up along antithetic faults (fig. 6*B*). Movement on the antithetic faults results in widening of the narrow prism to fill the opening above the dikes. Hence the prism changes dimensions. Because the product of the three dimensions of the prism must be the same before and after widening in order to maintain constant mass and volume, the increase in the width requires a decrease in the height of the prism (fig. 6*B*). The decrease in height means that the widened part of the prism above the dikes sinks and becomes a graben.

No reason exists for the blocks bordering the narrow prism of figure 6A to remain stable. They will break up in a manner similar to the grabens, and will tend to become shorter columns of rock. Hence, both the grabens and the horsts will sink relative to sea level, but the grabens will sink more. Comparison of figure 2 with figure 3 shows that the complex fault pattern is largely due to faults in the horsts.

Figure 6.--NEAR HERE

Figure 6.—Effects of extension of the crust. Figures _A_ and _B_ observe the rule of constant volume and are to scale; _C_, _D_, and _E_ are schematic. In _A_ through _E_ the dikes were injected to a level of 5 km below the surface of the original crust. (1) vertical fracture, (2) open space, (3) vertical bordering fault, (4) trough above graben may contain thicker sediments and coarser clastics, (5) trough may contain conglomerate, (6) older sediments may be restricted to central part of the basin, (7) younger sediments extend to sides of basin, (8) youngest sediments may fill only interior part of basin, (9) top of dikes, (10) dikes below (9), (11) fracturing in sediments above active grabens.

(_A_) A dike 150 m wide is injected causing vertical fractures such as (1) and open space (2).

(_B_) The unstable graben changes shape but not volume to fill the open space. It does this by collapsing, the blocks moving downward and outward along antithetic faults dipping about 45 degrees. The graben shown as 3 km wide shortens 238 m and widens 150 m. The surface of the potential graben sinks and the rocks above the level of the dikes become a graben.

(_C_) A narrow horst is unstable, as are the grabens, and will widen, causing it to break apart and shorten, thereby sinking at the surface. The original surface is indicated by the dashed line. An example is the horst between the Dixon Springs and Eichorn grabens which are shown on figure 2.

(D) A wide horst may not break apart entirely at first, leaving an inverted keystone-like block in the central part. This block may rise isostatically due to loss of mass from its sides. It may rise above the surface of sedimentation, become eroded, and shed gravel into adjacent troughs. With enough extension, the keystone block will break up and sink at the surface. An example is the large horst southeast of the Rock Creek graben (fig. 2).

(E) A belt of dikes, shown 25 km wide with the greatest amount of intrusion (widening) in the central part, will cause sinking at the surface along the length of the belt. The greatest amount of extension may vary with time and may migrate both across and along the belt. The borders dip gently toward the basin. Continued injection of dikes, extension, and sinking are required for continued sedimentation as in the Upper Mississippian and Pennsylvanian parts of the Illinois basin. The Cretaceous and Tertiary sediments of the embayment are probably due to a later episode of injection along the belt of dikes. Features of sedimentation such as (4), (5), (6), (7), and (8) indicate active extension during the time of sedimentation. Extension features such as (11) indicate extension after consolidation of the host sediments.

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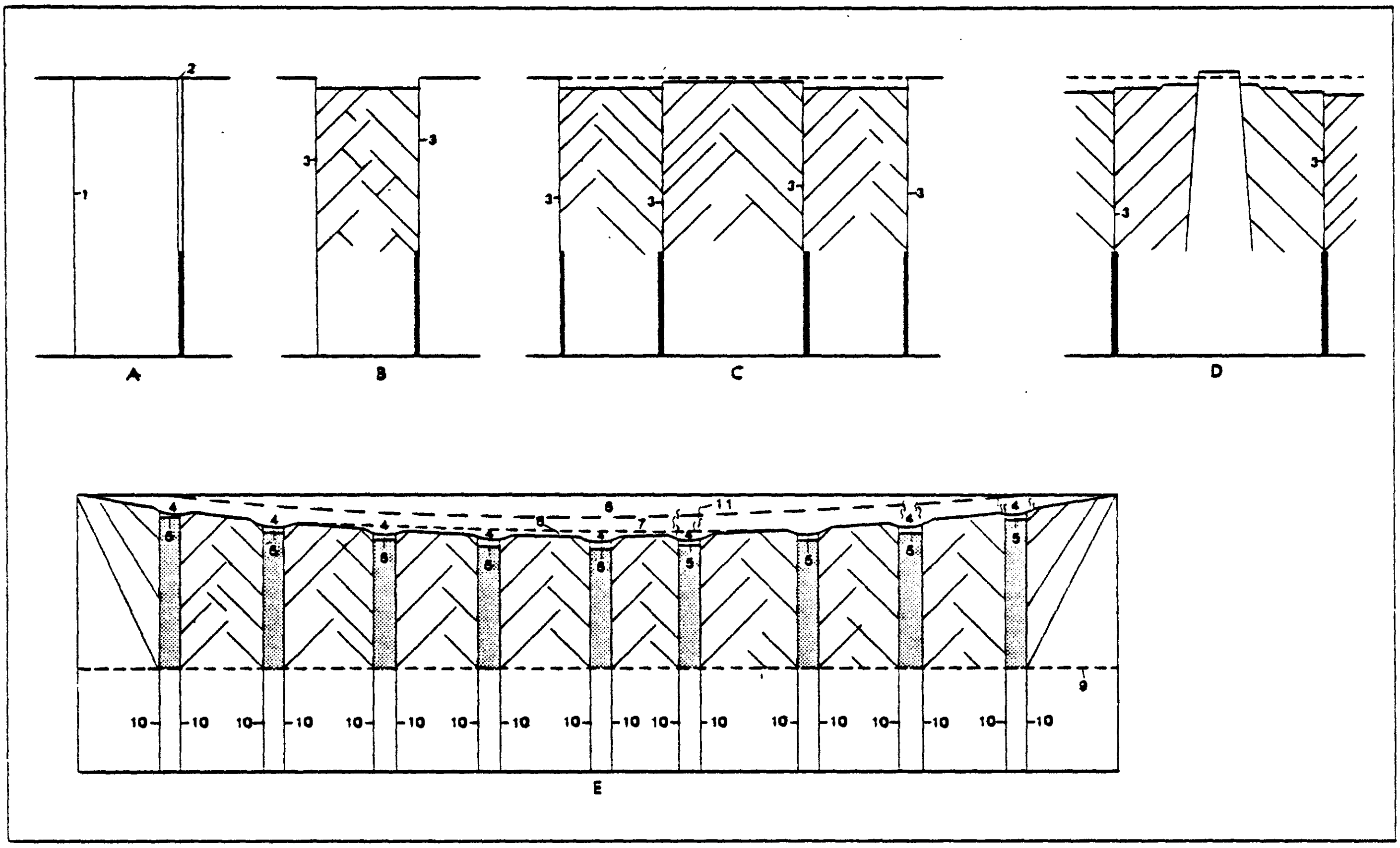


