A Paleoseismic Investigation of the McGregor Fault, East-Central New York

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ABSTRACT

The structural disposition and movement potential of the McGregor Fault System in Eastern New York State are reviewed from the perspective of a paleoseismic investigation. The McGregor Fault zone has been located beneath deltaic sediments west of West Glens Falls using commercial and residential bore hole data and microgravity surveys. A 34 m long trench was excavated in overburden near the western margin of the McGregor Fault zone to a depth of 3.2 m. Stratified fluvially deposited sediments were exposed which showed no evidence of earthquake induced deformation or offset. However, the overburden present here does not possess optimal qualities necessary to record liquifaction events. Also, the total thickness of the sediment and the large width of the fault zone relative to the trench length make encounters of earthquake induced offset effects unlikely at this locality.

ACKNOWLEDGEMENTS

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1.0 INTRODUCTION: STATEMENT OF THE PROBLEM, AND THE RESEARCH TECHNIQUE

The problem to be addressed in this study was to determine whether any evidence of geologically recent (post-glacial) fault activity could be found on the McGregor fault of New York State, and indirectly, to make evident the need for such an investigation for estimating earthquake risk (given the potential implications). Earthquake hazard assessments in intraplate areas such as the eastern United States typically involve a great deal of ambiguity, one source of which stems from a reliance upon historic events. Nishenko and Bollinger (1990) forecast that the central and eastern United States has approximately two-thirds the likelihood of California to experience an earthquake with a "comparable damage area" within the next 30 years. This is because for a given earthquake magnitude, seismic waves are poorly attenuated in the East and peak horizontal ground accelerations are greater (Nuttli, 1981). Also, the eastern United States contains a very high population density, and the buildings are typically very old. Moderate to severe earthquakes have occurred in or near major metropolitan areas (New York City, for example; in 1884, 1893, 1916; Coffman & Hake; 1973), and are likely to recur, yet current construction codes (outside of Boston) take no note of the possibility. The regulation of building safety is an established responsibility of government, and while the public remains completely ignorant of any risk whatsoever, most elected benefactors are similarly complacent in assuming that no risk exists, and no precautions are necessary or justified. Admittedly, the risk to any one individual is small, and while new structures in earthquake prone areas can be designed to better resist seismic activity for about

2% of the cost of the building (Whitman, 1989), what legislator would risk political embarassment by proposing such protection (better to wait until after a disaster)?

While earth scientists can mathematically estimate the probability of a future earthquake, they lack an accurate model to explain the mechanisms responsible for intraplate (within plate boundary) activity, and they are even less prepared to predict (or constrain) the location and magnitude of the next significant event. Geoscientists seem to be in total agreement, however, that the brief human historical earthquake record of this region will not simply repeat itself for the next 350 years.

A basic feature of seismic behavior is that earthquakes have an ominous propensity to recur in areas where they have occurred before. At first glance, this observation may not seem true for all intraplate activity. However, it has been demonstrated that intraplate earthquake episodes commonly have long recurrence intervals. It is on this premise that the task of assessing earthquake risk is primarily based. Paleo-seismological techniques have successfully extended the age resolution of this task and demonstrated instances where the potential hazard is much greater than previously thought. This has been accomplished by assessing the post-depositional features of prehistoric soft sedimentary deposits in combination with age data to obtain an account of earthquake occurrences much longer than that of the human historical record. This type of paleoseismic investigation includes trenching through overburden deposits and requires that certain prerequisites are satisfied. These include i) the identification of a potentially active fault; ii) finding a locality where the fault is buried by shallow, undisturbed (by man), stratified overburden; iii) precisely locating the fault at depth where it intersects the overburden. The bulk of effort spent on the present investigation involved satisfying these conditions. They are the foundation of this type of research and ensure that the trench is constructed in the location most likely to yield indications of movement on the fault, or attendant seismic disturbance(s).

There is some indirect evidence which implies that the McGregor Fault has been active within Neogene time. Much of this evidence will be reviewed in order to justify the selection of the McGregor Fault as a candidate for this investigation.

1.1 THE MCGREGOR-SARATOGA-BALLSTON LAKE FAULT SYSTEM

The McGregor-Saratoga-Ballston Lake Fault System forms the southeastern border of the Adirondack Mountains from just north of Glens Falls to Saratoga Springs, where the fault splays. From Saratoga Springs, the main fault extends southward to the Town of Altamont (Figures 1.1, 1.2, and 1.3), where it is lost up-section. This fault system is the southern end of a longer zone of high angle faults that form the entire eastern border of the Adirondack Mountains and extend into Canada (Isachsen and McKendree, 1977). The fault system is actually an 85 km network of several fault segments, each ranging in length from 1 km to 15 km, typically maintaining an average NNE strike by alternately following a northerly, and then a northeasterly striking segment. Like similar large faults in the Mohawk Valley and eastern Adirondack regions, the displacements are primarily downstepping toward the east (Figure 1.4). Along the northern segments an obtrusive topographic scarp delineates the fault but decreases gradually in relief from 355 m west of Glens Falls, to 235 m north of Wilton, and to 85 m north of Saratoga Springs. Minimum fault displacements (stratigraphic throw) at the same localities are 730 m, 600 m, and 440 m respectively (Geraghty & Isachsen, 1981). An exact estimate of the stratigraphic displacement is not possible north of Daniels Road, Saratoga Springs, where, in addition to Paleozoic cover rocks, an unknown thickness of Proterozoic basement has been eroded from the upthrown block as well.

In rare exposures, the fault zone characteristically contains brecciated angular rock fragments (up to 1 m long) from the upthrown block or both fault blocks. Locally, closely bracketing outcrops

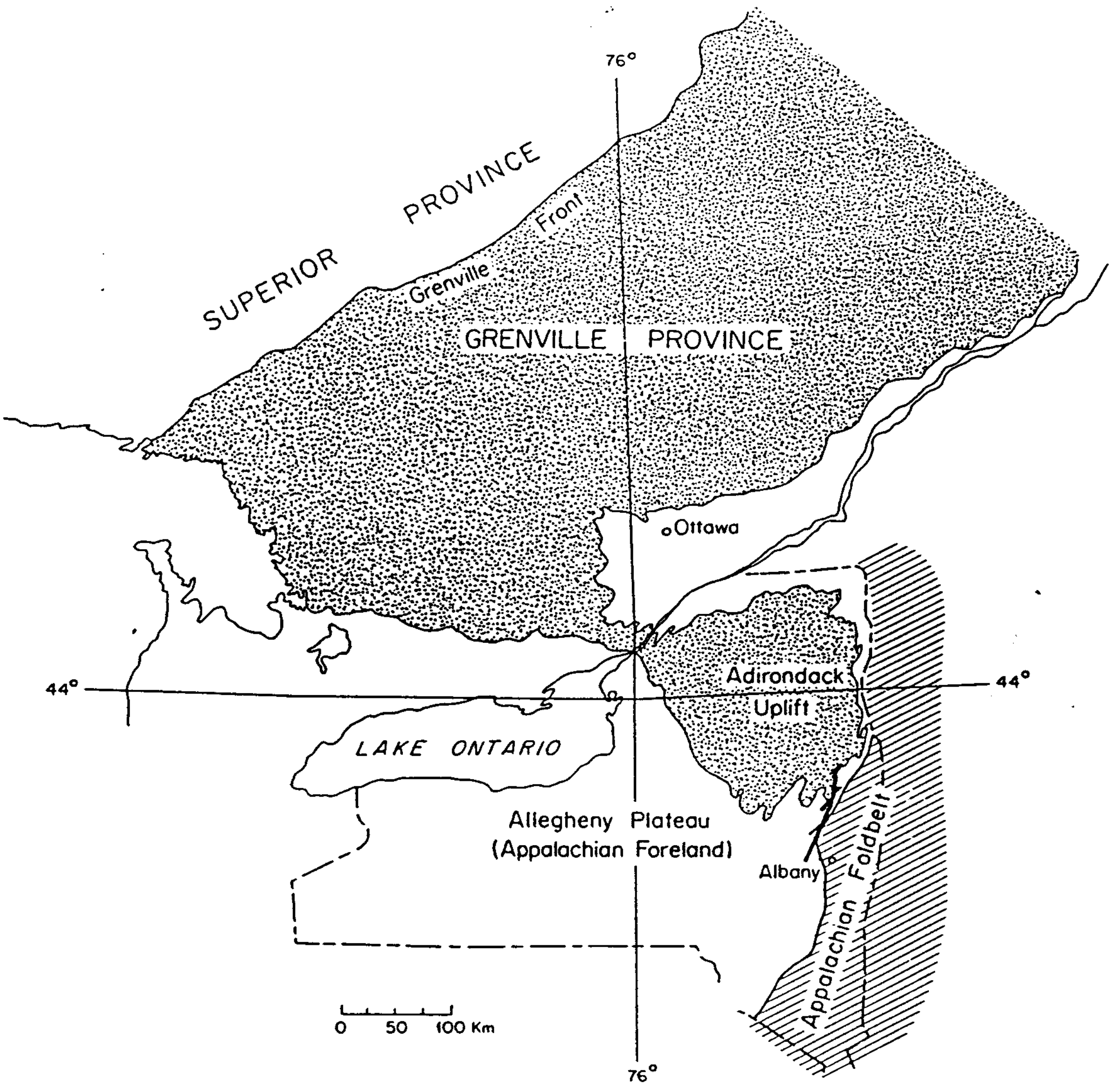


Figure 1.1 Map showing location of the McGregor-Saratoga-Ballston Lake Fault System (heavy lines) in relation to the Adirondack Uplift (an extension of the Grenville Province) and the boundary between the Appalachian Foldbelt and Foreland (From Geraghty and Isachsen, 1980).

Figure 1.2 Topographic map showing fault controlled relief. Heavy lines represent the McGregor Fault System.

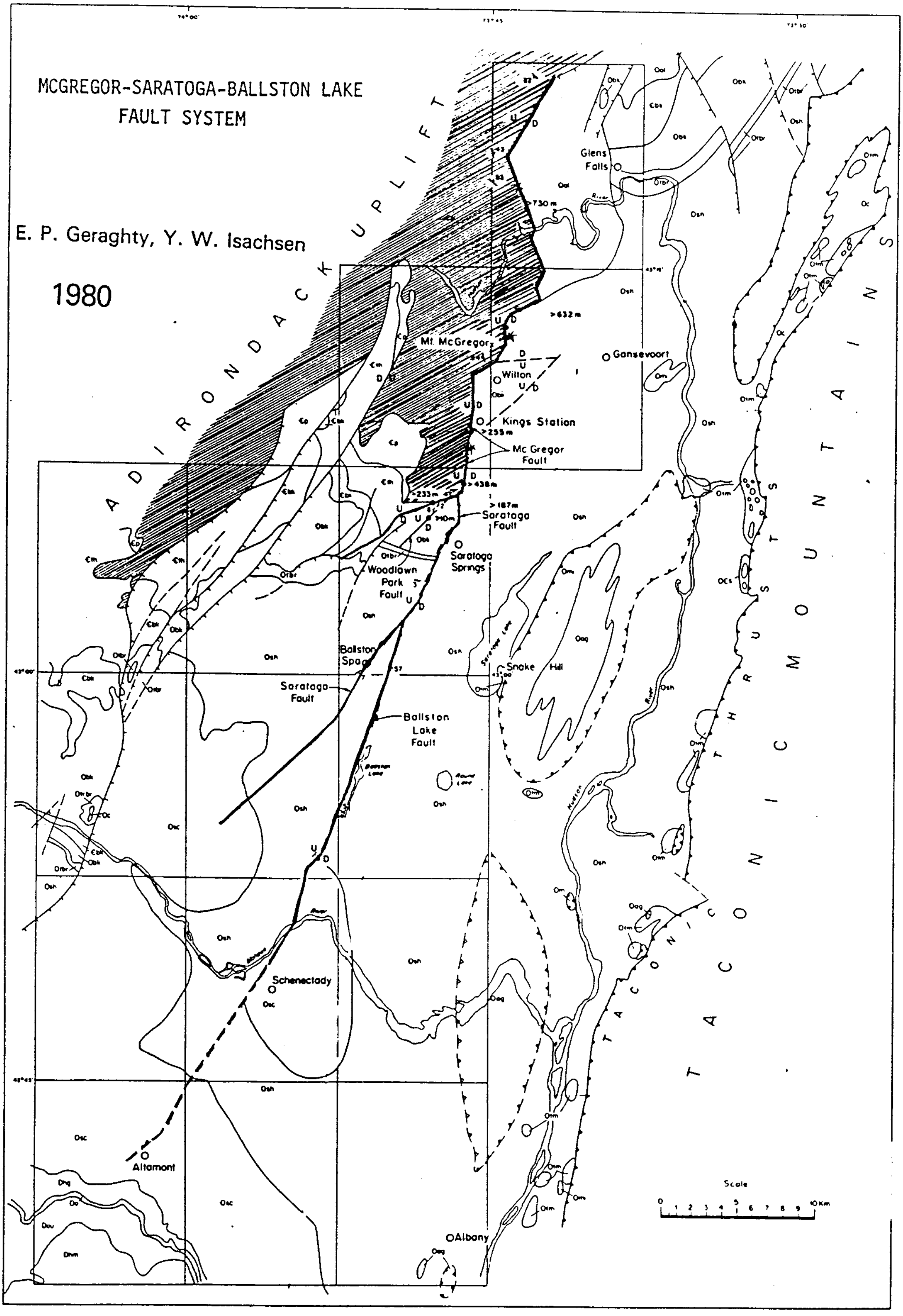
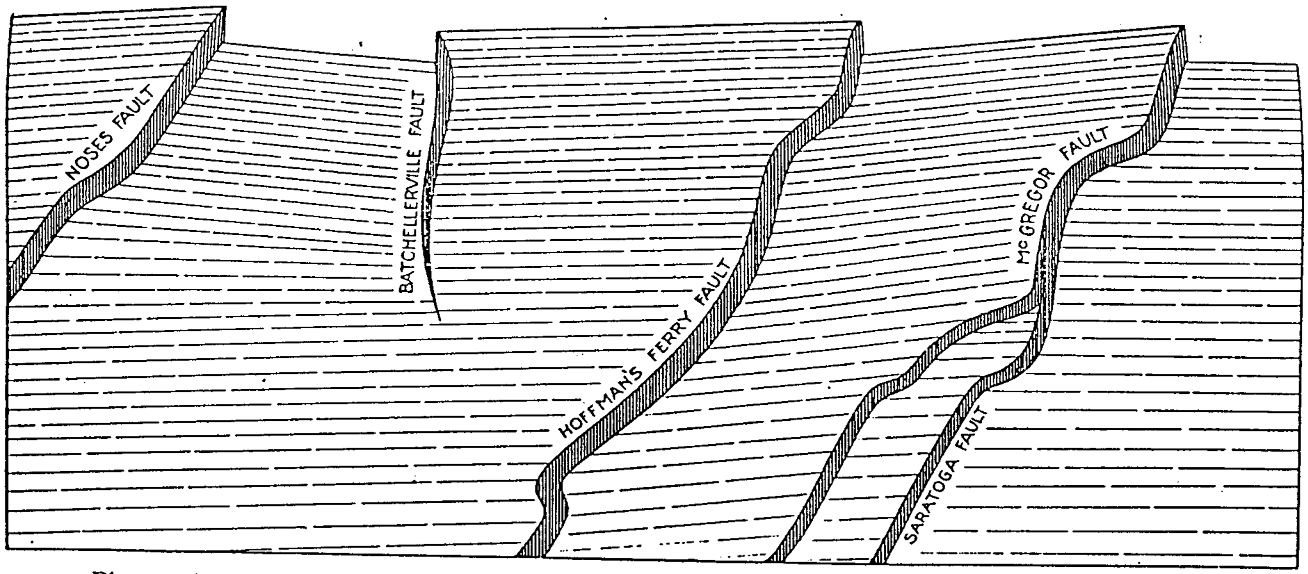
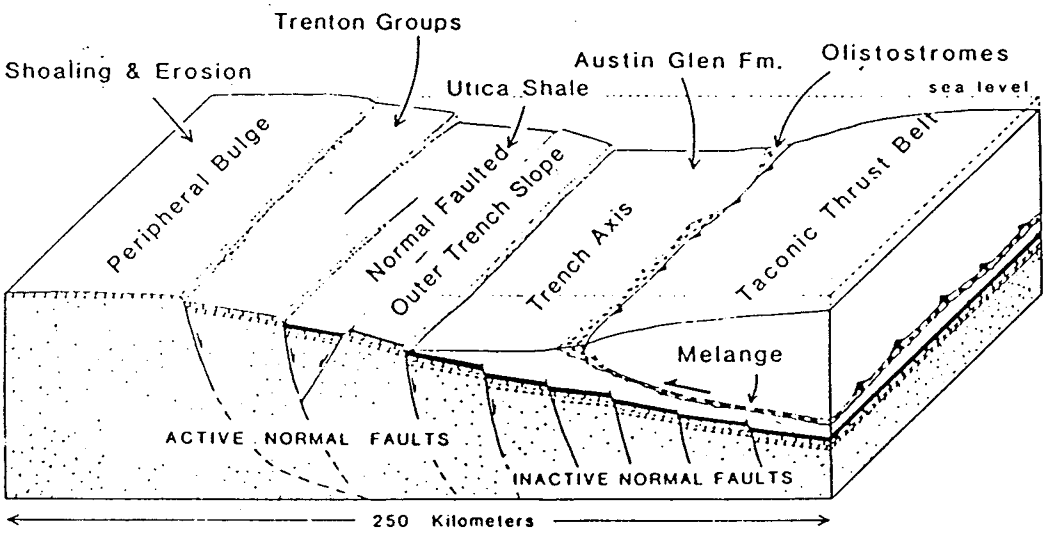


Figure 1.3 Map of the McGregor-Saratoga-Ballston Lake Fault System. Parallel lines indicate proterozoic gneisses; other lithologies are paleozoic. The location of narrow horses, slivers of Utica Shale wedged within the fault zone are noted by the *'s, north and south of the Wilton Horst. U/D indicates upthrown vs. downthrown fault relations. (from Geraghty and Isachsen, 1980).



- Diagram of the four principal faults of the Saratoga and Broadalbin quadrangles to illustrate the manner in which they longitudinally slice the country. Minor faults and branch faults are omitted. The Batchellerville fault also is not a member of the principal group but a subsidiary break in the block between the Noses and Hoffmans Ferry faults.
Vertical scale greatly exaggerated.