Fig 6

The hypothesis illustrated herein requires the injection of new (noncontinental) material into the continental crust or into the rocks below it. This causes extension and thinning of the continental crust. A thinner crust stands lower in elevation at its upper surface. Injection of crustal material into higher parts of the crust is merely rearrangement of the crust and causes no thinning.

Sinking of both grabens and horsts is essential to the hypothesis. However, some rising at the surface may occur in larger horsts (fig. 6D). A wide horst will not break up as readily as a narrow graben because it is larger. The upper part of the sides of the horst may collapse leaving the central part largely intact. If such occurred, the central part would rise isostatically. The rising part of the horst would be reflected in the contemporaneous sediments. If the surface of sedimentation built up faster than the horst rose, the sediments would thin across the horst. If the horst rose faster than the surface of sedimentation, the sediments would be absent. If the horst rose above sea level or above an adjacent alluvial plain, it would be eroded. Eventually, with continued extension, the horst will break up and its top will sink.

The effect of the intrusion of dikes in a wide belt, with the dikes decreasing in width from the center of the belt, is shown in figure 6E. The greatest extension of the crust by the dikes is in the center of the belt. Hence the greatest vertical shortening, to maintain constant volume above the dikes, occurs in the center of the belt. The near-surface expression of the process is a series of faults downthrown toward the axis of the belt. The surficial expression is a basin being filled or filled with sediments.

The borders of such a basin of narrow grabens do not rise and tilt back away from the basin. Loss of mass in any original column of the crust and the tendency of that column to rise is more than offset by addition of mass of the dikes and by sinking of the surface due to widening. Hence the borders slope gently toward the basin and become covered with sediments.

Mantle affinities

The mantle is the source of the dikes shown on figure 6. Nothing of the general geology of the region indicates igneous activity of crustal materials. Numerous dikes in the region are peridotite or altered peridotite (Clegg and Bradbury, 1956). The known dikes are too small to account for the required horizontal extension, but they may be small offshoots from larger bodies at depth. The source of peridotite is the mantle. Diatremes are associated with the peridotites (Clegg and Bradbury, 1956). The diatremes are extremely brecciated bodies containing fragments of peridotite, gneiss, and carbonatite.

The district contains a tremendous amount of fluorine. Almost 10 million tons of fluorspar have been mined since 1880 (Pinckney, 1976). Reserves are large, and probably much more remains to be found. Such a large quantity of fluorine is unlikely to have been contained in crustal rocks and then released to be redeposited in Carboniferous rocks. The fluorine probably came from the mantle by migrating up through the grabens and along the bordering faults.

A heat source is needed to account for the temperatures from fluid inclusions from mineral deposits of the fluorspar district. Many temperatures of homogenization of fluid inclusions have been determined, and they commonly are between 120 and 150 degrees Celsius (Pinckney and Rye, 1972; Pinckney, unpublished data). Homogenization temperatures in quartz from along a fault in the Deardorf ore body are as high as 179 degrees C. These temperatures are too high for fluids originating in the Illinois basin if the temperature gradient in the basin was a normal geothermal gradient of 25 degrees C per km. The deepest part of the Illinois basin is only 4 km deep so that temperatures of 110 degrees would be expected if the fluids came from the deepest part of the basin and lost no heat in ascending to the site of the ore deposits. A heat source is indicated, and heat flow in the district must have been high during mineralization.

The ratio S^{34}/S^{32} in galena and sphalerite from the Hill mine (Pinckney and Rafter, unpublished data) is nearly the same as the sulfur isotope ratios in large sills of basic composition (Thode, 1963). The combination of peridotites and diatremes, fluorine, high temperatures, and sulfur isotopes is strongly indicative of activities in the mantle.

Sedimentary confirmation

The sediments should contain a record of tectonism if the Illinois basin originated by the method proposed herein. Some evidence is available for the Pounds Sandstone Member of the Caseyville Formation, representing basal Pennsylvanian rocks. The sedimentary environment for the Pounds in southern Illinois was a prograding delta plain north of the fluorspar district (in the graben belt, fig. 2) and a marine-influenced, lower-delta plain west of the district (west of the graben belt) (Mansfield, 1979). A line of measured sections along the outcrops (GGP to PEP, fig. 3), in the graben belt shows varying thicknesses of the Pounds Sandstone Member (Morrow, 1979). Across the horst between the Dixon Springs and Eichorn grabens, the sandstone is 10 m thick. Its thickness increases to 14.7 m (47 percent) in 2 km in going from the horst to the Dixon Springs graben, GGP, and to 12.7 m (27 percent) on the edge of the Eichorn graben. The thicker sections reflect greater sinking of the grabens during deposition.

Four types of slump structures were found in Pounds in the graben belt. Mansfield (1979) reported three types: oversteepened crossbeds, z structures, and slump flows. The motions within these features were horizontal, but they are of noncompressional origin and frequently result from seismic shocks (Mansfield, oral commun., 1979). After the conference in Urbana in May 1979, Mansfield, Marrow, and I reexamined some outcrops of the Pounds Sandstone Member. A fourth type of slump structure was found as crossbeds deformed into numerous recumbent folds. The order of increasing deformation is oversteepened crossbeds, z structures, recumbent folds, and slump flows. The slump structures were found only in sandstone in the graben belt (from Marrow's sections GGP to PEP, fig. 3), and they are absent in sandstone several miles to the west of the graben belt. Other areas in the graben belt were not studied. The thickening of Pounds Sandstone Member in two grabens, the slump structures in the graben belt, and the lack of slumping away from the belt seems to be evidence that sedimentation reflects tectonism during deposition of the sandstone.

Extensions of the rift

The grabens of figure 2 are known only in the small area of the fluorspar district. None of the literature has yet related the grabens to the underlying structure of either the Illinois basin or the embayment, nor how the grabens have been generally recognized as the principal structural feature of the district. Heyl and Brock (1961) proposed thrusting along the north side of the district, along the Shawneetown-Rough Creek fault zone. In general, the faults seem to have been considered as of minor importance and the district as a much-faulted area, a bit of a structural anomaly, in a stable region. In contrast, data from the present study show that the faults have a simple pattern. Such a simple pattern can result only from a simple cause. That cause is considered to be a belt of dikes under the faults that border the grabens.

The extent of the belt of faults is not completely known. Faults are known to extend from the district north-northeastward along the Wabash Valley fault zone (fig. 1) and upward through younger Pennsylvanian sediments into the Mattoon Formation (Willman and others, 1967). They extend southward to the unconsolidated sediments of the Mississippi River embayment. No reason seems to exist for the faults to stop at the northern edge of the embayment. Some workers (Amos and Wolfe, 1966; Amos and Finch, 1968) show Tertiary and Cretaceous formations against Mississippian formations along faults, but they lose the faults further south in the areas of these unconsolidated sediments. Presumably the faults extend southward under, and possibly up into, the embayment sediments. Figure 1 shows the fault belt extended southwestward through the embayment. The edges of the extended belt are along strong magnetic anomalies recently found and ascribed to basic igneous intrusions at shallow depth (Hildenbrand and others, 1978).

North of the fluorspar district the fault belt, if extended, is adjacent to the deepest part of the Illinois basin and extends to the La Salle anticlinal belt (fig. 1). From structure contours (Cohee, 1962), the La Salle belt consists of a narrow anticlinal ridge bordered on each side by narrow troughs and has as much as 760 m of structural relief. This form is consistent with a band of narrow horsts bordered on each side by narrow grabens.

Other structures in the basin (Cohee, 1962) are faults and monoclines with the down-stepped side toward the axis of the basin or toward the Moorman syncline (fig. 2). The Moorman syncline trends easterly nearly parallel to grabens in Kentucky that branch from the main northeast-trending belt. This syncline is probably underlain by grabens, some of which are shown on figure 2.

Discussion

Many of the structural, stratigraphic, and sedimentary features of the Illinois basin appear to correlate with the narrow graben hypothesis proposed herein. Gray (1979), Atherton and Palmer (1979), and Rice and others (1979) gave detailed summaries of a large amount of data, some of which is discussed herein. The present basin appears to have been well established by Valmeyeran time when the Borden delta was being deposited on the edge of the basin and the Springville Shale, as foreset beds of the delta, spread across the deep (300 m) starved basin. The Fort Payne Formation, overlying the Springville Shale, occupies the deepest part of the starved basin. It is an unusual rock in that it is a very fine grained limestone, lacking clastic material, contains much chert, and thins rapidly along one edge. It is extensive to the east and southeast in the Moorman syncline, but it is thickest along its western edge, a line about 200 km long (fig. 1) in southeastern Illinois, and thins rapidly to the west. The thick western edge built up along an axis (fig. 1) nearly parallel to and partly on the axis of the graben belt and Wabash Valley fault zone with a deep and open basin to the west. The thick western edge therefore appears to be structurally controlled, possibly along sea-bottom fissures. This suggests that the Fort Payne rocks are, at least in part, chemical precipitates of $CaCO_3$ and SiO_2 deposited where warm solutions, rising along faults, emerged on the sea floor. Around Hicks Dome the writer found much disseminated fluorite in the Fort Payne cherts, but none in the overlying Harrodsburg Limestone. Following deposition of the Fort Payne Formation, the deep water trench between the Borden delta and the Fort Payne cherts was filled by Harrodsburg and Ullin Limestones. The whole sequence was then covered by the shallow-water Salem Limestone, a crossbedded biocalcarenite. The Salem Limestone thins to a knife edge to the northwest over the Borden delta. The Salem Limestone probably represents the end of a tectonically induced sedimentary sequence by which the basin was filled.