Figure 1.4 From Cushing and Ruedemann, 1914



—Schematic paleogeographic block diagram of the Taconic foreland basin, New York, near the close of the Taconic Orogeny. Stratigraphic units are represented on the front face as follows: dolostone pattern = Cambrian and Lower Ordovician; limestone pattern = Black River and Trenton Groups; black = Utica Shale; stipple = Austin Glen Greywacke; conglomerate pattern = melange and/or olistostromes.

Figure 1.5 From Bradley and Kusky, 1986

constrain the fault zone width to typically less than 10 m. Where bedding is exposed near the fault, it is conspicuously drag-folded, especially in the shales of the downdropped (eastern) block. Exposures of the Utica shale just east of the main fault north and south of the Wilton Horst (named here for the first time, Figure 1.3) dip toward the east as much as 65°. Fault drag effects are rarely displayed in the more competent beds of the upthrown block although Isachsen et al., (1981) reported Gailor Dolostone beds (lower Ordovician) with dips of 44° E, 85 m west of the inferred fault zone just north of Saratoga Springs, that flatten to a 6°E dip 100 m further west.

The fault dip (and fault type) of the McGregor can not be directly determined because the actual fault slip surfaces are not exposed. From the mapped straight fault traces and the large vertical displacements involved, very steep dips are suggested with either normal or high angle reverse faulting indicated. The Hathorn #1 well ("spring") in Saratoga Springs is a deep logged boring (1006 ft.) on the downthrown (east) side of the main fault and constrains the dip of an east-dipping fault plane to an angle greater than 74°, because it does not pass through the main fault zone. Hills (1965) interpreted the McGregor fault to be normal based on the orientation of minor normal faults in the Glens Falls area which were steeply dipping and parallel to traces of the major fault. Cushing & Ruedemann (1914) include the McGregor-Saratoga Fault in their discussion of the Mohawk Valley faults and conclude that all are nearly vertical normal faults, and Fisher (1980) also mapped similar northerly trending faults (west of the McGregor) as normal faults. Bradley & Kidd (1991) also conclude the McGregor is a normal fault based on the strike and dip of about 010° and 70° E (respectively) of minor faults in

Proterozoic basement rocks just west of the fault at Kings Station, and the dip of 65° E of the Rock City Falls fault, a western splay of the McGregor as mapped by Fisher (1971). However, Geraghty and Isachsen (1981) examined the attitudes of subsidiary fractures near the major fault traces and found that a great majority were the result of proximate high-angle (greater than 60° E) reverse faulting (of the main fault). They caution that this conclusion is supported only to the extent that fault-related fractures are consistently parallel to the attitude of the major structure(see Firman, 1960).

The tendency to categorize the original movement on the McGregor fault system as normal, in unison with similarly oriented normal faults to the west, seems justified. However, the fault is an old one, and reactivations may overprint previous movement. Thus, reverse movement could be considered possible.

Throughout its length the Fault System branches in connection with horsts and graben, asymmetric half graben, and horst wedges. The Lake George Graben is a northern extension of the McGregor Fault, about 5 km north of Glens Falls (Fisher, 1985). Near Kings Station, 6 km north of Saratoga Springs, a branch fault, the Gurnspring fault (Cushing and Ruedemann, 1914), splays off toward the northeast, forming the lower boundary of the Wilton Horst block contained to the north by a probable east-west fault north of Wilton (Figure 1.3). About 2 km north of Saratoga Springs, the McGregor fault spawns dendritic branches (the Woodlawn Park fault and at least one unnamed fault, Figure 1.3) which splay toward the west, greatly diminishing the throw of the main fault along its southern extension, the Saratoga fault. Branches commonly occur at curves along the main fault, and the throw of the main fault at that point is then divided among the branches. The main fault zone width can increase to more than 300 m at intersections with branch faults, as near Kings Station, for example (Willems et al., 1983; Bosworth and Putman, 1986).

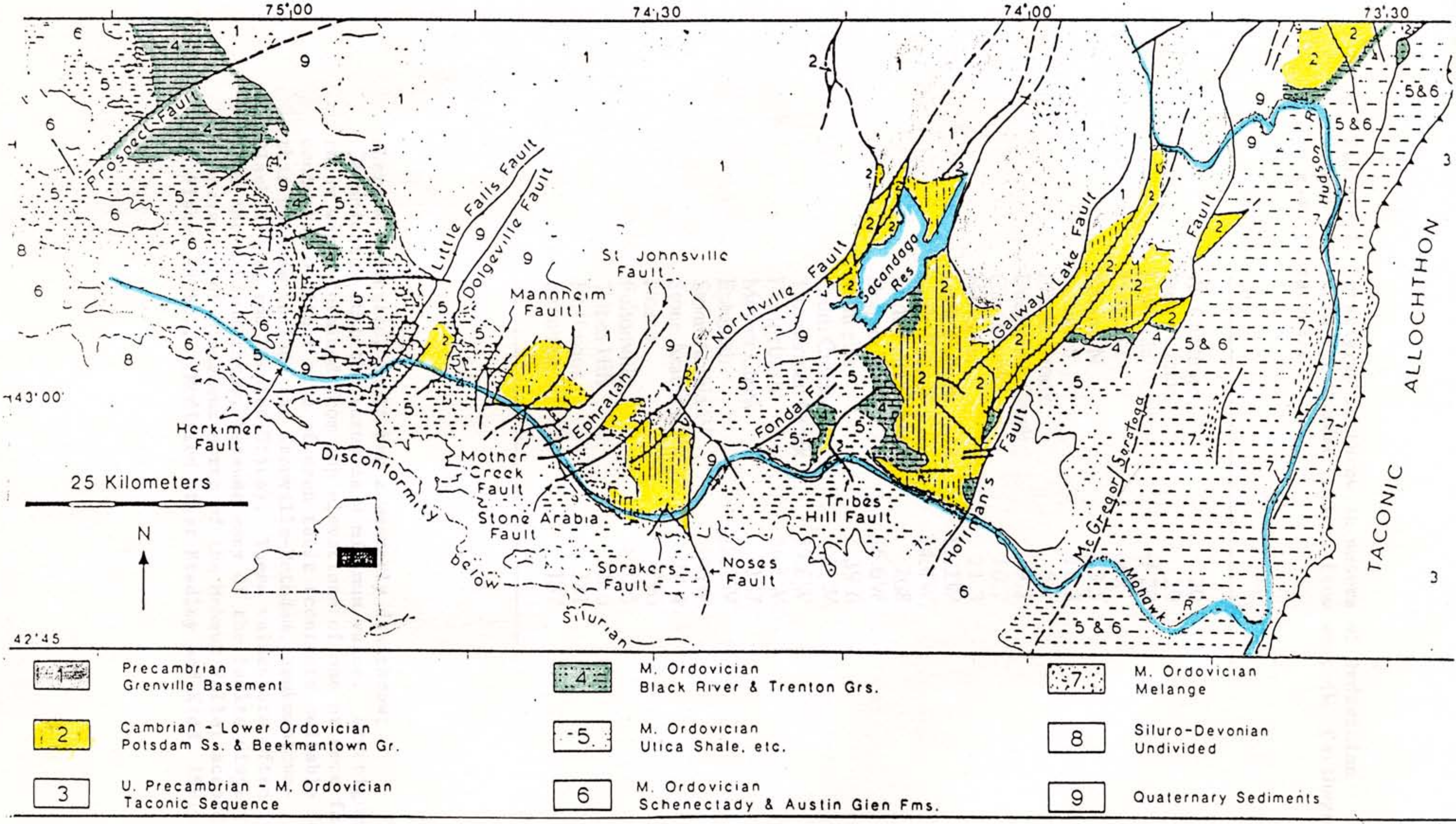
The Saratoga fault is renowned because of its association with the carbonated mineral springs in the city, however the stratigraphic displacement there is rather innocuous - 187 m (Geraghty & Isachsen, 1981) compared to that further north. On its course through the city there is commonly associated with it a narrow fault horse (up to 40 m wide) east of the main fault trace. The horse is a wedge of Black River Group (middle upper Ordovician) limestone caught up within the fault zone between upthrown lower Ordovician Gailor dolostone to the west and downthrown middle Ordovician Utica shale to the east (Cushing & Reudemann, 1914). Presently, this horse is no longer exposed, but can be located using bore hole log data and non-invasive geophysical techniques.

1.2 FAULT MOVEMENT HISTORY AND EVIDENCE FOR RECENT REACTIVATION

The time of the initial formation of the McGregor-Saratoga-Ballston Lake Fault System dates perhaps to Late Proterozoic episodes of block faulting in northern Adirondack rocks according to Fisher (1977). Mylonitic textures in quartzite fault breccia within the McGregor fault zone at Kings Corners were considered by Bosworth & Putman (1986) to represent material from a ductile shear zone over 12 km deep, brought to the surface by the faulting and uplift of Adirondack basement rocks. They infer that the "timing is consistent with rift development accompanying crustal extension and initial opening of the Iapetus Ocean" in late Proterozoic time. During subsequent uplift, continued faulting under shallower brittle conditions fractured the older ductile fault zone and later involved blocks of Ordovician dolostone. Since 12 km of uplift is inconsistent with the smaller displacement of the Proterozoic-Paleozoic unconformity along the McGregor fault, post-Taconic movement is unlikely to have been exclusively responsible for the uplift. Some North-northeast striking mafic dikes are within 1 km of the McGregor fault, and at least one is truncated by the Paleozoic-Proterozoic unconformity (Fisher, 1977; Fisher, 1984; Bosworth & Putman, 1986; Isachsen, et al., 1988).

It is likely that a reactivation of the fault system occurred during the Taconic Orogeny when, during the late Llandeilian epoch (Rowley and Kidd, 1981), the eastern North American craton began colliding with a microcontinent which contained a volcanic arc (Delano, et al., 1990). Unconformities on the North American carbonate shelf are associated with the development of typically eastward dipping normal faults (Zen, 1967; Rowley and Kidd, 1981), which are considered to have formed in response to loading and bending of the continental margin in the Taconic closing zone (Bradley and Kidd, 1991;; Bradley and Kusky, 1986; Figures 1.5 and 1.6). Figure 1.6 is a map of structures in the Taconic Foreland basin in the Mohawk Valley of New York, and Table 1.1 describes a general increase of normal fault displacements therein, progressing from west to east. The fact that the Ballston Lake Fault apparently hinges out within the Ordovician age Schenectady formation and that several Mohawk Valley faults are flanked by conglomerates in a manner consistent with synsedimentary faulting is suggestive that the Ballston Lake Fault (and perhaps the whole of the McGregor Fault System) was also active at that time.

The liklihood of any Devonian or later movement along the southern portion of the fault system is constrained by evidence from bore hole log information which indicates that some Mohawk Valley faults continue southward within Ordovician strata beneath the undisturbed Devonian strata of the Helderberg escarpment (Bradley & Kidd, 1991). However, the comparatively large, progressively increasing displacement of the McGregor fault toward the north suggests that it has been active during post Taconic time. The nature of this "selective" displacement makes it difficult to attribute the whole of the reactivation to the Acadian (Middle Paleozoic) or Alleghenian (Late Paleozoic) orogenies, or any extension that resulted from a relaxation that followed those collisions. It is interesting to note that to the north, some normal faults similarly oriented to the McGregor, which form part of the southeastern boundary of the Lake Champlain Graben are known to



—Geologic map of the Taconic foreland basin in New York showing the distribution of rock units and the locations of structures and localities mentioned in the text. Unit 5 includes the Canajoharie Shale: Unit 6 includes the Austin Glen Greywacke and the Frankfort Formation. Units 5 and 6 have been left undivided in the area of little outcrop and mild deformation to the east of the McGregor Fault. Unit 9 is only shown where the distribution of covered bedrock units cannot be reasonably inferred. Adapted from Fisher et al. (1970). Fisher (1980). and Bosworth and Vollmer (1981).

Figure 1.6 Map modified from Bradley and Kusky, 1986.

Table 1.1 Stratigraphic throw in meters of Ordovician Normal Faults in the Mohawk Valley, from west (W. Carthage) to east (Saratoga).

W. Carthage	20
Stony Creek	20
Lowville	+/-20
Prospect	-58
Poland N	33.5
Poland S	18.3
Shedd Brook	6.1
Buttermilk Creek	9.2
City Brook	10.7
Herkimer	21.3
Little Falls A	12.2W
Little Falls B	24.4W
Little Falls C (main)	268
Dolgeville	36.6W
Manheim	39.6
Crum Creek	61W
Timmerman Creek	39.6W
Kringsbush	36.6W
Mother Creek	137
Ephratah	≥162W
Stone Arabia W	146W
Stone Arabia E	24.4
Noses	296
Fultonville	≥30.5
Tribes Hill	68.6
Hoffmans	381
Saratoga	137

Where W indicated, fault has a westerly downthrow; > indicates that the throw is a minimum value. Stratigraphic throw was determined from map elevations of one or more facies contacts on either side of each fault (contacts suitable for this purpose include the Grenville-Potsdam, Beekmantown-Black River, and Trenton-Utica). These values are often less than the maximum throw, because many of the faults display an increase in displacement north of the Mohawk Valley according to Fisher (1980). (Modified after Bradley and Kidd, 1991).

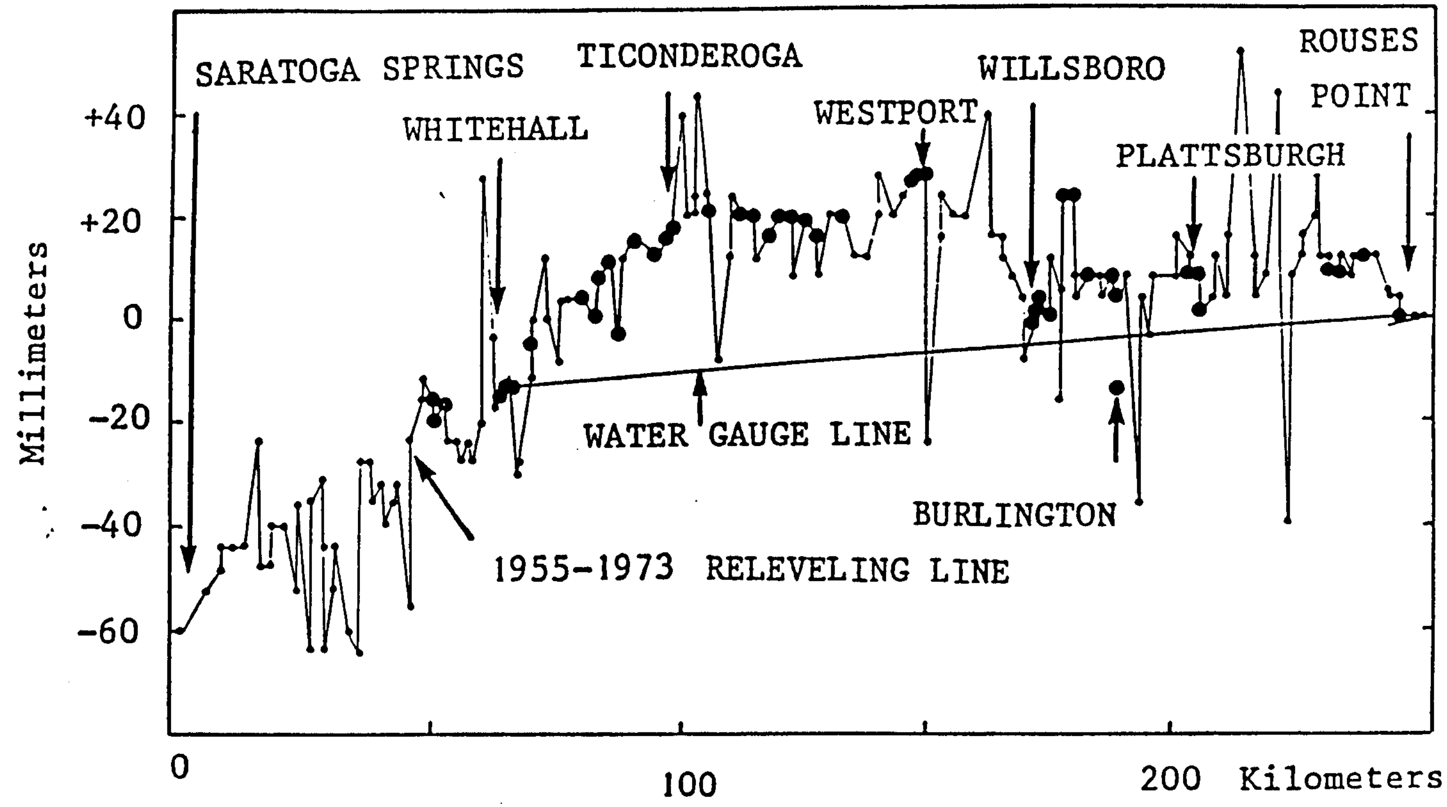
offset Taconic thrust faults, demonstrating that post Taconic movement has occurred on them. Examples include the Mettawee River Fault (Fisher, 1985) and the Warren Hollow and Poultney River Faults (Steinhardt, 1983).

However, several lines of indirect evidence suggest that the McGregor fault escarpment may be a result of Neogene movement, rather than simply a fault-line erosional scarp. For example, both outcrop constraints and shallow seismic refraction profiles (Knapp, 1978) demonstrate that the fault location coincides with the break in slope at the base of the escarpment. Also, slivers of easily eroded Utica shale in the downthrown fault block have been drag folded upward and are exposed against the escarpment face (at *, Figure 1.3), suggesting that they were dragged into their present position recently, with little time to allow for removal and/or burial or for any significant erosional retreat of the scarp.

1.3 NEOGENE ADIRONDACK UPLIFT AND POSSIBLE FAULT ACCOMODATION

The Adirondack Mountain massif is an anomalous feature of the eastern North American craton, considering its confined dimensions with respect to its stark, abrupt elevation. Quite distant from a plate or continental margin, this enigmatic uplift is identical to those associated with hot spots but with no associated vulcanism (Delong, et al., 1975). It is an elliptical domical uplift 1600 m in elevation, with about a 200 km long axis trending about N17°E, and a short axis length of about 140 km (Isachsen, 1975). The breached dome exposes 1.1 B.Y. Proterozoic basement rocks of the Grenville Province of the Canadian Shield, to which it is connected toward the northwest by the Frontenac Arch. Toward the east it is bordered by the Champlain trough, and its southern and southeastern margins are extensively fractured with block-faulted graben containing adjacent Paleozoic rocks (Fisher et al., 1971).