Another and similar sequence started very soon thereafter, but in shallower water, with deposition of the St. Louis Limestone. This formation is also very fine grained limestone with abundant chert in places decreasing in abundance upward into the overlying biocalcarenitic, oolitic, and very pure Ste. Genevieve Limestone. This sequence ended in the southern part of the basin with local erosion and the influx of the Rosiclare Sandstone Member of the Ste. Genevieve Limestone. Similar sequences, but in shallower water, are present throughout the Chesterian (Valmeyeran) strata (figure 10; see especially Atherton and Palmer, 1979, fig. 9). Higher in the section, such cyclic sequences are well known in the Pennsylvanian cyclothems, differing chiefly in that the rocks are mostly continental. Each sequence may represent an episode of intrusion, faulting, and sinking.

The southern edge of the Illinois basin is uplifted, exposing the deeper rocks of the fluorspar district and surrounding area of figure 2. These deeper rocks, of Mississippian age, are as much a part of the depositional basin as are the Pennsylvanian cyclothems. They, too, are cyclic, but they are much faulted. Hence the faulting must have progressed upward through the section. The observed faults extend into the Pennsylvanian rocks in the Moorman syncline to the east where they can be mapped from outcrops, but they appear to end to the northeast in Illinois where the basinal rocks are covered by glacial deposits and are mapped by subsurface methods.

Outcrops of the Caseyville Formation should be studied in detail for additional features of faulting concurrent with sedimentation. Most other Pennsylvanian beds cannot be studied in much detail because of lack of exposures. However, Pennsylvanian coal beds are again being actively mined and exposed, even if briefly, and they can yield much data. Recently the number 6 coal bed has been shown to have down-dropped blocks between vertical faults (Popp and De Maris, 1979), tension fractures in coal filled with sulfides, calcite, and barite (Cobb and Krausse, 1979), and clay dikes injected into vertical fractures (Krausse and Damberger, 1979). All of these features require extensional tectonics that are post-coalification. Possibly, patterns in coal deposition exist and reflect a genetic connection between sedimentation and underlying grabens. Bauer (1979) noted that the Du Quain monocline in Illinois (probably overlying a fault) was structurally active during the time of the numbers 5 and 6 coal swamps.

Tectonic activity of the Du Quain monocline correlates with the north-trending veins of sulfides of Cobb and Krausse (1979). It also may account for the fluid inclusion temperatures of 80°C in the sphalerite veinlets (Cobb and Krausse, 1979) by the upward flow of hydrothermal fluids along the fault. Similar veinlets are reported for Iowa and Missouri and may be common in many coal basins. If so, the veinlets may be indicative of grabens and a distribution pattern would be diagnostic. In turn, belts of high sulfur coal might be predicted from the tectonic pattern so that the coal could be selectively mined and treated.

Mineralization such as occurred in the fluorspar district may have been more extensive than the exposed district. The exposure of the district is related to uplift on the arch and erosion. Because mineralization is related to the faults and presumably the underlying dikes, it could have occurred in other places along the rift zone. If so, mineral deposits of Mississippi Valley type could be present both in the Illinois basin and in the embayment.

Basins of extensional origin may be common on the continents, with the underlying structure largely covered with sediments, but reflected in the sediments. With extension due to dikes or other igneous bodies, contributions from depth can be expected in the basins. The most likely contributions are hydrothermal solutions that bring in heat, metals, and the nonmetals, notably fluorine and sulfur. The heat accelerates the maturation of kerogen and coal and the conversion of gypsum to anhydrite. The dissolved elements add to the complex chemistry of the basins as minor elements and as mineral deposits of both epigenetic and syngenetic origin.

APPENDIX

Mechanics of stress

Stresses in the district

Hubbert (1951) clearly pointed out that in stable areas the greatest force acting on the rocks of the crust is the force of gravity which is vertically directed. If stress from outside the area, such as from tectonism or high relief, is absent, the greatest principal stress, σ_1, results only from the force of gravity acting on the mass of the rocks; σ_1 is therefore vertically directed. It acts on all horizontal planes in the rock column, is equal to the weight, ρgz, of the overlying rocks, and is

(1) $$\sigma_1 = \rho gz$$

where ρ is average density, g is the force of gravity, and z is depth. Consequently, the intermediate and least principal stresses, σ_2 and σ_3 respectively, are horizontally directed, act on mutually perpendicular vertical planes, and are the confining stresses of rock mechanics.

Rocks in the upper part of the crust, and probably to considerable depth, are elastic materials whose behavior under stress conforms to Hooke's law. From the three axes, or three dimensional, equations of Hooke's law the confining stresses resulting from σ_1 in stable areas are given by Howard and Fast (1970, p. 19) as

(2) $$\sigma_2 = \sigma_3 = \sigma_1 \frac{\nu}{1-\nu} = \rho gz \frac{\nu}{1-\nu}$$

for elastically isotropic rocks, where ν is Poisson's ratio. Hence the stresses before tectonism result entirely from the mass of the rocks, gravity, and an elastic property of the rocks. The principal stresses given by (1) and (2) can be resolved into components of shear and normal stresses acting on any inclined plane. The usual equations for these stresses are (3), and (4), which are given in figure 7.

Figure 7.--NEAR HERE

Figure 7.--Two well known methods of defining a stress system by means of two principal stresses σ_1 and σ_3. The systems so defined are called two dimensional or plane stress, and their use assumes that no change occurs in the third orthogonal axis σ_2. In (a) the σ_2 plane is parallel to the plane of the drawing. The greatest principal stress, σ_1, is directed (arrow) onto the σ_1 plane (horizontal) and the least principal stress, σ_3, is directed onto the σ_3 plane (vertical). The stress on all other planes normal to σ_2 plane can be determined by resolving σ_1 and σ_3 onto those planes as given in standard texts. The stresses so derived are a stress directed perpendicularly onto the plane, called normal stress, σ_n, and a stress directed parallel to the plane and called shear stress, τ. The plane is defined by the angle α and the stresses are given in equations (3) and (4). Note that α of this figure is θ in texts of stress mechanics, and ∂ of this figure is θ, fracture angle, of rock mechanics (fracture mechanics); see for example Handin (1966, fig. 11-4), and compare to Jaeger and Cook (1969, fig. 2.3.1). For this paper σ_1 is the load of the overburden and is vertically directed onto horizontal planes; σ_3 is due to σ_1 acting through the elastic properties of the rocks and is horizontally directed onto vertical planes.

Figure (b) is Mohr's (Jaeger and Cook, 1969, p. 15) simultaneous solution of all values of equations (3) and (4) resulting from different values of α for preset values of σ_1 and σ_3. The result is a circle of radius $(\sigma_1-\sigma_3)/2$ centered on the σ_n axis at $(\sigma_1+\sigma_3)/2$. Two examples are given, one each for a set of conditions defined by σ_1 and σ_3 and the stresses on the plane, P. The normal stress, σ_n, and shear stress, τ, for P can be read directly from the diagram. Any plane, P, is a potential fracture plane and is defined by a point on a Mohr stress circle.

Top.

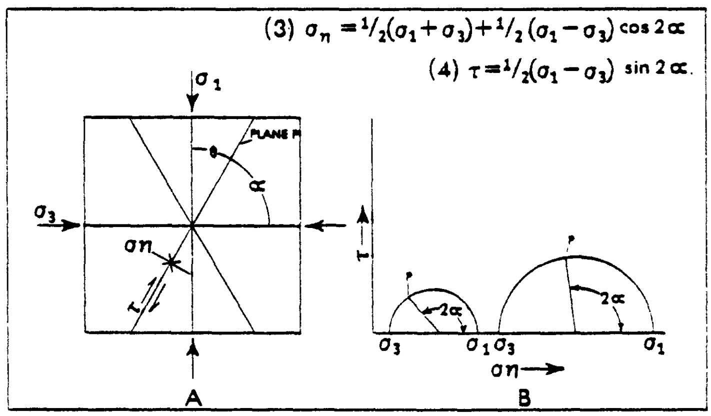

(3) $\sigma_n = {}^1/_2(\sigma_1+\sigma_3)+{}^1/_2(\sigma_1-\sigma_3)\cos 2\alpha$
(4) $\tau = {}^1/_2(\sigma_1-\sigma_3)\sin 2\alpha$.
σ_1
PLANE P
θ
α
σ_3
σn
τ
τ
P
2α
P
2α
σ_3
σ_1 σ_3
σ_1
σn
A
B

Fig 7

The stresses given by (1) through (4) are carried by the framework of solid grains in the rocks. The stress to which a porous rock filled with fluid reacts is the effective stress, $\bar{\sigma}$, of Terzaghi (Terzaghi, 1925, 1943) given by (5)

$$\bar{\sigma} = \sigma - u \tag{5}$$

where u is called the neutral stress and is the piezometric pressure of fluid in pores of the rocks. For most sediments with interconnected pores open to the surface, u is $\rho_f gz$, where ρ_f is the density of the contained water. Equations (1) and (2) therefore become (5) and (6)

$$\sigma_1 = \rho gz - u \tag{5}$$

$$\sigma_2 = \sigma_3 = \rho gz \frac{\nu}{1-\nu} - u \tag{6}$$

By this usage the normal stresses of (1), and (2), and σ_n of (3), figure 7, are reduced by the amount of the neutral stress to become the effective stress. Because Poisson's ratio is rather small for most sedimentary rocks (Birch, 1966, table 7-15), the confining stresses, σ_2 and σ_3, for many sedimentary rocks is small. For example, Poisson's ratio is 0.25 for the Redwall Limestone in Arizona (Birch, 1966, table 7-15, p. 165). If this limestone were carrying a load of overburden 1 kilometer thick of density 2.37, σ_1 would be 232 kg cm^{-2} and σ_2 and σ_3 would be 77 kg cm^{-2}. If the water table were at the surface, the neutral stress would be about -98 kg cm^{-2}, and the normal stresses on all planes in the rock would be reduced by 98 kg cm^{-2}. This would make the effective principal stresses 134 kg cm^{-2} for σ_1 and -20.6 kg cm^{-2}

for σ_2 and σ_3. The usage here is that compressive stress is positive (+) and tensile stress is negative (-). Tensile stress can therefore be expected in some sediments under nontectonic conditions. Any slight horizontal extension of the crust would increase the tensile stress across vertical planes.