Although composed of Proterozoic rocks, the Adirondack mountains are not Proterozoic features. This is noted by the fact that onlapping Cambrian and Ordovician strata around the perimeter of the dome dip away from it (Isachsen, 1966; Fisher et al., 1980). Releveling surveys done along the eastern margin of the Adirondacks have shown an uplift rate of 3.7 mm/yr near the dome center (Isachsen, 1976; 1978) and 2.2 mm/yr along the eastern margin at the same latitude (Isachsen, 1975; Figure 1.7). The same technique has been applied to determine a rate of uplift of the Swiss Alps at 1 mm/yr (Schaer and Jeanrichard, 1974). The domical configuration of the area undergoing uplift combined with



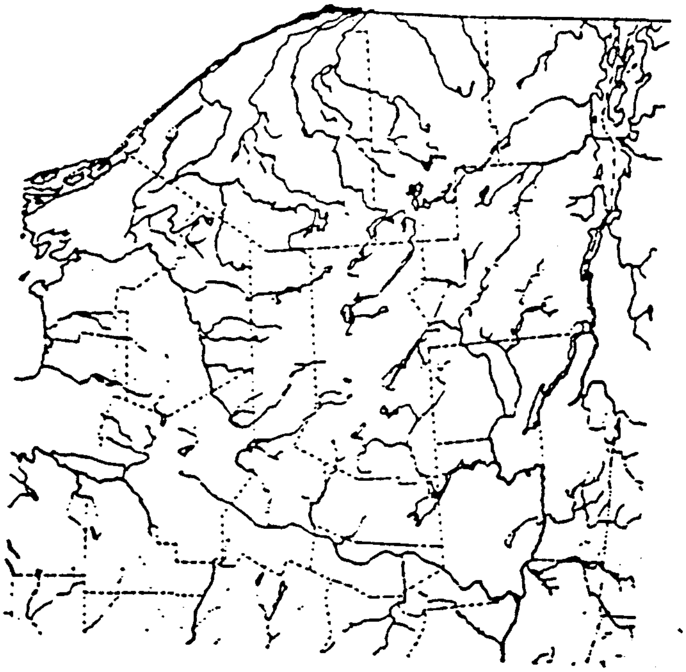
The 1955-1973 releveling line "calibrated" to the water level gauge profile line for the same period. The releveling line has been rotated so that the Whitehall-Rouses Point slope corresponds to that of the water gauge profile.

Figure 1.7 From Barnett and Isachsen, 1980 subsidence along the northeastern portion argues against a glacioisostatic mechanism of origin (Isachsen, 1981). Also, ancient Champlain Sea water levels in the Champlain Valley indicate that glacioisostatic rebound for that area was essentially completed 10,000 yrs. ago (Wagner, 1972; Cronin, 1977). Although there exists some controversy regarding the reliability of releveling techniques (Mark, et al., 1981), support for the results can be garnered from the consistency of maximum resolved uplift rates near the center of the dome and the correlation of that data with water level gauge profiles demonstrating tilt and differential displacement of the Lake Champlain basin (Barnett and Isachsen, 1980); Figure 1.7).

The topographical similarity of the Adirondacks to that of hot spot localities has not gone unnoticed. Burke & Whiteman (1973) suggested that the majority of African hot spot populations on the African continent are all about 1 km high and 100-200 km across. Also, while there does exist a range in the diameters and amplitudes of the uplifts, and in the amount of volcanism, there is no apparent correlation because large and/or high structural uplifts are not necessarily associated with much, or any, volcanic accumulation (Burke, et al., 1979). Typical examples of circular high spots (hot spot type uplifts-without any volcanic component) in continental environments include the Adirondacks; the Putorana massif in Siberia; and the Serra do Mar in Brazil, as well as several in Africa (Fouta Djallon, Angola, Adamawa, and High Veldt; McGetchin, et al., 1979).

Other types of anomalous intraplate topography should not be confused with high spots. They include old (up to about 500 m.y.); very elongate collisional orogenic belts such as the Appalachians (the present day Appalachian Mountains are a reactivated uplift zone from Maine to Georgia - the Adirondacks could be a part of it; Putman, personal communication), and forebulges due to lithospheric flexure in front of active thrusting zones (eg. central India). It has been shown that the African plate has a larger quantity of hot spots than any other plate (Burke & Kidd, 1975; Burke & Wilson, 1976; Thiessen, et al., 1979) and the stationary behavior of the African continent over the convecting mantle since about 30 m.y. ago (Burke & Wilson, 1972; Thiessen, et al., 1979) was likely to have been an important factor in trapping and concentrating sub-lithospheric heat sources (Briden and Gass, 1974). Lithospheric thinning is the most plausible (and perhaps exclusive) explanation for the uplifts associated with hot spots and high spots, as ascending convection currents in the mantle lead to pressure-release melting and the resulting magmas penetrate the plates (Morgan, 1972). Lesotho, South Africa, was a high-spot (elev. 3,800+ feet) up until 1983, when for the first time it extruded volcanics (Kidd, personal communication). It should be noted, however, that although typical of high spot/hot spot topography, a thermal anomaly has not been observed to be associated with the Adirondack uplift (Sneeringer, et al., 1983).

Other evidence in support of the suggestion that the Adirondack dome is young and rising include the fact that the stream drainage pattern is radial (Figure 1.8), has poorly accomodated the variously resistant Proterozoic rock types (marbles, quartzites, gneisses, for example; Fisher, et al., 1971), and has not incised too deeply into the McGregor escarpment. Although the dome is an area of low and occasionally moderate magnitude seismic activity, a direct relationship



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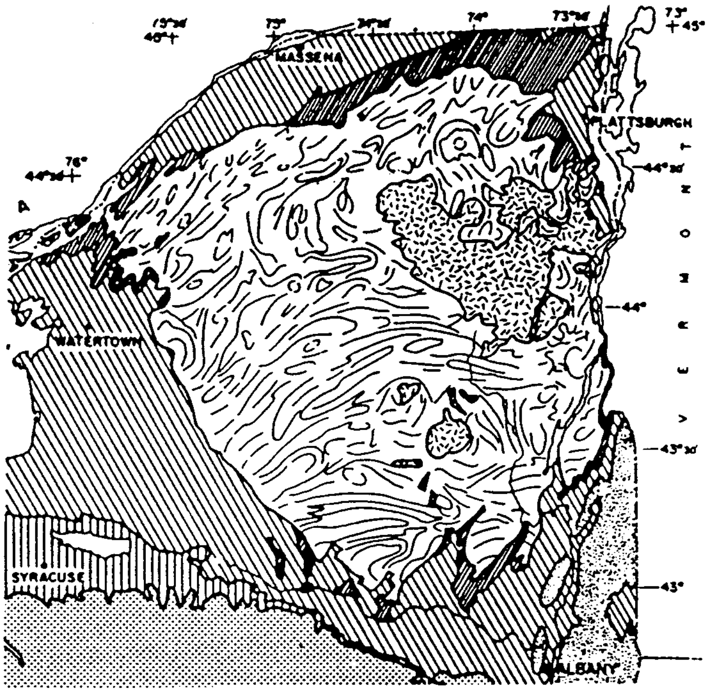


Figure 1.8A
Shows radial drainage pattern upon the Adirondack domical uplift, which has developed in spite of various influences of structural trends and metamorphic lithologies.
Figure 1.8B

From Isachsen, Geraghty and Wiener, 1981

B

between the uplift and the present pattern of seismicity has not been established. However, faults that are accomodating the uplift must exist. At Wilmington Notch (near Lake Placid), there is a fault which contains fresh sandy fault breccia and gouge (scarcely weathered angular fragments), with slickensides. Activity on such faults is certainly possible. Speculation that the McGregor fault system may be active is not unjustified.

2.0 SOME EARTHQUAKE HISTORY OF SOUTH EASTERN CANADA AND NORTH EASTERN UNITED STATES AND THE FAULT MOVEMENT POTENTIAL IN THE REGIONAL STRESS FIELD

Though comparatively rare, the northeastern United States does contain zones of relatively high seismicity. New York and Massachusetts have experienced numerous quakes, several quite severe. The region is also affected by large earthquakes occurring in Canada, mainly in the St. Lawrence Valley and the Laurentian Trough. The crustal adjustments occur to depths of 25 km (Mitronovas, 1983).

Before the arrival of European settlers, Indians of the Mohawk Valley Iroquois League explained the earthquakes as a result of the shifting feet of The Great Turtle that supported the earth on its back (Dudley, 1727). Indian legends describe at least one drastic event in which an isolated tribe (the Tyadaghtons) was wholly swallowed up by the earth in a great tragedy in the vicinity of a tributary of the Susquehanna River (Eckert, 1978). More recent information on historical and instrumental earthquake locations is compiled in figure 2.1. The data suggests a broad NW trending zone of seismicity from northeastern New York to Kirkland Lake, Ontario; and a NNW trending zone from Boston, Massachusetts to central New Hampshire. There is also a small cluster of activity in the Attica-Dale area of western New York, and around Charlevoix on the lower St. Lawrence River.

The most thorough survey of seismic activity in the McGregor area was done by Mitronovas (1983). Some of that data (Table 2.1 and 2.2)

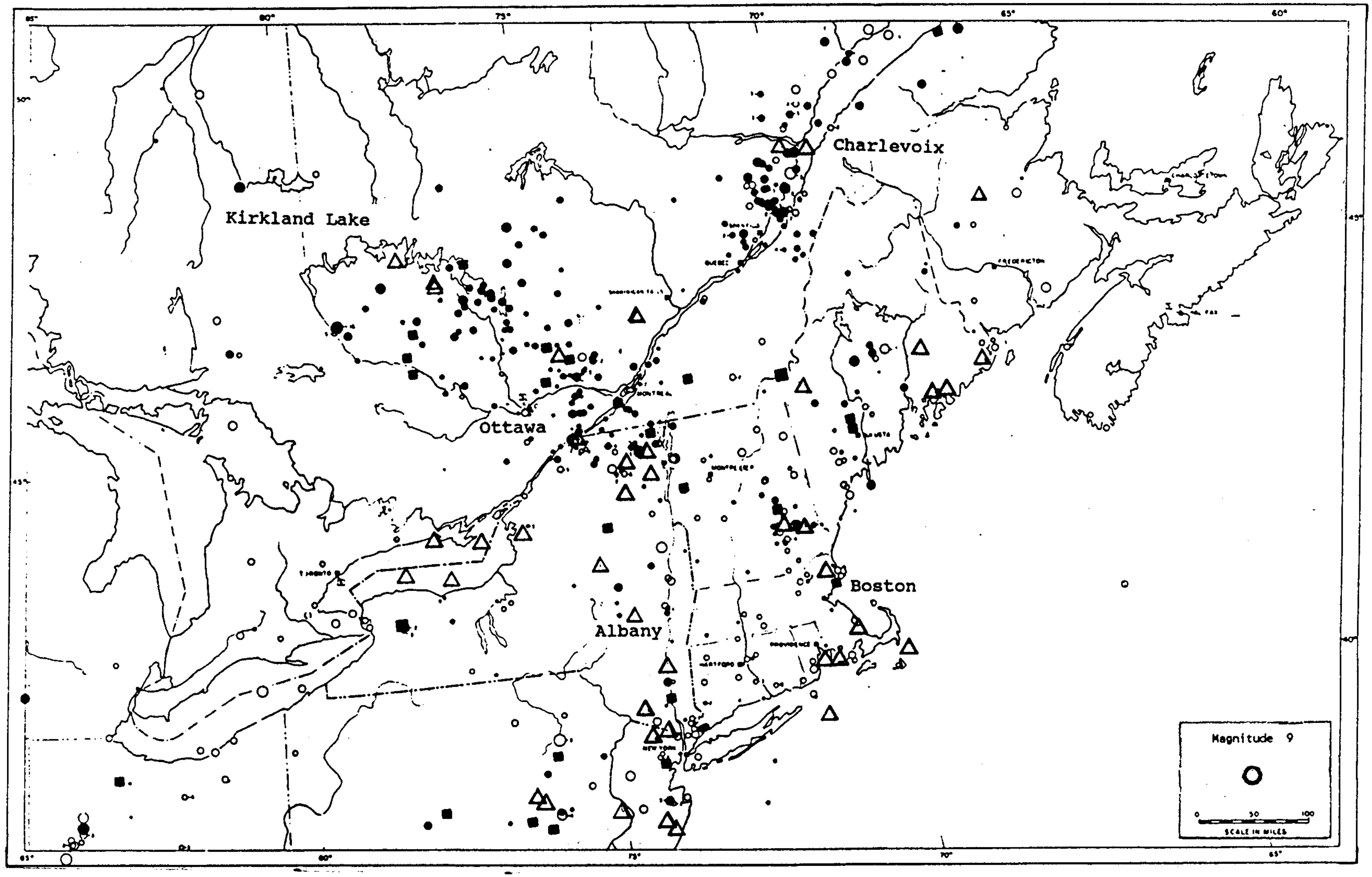


Figure 2.1 Epicenters from 1974 to 1980 of magnitude 3.0 or greater (triangles, from Nottis, 1983) are plotted on the map of Sbar and Sykes (1977; squares), USGS and NOAA epicenters from 1961 to 1972, plotted on the map of Smith (1966), epicenters from 1928 to 1959 (circles).

Table 2.1

Historical Earthquakes in the Mohawk Valley and Southern Adirondack

DATE	ORIGIN TIME (EST)	LOCALITY	LATITUDE DEG. (N)	LONGITUDE DEG. (W)	MAXIMUM[b] INTENSITY (I_o)	MAGNITUDE
July 6, 1775	10:55 -	Lake George	43.50	73.90	V	-
July 6, 1775	18:51 -	Lake George	43.50	73.90	IV	-
July 6, 1775	19:41 -	Lake George	43.50	73.90	IV	-
Jan. 11, 1847	23:30 -	Albany	42.70	73.80	III-IV	-
July 9, 1847	"AM"	Glens Falls	43.40	73.70	III	-
Dec. 17, 1855	14:00 -	Warren Co.	43.50	73.80	IV	-
May 11, 1877	10:02 -	Schoharie	42.60	74.40	IV	-
Dec. 28, 1878	21:32 -	Schoharie	42.70	74.30	III-IV	-
Mar. 18, 1881	21:30 -	Schenectady	42.80	73.90	IV	-
Apr. 2, 1882	06:30:07	Amsterdam	42.90	74.20	III	-
Apr. 2, 1882	08:10 -	Amsterdam	42.90	74.20	IV	-
Aug. 10, 1889	08:40 -	Warrensburg	43.50	73.80	V	-
May 25, 1890	17:10 -	Little Falls	43.00	74.80	V	-
Dec. 17, 1894	03:00 -	Coeymans	42.48	73.80	IV-V	-
Jan. 5, 1916	08:56 -	Chestertown	43.60	73.70	V	-
Feb. 2, 1916	23:25 -	Mohawk Valley	42.80	73.90	V-VI	-
Nov. 1, 1916	21:30 -	Glens Falls	43.40	73.60	V	-
Oct. 1, 1917	21:30 -	Glens Falls	43.30	73.70	III	-
Apr. 20, 1931	14:54 -	Warrensburg	43.50	73.80	VII	4.7
Oct. 29, 1933	-	St. Johnsville	43.00	74.40	IV	-
Nov. 1, 1935	01:30 -	Richmondville	42.60	74.60	II	-
Apr. 13, 1938	01:00 -		43.17	73.12	II	-
Oct. 21, 1939	03:59:33	Glens Falls	43.30	73.30	II	-
Oct. 25, 1939	09:46 -	Hudson	42.20	73.90	II	-
Apr. 11, 1940	20:58 -	St. Johnsville	42.80	74.60	II	-
Oct. 2, 1942	17:29:31	Albany	42.60	73.80	-	3.0
Aug. 24, 1952	19:07 -	Johnstown	43.00	74.50	V	-
Mar. 31, 1953	02:50 -		43.70	73.00	III	-
July 1, 1963	14:59:12	Albany	42.60	73.80	-	3.3
May 23, 1971	01:24:27	Blue Mountain Lake	43.90	74.48	V	4.1
May 23, 1971	04:29:59	Blue Mountain Lake	43.93	74.47	V	3.8
Jun. 20, 1971	21:48:31	Blue Mountain Lake	43.90	74.48	IV	3.4
July 10, 1971	03:15:01	Blue Mountain Lake	43.91	47.44	V	3.6
Dec. 20, 1971	06:44 -	Blue Mountain Lake	43.90	74.60	-	2.0
Mar. 15, 1972	07:10 -	Old Forge	43.70	74.70	-	2.6
June 16, 1972	04:01:58	Schenectady	42.80	73.90	-	2.0
Mar. 6, 1973	23:05:56	Blue Mountain Lake	43.81	74.45	-	1.9
Mar. 21, 1973	22:10:30	Blue Mountain Lake	43.89	74.42	-	-
June 11, 1973	05:08:31	Newcomb	43.95	73.98	-	-
July 15, 1973	03:30 -	Blue Mountain Lake	43.90	74.40	-	3.4
July 15, 1973	05:32 -	Blue Mountain Lake	43.90	74.40	-	-
Oct. 21, 1973	04:24:45	Blue Mountain Lake	43.82	74.45	-	2.2
Sep. 11, 1974	15:54:13	Blue Mountain Lake	43.83	74.19	-	2.2
Sep. 15, 1974	90:01:17	Schroon Lake	43.89	43.92	-	1.7
Sep. 18, 1974	01:23:09	Lake George	43.40	78.80	-	2.5
Nov. 19, 1974	04:23:28	Stony Creek	43.50	74.00	-	2.3
Jan. 27, 1975	05:40:10	Near Vt. Border	43.78	73.36	-	1.7
June 22, 1975	06:30:26	Old Forge	43.65	74.99	-	1.9
Aug. 3, 1975	23:58:22	Blue Mountain Lake	43.87	74.15	-	2.1
Sep. 11, 1975	11:54:58	Raquette Lake	43.89	74.66	-	2.6
Sep. 23, 1975	10:11:16	Raquette Lake	43.90	74.65	-	1.6
Nov. 3, 1975	15:54:55	Raquette Lake	43.87	74.64	-	3.9
Nov. 3, 1975	16:06:40	Raquette Lake	43.89	74.65	-	-
Nov. 5, 1975	06:53:11	Raquette Lake	43.89	74.65	-	-
Nov. 17, 1975	17:36:41	Raquette Lake	43.90	74.64	-	-
Nov. 18, 1975	16:00:13	Raquette Lake	43.89	74.64	-	-
May 9, 1976	08:56:03	Raquette Lake	43.88	74.65	-	1.0
May 9, 1976	22:19 -	Raquette Lake	43.88	74.65	-	-
Aug. 19, 1976	19:47:53	Raquette Lake	43.89	74.64	-	-
Aug. 21, 1976	09:00:12	Raquette Lake	43.89	74.66	-	-
Sep. 17, 1976	20:15:23	Blue Mountain Lake	43.82	74.20	-	1.6
Nov. 5, 1976	13:07:48	Raquette Lake	43.79	74.59	-	-
Apr. 5, 1978	09:45:49	Newcomb	43.85	74.24	-	2.6
Nov. 7, 1979	20:13:16	Blue Mountain Lake	43.88	74.50	-	-
Nov. 28, 1979	21:58:44	Blue Mountain Lake	43.79	74.49	-	2.5

b. MM-Modified Mercalli Intensity Scale

From Mitronovas, 1983

Table 2.2

Earthquakes Detected Near Glens Falls
(May 1979 to September 1980)