Mechanics of fracture

Rocks fracture along planes, and any plane in a solid body is a potential fracture plane. The planes, designated by the angle α in figure 7, each are characterized at the time of fracture by particular values of both shear stress, τ, and effective normal stress, $\bar{\sigma}_n$. Shear stress tends to tear the material apart. Normal stress on any plane tends to hold the material together if the normal stress is compressive, and to tear the material apart if it is tensile. A simple means of showing the shear and normal components of stress on every plane is the stress circle, credited to Otto Mohr in 1900 (Jager and Cook, 1969). According to Mohr's hypothesis, because the state of stress in a body changes up to the time of fracture, a stress circle can depict the state of the stresses on all planes at any time, such as at the time of fracture. Similarly, because a given material can fracture under a variety of stress conditions, a different stress circle of a different size would depict the state of stress in the body for each set of conditions. The conditions are easily defined by a statement of the principal stresses. For intermediate sets of conditions, the state of stress on all planes would be represented by circles of intermediate size and they would be centered at intermediate positions on the $\bar{\sigma}_n$ axis of a shear stress-normal stress diagram (fig. 8). A line tangent to all the circles depicting a group stress condition at fracture separates conditions under which fracture occurs from those under which it does not occur. Such a line, called the fracture line, defines the values of shear stress and effective normal stress on the planes along which the material will fracture. Because the line is tangent to the circles it also defines, by the point of tangency, which planes of figure 7 will become the fracture plane. What is usually unknown is the position of the line on the diagram, its curvature, and how far it can be projected. The nature of the line must be dependent upon some property of the material.

Griffith theory

Griffith (1920, 1924) calculated the form of fracture lines using unconfined tensile strength as the property of materials. His line is a parabola whose size is defined by letting tensile strength be represented by the distance from the focus to the vertex of the parabola. Such a parabola is shown in figure 8, and in terms of two principal stresses the parabolas are given by (7)

$$\frac{(\bar{\sigma}_1 - \bar{\sigma}_3)^2}{\bar{\sigma}_1 + \bar{\sigma}_3} = -8T \tag{7}$$

where T is tensile strength and is inherently negative if compressive stresses are positive, and where the limits of the relation are given by (8) and (9)

$$\bar{\sigma}_1 + 3\bar{\sigma}_3 > 0, \text{ and if} \tag{8}$$

$$\bar{\sigma}_1 + 3\bar{\sigma}_3 < 0, \ \bar{\sigma}_3 = T \tag{9}$$

so that tensile strength is the limit of $\bar{\sigma}_3$.

Figure 8.—NEAR HERE

Figure 8.—Griffith's (1920, 1921) criterion of fracture plotted on a σ_n-τ diagram.

For a given material fracture occurs on different planes P at different values of α as stress conditions of σ_1 and σ_3 are changed. The values of σ_n and τ on the fracture planes are given by the fracture line which is a parabola. The position of the parabola on the σ_n axis is determined by a property of the rock, tensile strength, T, which is plotted as a negative (-) quantity because compressive stress is considered as positive (+). The origin of the diagram is at the focus of the parabola. Because the fracture plane P is a point on Mohr's stress circle, contact of the stress circle and fracture line define conditions of stress in terms of σ_1 and σ_3 at the time of fracture for a rock characterized by a value of T. The relations of the stress circle in contact with the fracture line are given by equations (10) and (11). Hence using the load of the overburden at the time of faulting for σ_1 and the dip angle α (2θ is the supplement of 2α because θ is the complement of α) the equations can be solved for T and σ_3. The direction of change of σ_3 tells if the stress system changed by becoming more compressive or more tensile. The scale is the same for both axes.

Top

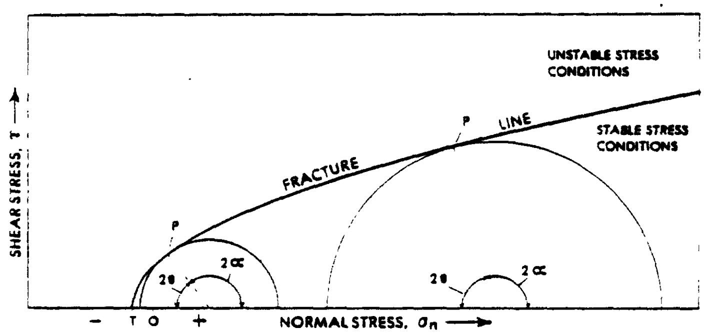

UNSTABLE STRESS CONDITIONS
STABLE STRESS CONDITIONS
FRACTURE LINE
P
P
2θ
2α
2θ
2α
SHEAR STRESS, τ
−
T
O
+
NORMAL STRESS, σ_n

Fig 8

The relations among Griffith's fracture line, stress circles representing the stress conditions at the time of fracture, and the plane on which fracturing occurred were worked out in the present study. Those pertinent to the problem here are (10) and (11)

$$\bar{\sigma}_1 \frac{\cos^2 2\theta}{1 + 2 \cos 2\theta} = T \tag{10}$$

$$\bar{\sigma}_3 \frac{\cos^2 2\theta}{1 - 2 \cos 2\theta} = T \tag{11}$$

where θ is the complementary angle of α so that 2θ is the supplementary angle of 2α in the stress circle. These equations contain only two independent variables.