DATE	ORIGIN TIME (EST)	LOCALITY	LATITUDE DEG.(N)	LONGITUDE DEG.(W)	DEPTH (km)	M_C
01/04/80*/	15:27:35.14	Whitehall	43.590	73.299	0.0	1.6
03/11/80	06:08:46.02	Athol	43.511	73.795	0.0	1.8
04/24/80	13:31:04.03	Lebanon Springs	42.499	73.425	0.0	1.7
05/05/80a/	07:18:54.42	Near Glens Falls	43.314	73.652	1.8	-0.2
05/05/80	16:11:13.27	Lake George	43.437	73.645	0.0	2.2
05/12/80a/	13:30:18.36	Near Glens Falls	43.328	73.650	5.6	0.2
05/13/80a/	07:36:13.37	Near Glens Falls	43.317	73.655	1.8	-0.2
05/13/80a/	13:35:37.62	Near Glens Falls	43.346	73.689	4.6	1.1
05/14/80a/	07:11:33.98	Near Glens Falls	43.314	73.660	0.5	-0.2
05/15/80a/	07:03:36.24	Near Glens Falls	43.320	73.650	2.1	-0.3
05/22/80a/	05:17:49.47	Near Glens Falls	43.347	73.683	4.7	0.5
05/29/80a/	14:04:58.08	Menands	42.698	73.709	0.0	1.5
06/06/80	12:40:07.67	New Lebanon	42.466	73.440	0.0	1.8
06/24/80	13:46:28.22	Halfmoon	42.835	73.714	0.0	1.0

*/Uncertain whether man-made or natural

a/Earthquake sequence of 1980 near Glens Falls.

From Mitronovas, 1983

does suggest that the Glens Falls area, along with the Adirondacks and points north, tend to be more active seismically than the surrounding regions. While the Adirondacks are seismically active, an association between individual earthquakes and specific faults remains difficult to determine given typical errors in earthquake location analyses ($\geqslant$ 1 km in depth, 1/2 km in epicenter), and the fact that commonly neither the dip of a fault at the surface nor its change with depth are known.

Experience with earthquake activity in the western U.S. predisposes an attempt to try to relate the observed spatial trends of seismicity to the location of existing geological features, such as mapped faults and topographic lineaments. Adams (1989) suggests that almost all the significant earthquakes of southeastern Canada can be spatially and probably causally associated with the pre-existing faults of a Paleozoic aged rift system along the St. Lawrence and Ottawa Rivers (Figure 2.2). However, despite the continuity of these rift systems, seismic activity along them is spatially sporadic, albeit sometimes very intense. Coppersmith et al., (1987) found that 71% of magnitude 5 or greater (and all magnitude 7 or greater) historical earthquakes in eastern North America were spatially associated with ancient continental rift zones or passive margins (Figure 2.3). The thick Paleozoic sediments under the Lake Champlain and Lake George graben have led to the speculation that the faults bounding these ancient rift graben could be reactivated by the lakes that occupy them (Burke, 1977). There is a suggestion that a major tectonic zone exists from the New England Seamount chain through Boston to Ottawa (Sbar and Sykes, 1973; Fletcher et al., 1977), but the existence of historically low seismic activity in Vermont and western New Hampshire does not lend support. Sbar and Sykes (1977) propose that local

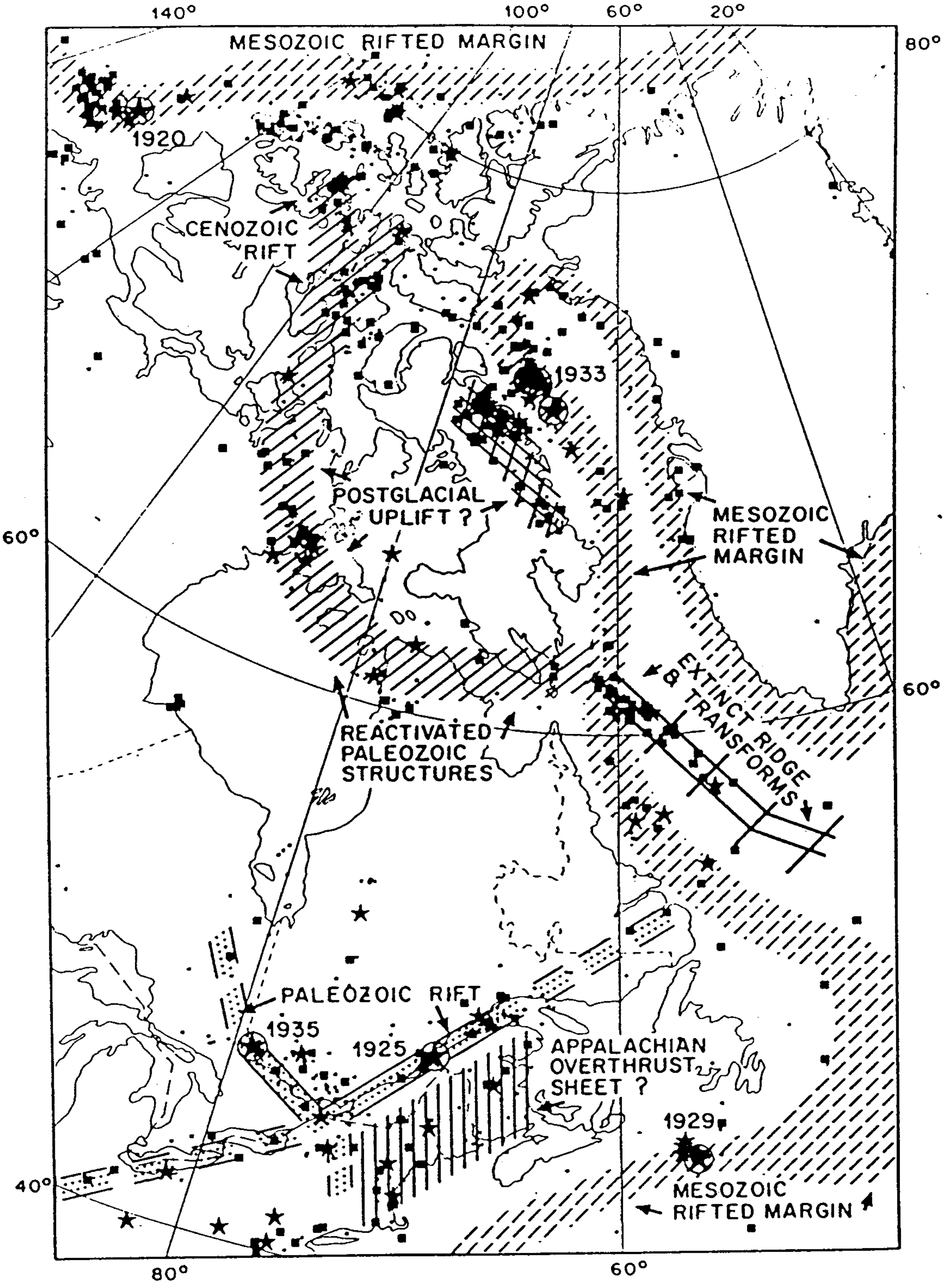


Figure 2.2 Earthquakes of eastern Canada together with an interpretative framework for the cause of the seismicity (From Adams, 1989).