The two independent variables can be obtained from fault dips and an estimate of overburden at the time of faulting. According to Hubbert's concept of the direction of the σ_1 axis (Hubbert, 1951), the angle α is the dip angle of the faults bordering the grabens in the district. The dips were determined in 39 places by using data gathered in mining and drilling operations. Many of the bordering faults contain veins of fluorspar. Much of the data used for determining the dips was given by Williams and others (1955). Only dips calculated from footwalls and consistent over a vertical interval of 30 meters or more are used. The fault dips are given in figure 9.

Figure 9.--NEAR HERE

Figure 9.--Dip of faults bordering grabens.

The dips are plotted as the angle between a horizontal plane extending across the graben and ending at its borders. Three faults dip toward a horst and probably represent motion on conjugate fracture planes that correspond to faults dipping at the same angle toward the grabens, as indicated by the dashed line. Viewing the diagram as a frequency diagram, the greatest peak is at 90 degrees with lesser peaks at 69, 80, and 85 degrees.

Top

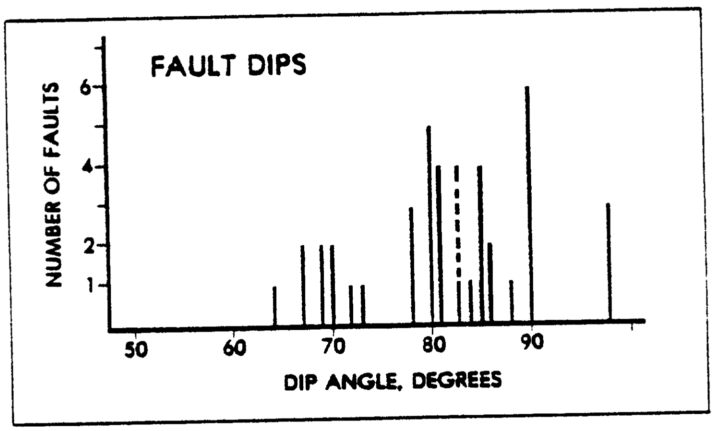

FAULT DIPS
NUMBER OF FAULTS
1
2
4
6
50
60
70
80
90
DIP ANGLE, DEGREES

Fig 9

Evidence of the earliest known faulting was given by Desborough (1961) who found faults in the lower part of the Caseyville Formation in the Pomona Quadrangle, about 60 miles west of the fluorspar district. The faults offset beds of the Caseyville Formation, and the fault scarps are covered by slightly younger Caseyville beds. Therefore, recognized faulting seems to have started about as early as the deposition of the oldest Pennsylvanian rocks in the area. Because many of the faults given in figure 9 are in the upper member of the St. Louis Limestone (fig. 10), the thickness of the overburden is taken as that of the Mississippian rocks above the base of the upper member. From figure 10 this is 493 meters. The average density used for the rock is 2.37 g cm^{-2}. This value was arrived at by comparing the lithologies represented in figure 10 to rocks of similar lithology for which densities are known (Birch, 1966, table 7-15). A value of 980 cm sec^{-2} was used for gravity. These values applied to equation (1) give a value of 114.5 kg cm^{-2} for the weight of the overburden, $\bar{\sigma}_1$. The Lower Pennsylvanian sandstones are chiefly fluvial and deltaic deposits so that the water table was near the surface, and the height of the piezometric column was the height of the rock column, 493 m. Using a density of 1 g cm^{-3} for groundwater gives the neutral stress as 48.2 kg cm^{-2} and from equation (5) the effective greatest principal stress was 66.2 kg cm^{-2} at the time of fracturing, giving a value of $\bar{\sigma}_1$ for use in equation (10).

Figure 10.--NEAR HERE

Figure 10.--Mississippian formations of the Illinois-Kentucky fluorspar district.

The chief ore-bearing beds are shown by a solid bar. Along many of the faults of figure 9 these beds are adjacent to the St. Louis Limestone.

Adapted from Pinckney (1976).

SYSTEM	SERIES	STRATIGRAPHIC UNIT	LITHOLOGY	APPROXIMATE THICKNESS, IN FEET (METRES)
Pennsylvanian	Lower Pennsylvanian	Caseyville Formation		
MISSISSIPPIAN	Chesterian	Kinkaid Limestone		40(12)
		Degonia Sandstone		30(9)
		Clore Limestone		112(34)
		Palestine Sandstone		56(17)
		Menard Limestone		135(41)
		Waltersburg Sandstone		36(11)
		Vienna Limestone		16(5)
		Tar Springs Sandstone		161(49)
		Glen Dean Limestone		75(23)
		Hardinsburg Sandstone		98(30)
		Golconda Formation		125(38)
		Cypress Sandstone		200(61)
		Paint Creek Formation		
		Bethel Sandstone		
		Renault Formation		79(24)
	Meramecian	Ste Genevieve Limestone: Levias Ls Mbr		20(6)
		Ste Genevieve Limestone: Rosiclare Ss Mbr	SR	30(9)
		Ste Genevieve Limestone: Fredonia Limestone Member		226(69)
		Upper member of St. Louis Limestone		230(70)
		Lower member of St. Louis Limestone and Salem Limestone		374(114)
		Warsaw Limestone		200(61)
	Osagean	Fort Payne Formation		558(170)
Miss. and Devonian	Lower Miss. and Upper Devonian	Chattanooga Shale		

Fig 10

Solving (10) for each dip angle gives a value of tensile strength, T, corresponding to each dip, that is, a value of T for each rock unit having a characteristic dip. Solving equation (11) for each dip angle and tensile strength gives a corresponding value for $\bar{\sigma}_3$ so that the stress system for each fault dip is completely described. The results of the calculations are given in shortened form in table 1.

Table 1.--NEAR HERE

Table 1 shows a progression of values of both tensile strength and $\bar{\sigma}_3$ corresponding to changes in dip angle. At the lowest dip angle, tensile strength is 11.2 kg cm^{-2} and changes to 22.1 kg cm^{-2} at 90 degrees. Similarly, at the lowest dip angle, $\bar{\sigma}_3$ is 6.86 kg cm^{-2} and changes to 22.1 kg cm^{-2} at 90 degrees. A remarkable correlation is evident with lower values of $\bar{\sigma}_3$ corresponding to the stronger rocks (larger absolute values of T). The question can now be posed: did $\bar{\sigma}_3$ increase or decrease (become more compressive or more tensile)? It decreased. The proof is given in figure 11 and below.

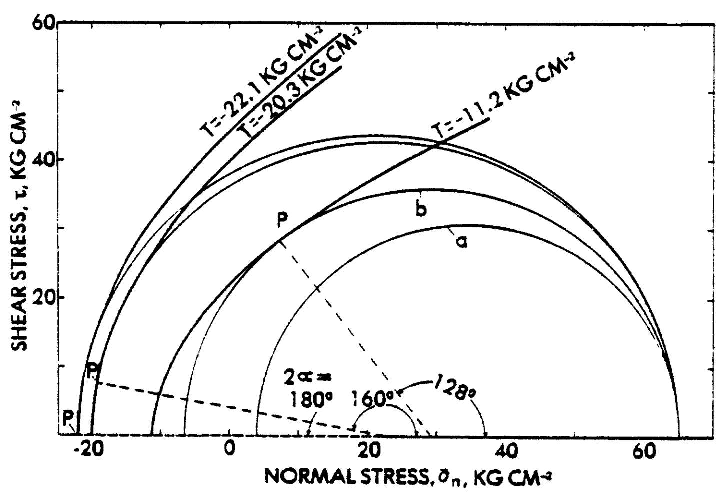
Figure 11.--NEAR HERE

Table 1.--Values of $\bar{\sigma}_3$ and T in kg cm^{-2} for rocks at the base of the upper member of the St. Louis Limestone from equations (10) and (11). The assumption is made that the rocks fractured at the end of Mississippian time so that the depth of overburden, z, was 493 m. Other values used are ρ, 2.37 g cm^{-3}; ρ_f, 1.0 g cm^{-3}; and g, 980 cm sec^{-2}, giving a value of $\bar{\sigma}_1$ of 66.19018 kg cm^{-2} for calculation. The values for α of 85 and 90 degrees are given to six figures to show that small differences do exist between $\bar{\sigma}_3$ and T for 85 degrees and that $\bar{\sigma}_3$ and T are equal for 90 degrees. At 90 degrees equation (9) is at the limit of zero where $\bar{\sigma}_1$ is 66.19018 kg cm^{-2}.

	Degrees				
d	64	69	80	85	90
$\bar{\sigma}_3$	-6.86	-12.9	-20.2	-21.6119	-22.0634
T	-11.2	-14.7	-20.3	-21.6170	-22.0634

Figure 11.--Three conditions of stress at three times of fracture. The value of $\bar{\sigma}_1$ was constant at 66.2 kg cm^{-2}. If an increase in $\bar{\sigma}_3$ occurred, a stress circle such as _a_ would become smaller, and the rocks would be stable (fig. 8). If, instead, a decrease in $\bar{\sigma}_3$ occurred, the circle would become larger, such as from _a_ to _b_, and could contact a fracture line. Hence fracture did occur by a decrease in $\bar{\sigma}_3$, causing the stress circle to become larger and to contact successively larger parabolas representing increasingly stronger rocks. The parabolas shown are for the faults dipping 64, 80, and 90 degrees.

Top



T=-22.1 KG CM-2
T=-20.3 KG CM-2
T=-11.2 KG CM-2
SHEAR STRESS, τ, KG CM-2
NORMAL STRESS, σn, KG CM-2
60
40
20
-20
0
20
40
60
P
P
P
b
a
2α =
180°
160°
128°

Fig 11

Figure 11 shows three parabolas representing fracture lines for rocks of tensile strength -11.2, -20.3, and -22.1 kg cm^{-2}. If before any deformation occurred, the conditions of stress can be represented by stress circle _a_, all rocks shown were stable because no plane exists for circle _a_ with shear stress great enough to cause fracture. Because the load and neutral stress are fixed, $\bar{\sigma}_1$ is fixed. Hence the stress conditions can change only by a change in $\bar{\sigma}_3$. If $\bar{\sigma}_3$ increased, the shear on each plane would decrease and the rocks would be even more stable. However, if $\bar{\sigma}_3$ were to decrease, the shear stress on each plane would increase, the compressive normal stress on each plane (except $\bar{\sigma}_1$) would decrease and could even become tensile, circle _b_. Circle _b_, with $\bar{\sigma}_3$ of -6.86 kg cm^{-2}, is tangent to a parabola representing rock tensile strength of -11.2 kg cm^{-2}, the weakest rock noted in table 1. The angle 2α for the point of tangency is 128 degrees, representing the fault dipping 64 degrees. Further decreases in $\bar{\sigma}_3$ would cause successively larger stress circles to become tangent to successively larger parabolas representing successively stronger rocks. Therefore, the fracturing of the rocks resulted from a decrease in $\bar{\sigma}_3$, and the succession of events is that seen by reading table 1 from left to right.

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