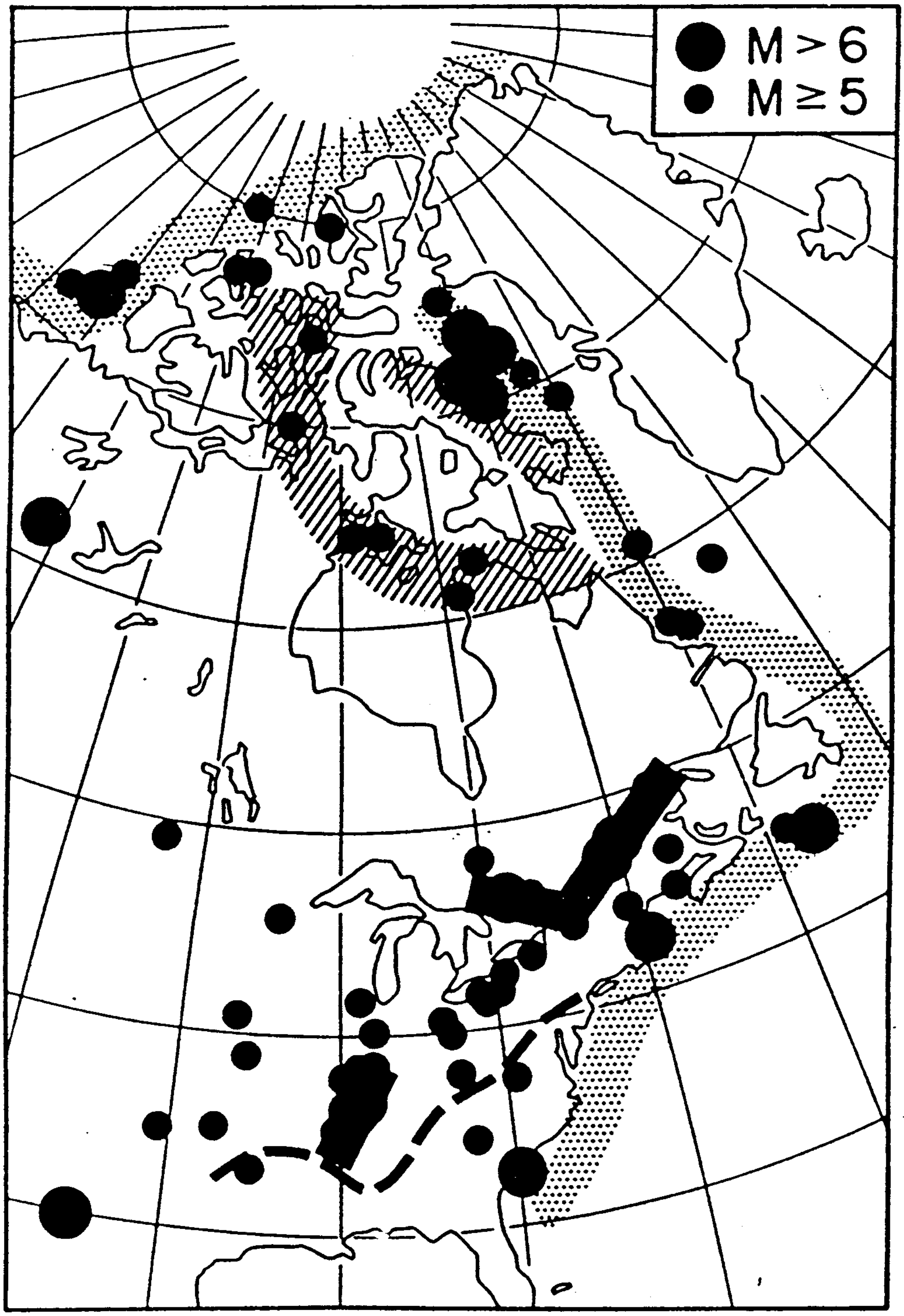


Figure 2.3 Schematic outline of Paleozoic rifting of the Precambrian craton (black); the buried edge of the craton in the southeastern U.S. (dashed line); Mesozoic rifting of the Arctic and Atlantic margins (stippled); and postglacial rebound reactivation of Paleozoic structures in the eastern Arctic (diagonal hatch marks). Superimposed on earthquake epicenters (From Basham, 1989).

variations in the stress field and the lack of suitably oriented pre-existing faults in Vermont may explain this. Linking earthquake activity to motions on a pre-existing "plane of crustal weakness", such as a fault, is an appealing tendency in earthquake investigation, for the reason that much has been learned of their relationships at plate boundaries. According to classic Andersonian faulting theory, frictional sliding is most likely to occur on those faults that are optimally oriented to the stress field (Jaeger and Cook, 1969). The "plane of crustal weakness" argument however, is based on the observation that faulting can occur on nearly any weak plane, even those striking nearly 90 degrees to the direction of the maximum principle stress direction, which is the case along much of the San Andreas fault system, for example (Zoback et al., 1987). But basic differences exist between the eastern and western U.S. in terms of the role of pre-existing faults in neotectonics. While intraplate seismic zones do show an association with zones of ancient tectonism, the mechanisms of intraplate seismogenesis are still poorly understood, and rarely has seismic activity in the east been (clearly) related to pre-existing structural features (such as faults or lineaments). Zoback et al., (1985) have speculated that discrete large-scale shear zones within the lower crust (which may play a role in continental fragmentation) may have a localized episodic role concentrating stress in the upper crust, analogous to strain accumulation and release along plate boundary faults like the San Andreas Fault.

Figure 2.4 shows a generalized stress orientation map of eastern U.S. and Canada (Zoback and Zoback, 1988). The data are from earthquake focal mechanisms and in situ stress measurements at depth, including stress-induced borehole enlargements ("breakouts") and hydraulic

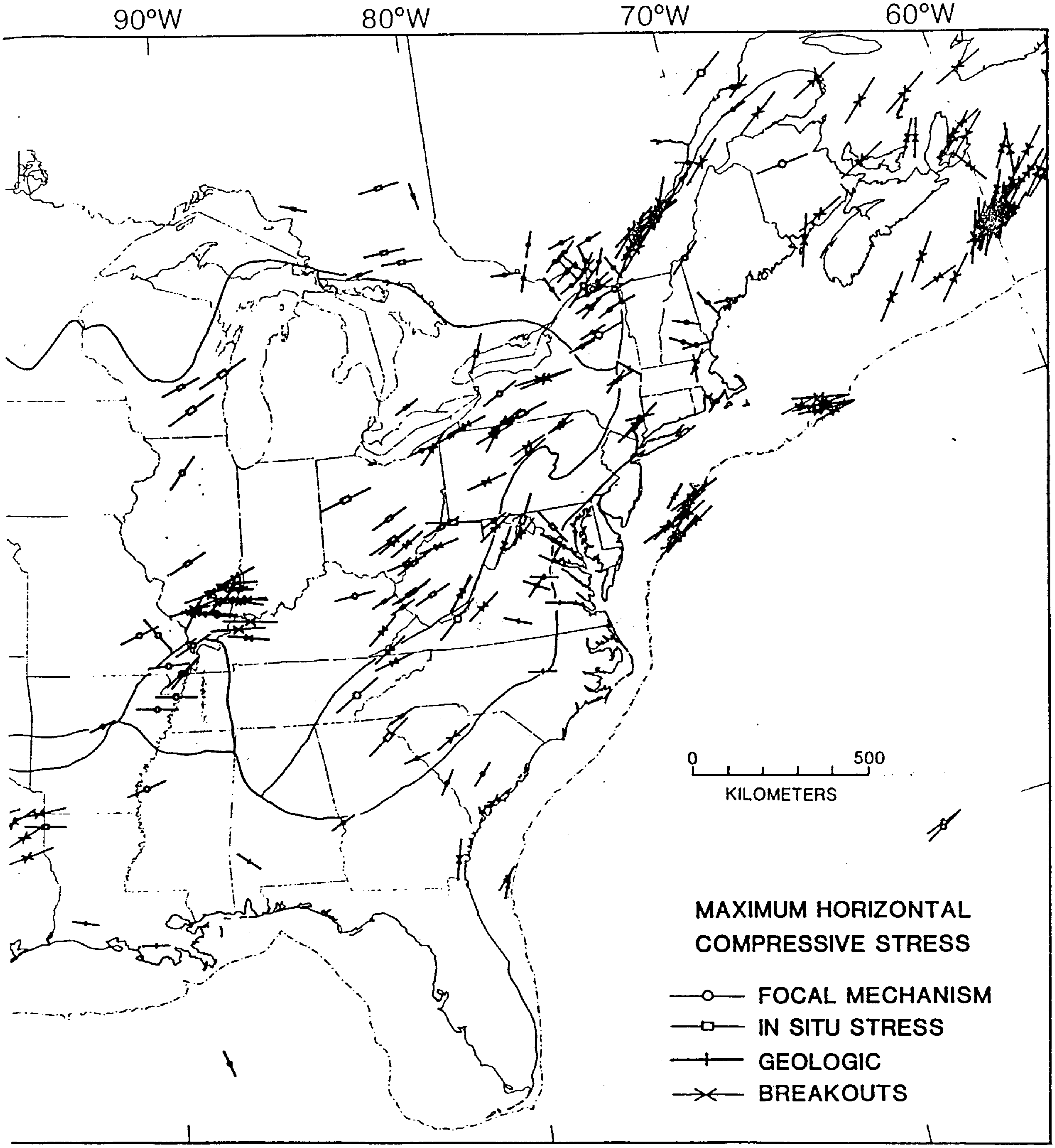


Figure 2.4 Current Tectonic Stress Field
(From Zoback and Zoback, 1988)

fracturing stress measurements. The large-scale direction of maximum horizontal compression is roughly ENE and appears to be of plate-wide tectonic origin, that is, consistent with the direction of absolute plate motion (Richardson et al., 1979; Woodward-Clyde, 1986; Zoback and Zoback, 1989). It has been suggested that areas where the local stress field orientation varies can be inferred to exist, based on collected earthquake focal mechanism solutions (Pulli and Toksoz, 1981). In New England, for example, earthquake focal mechanism solutions can locally show considerable variation in the orientations of the nodal planes and the P axes (Figure 2.5). Again, this could be explained as resulting from true deviations in stress orientation from area to area, or that slip is occurring on pre-existing planes of weakness in a relatively uniform stress field (or a combination of both explanations). Either way, it seems apparent that slip can occur on variously oriented planes.

For the most part, focal mechanism solutions in Northeastern America tend to describe thrusting (and reverse faulting) and some strike-slip motion indicating that the northeastern upper crustal region is largely under horizontal compression, and that the maximum compressive stress (σ_1) axis trends ENE, with nearly vertical minimum compression (Sbar and Sykes, 1977; Aggarwal and Yang, 1978; Yang and Aggarwal, 1981; Mitronovas, 1983; Gephart and Forsyth, 1985; Zoback and Zoback, 1988; in agreement with other stress indicator techniques). Earthquake foci of a swarm in the Blue Mountain Lake area of the southern Adirondacks (1971-72, M 1.9-4.1) define thrusting on a plane striking N12°W and dipping 25°E to 2 km depth, and a plane striking N31° E and dipping 59°E for events between 2-3.5 km deep (Sbar et al., 1972).

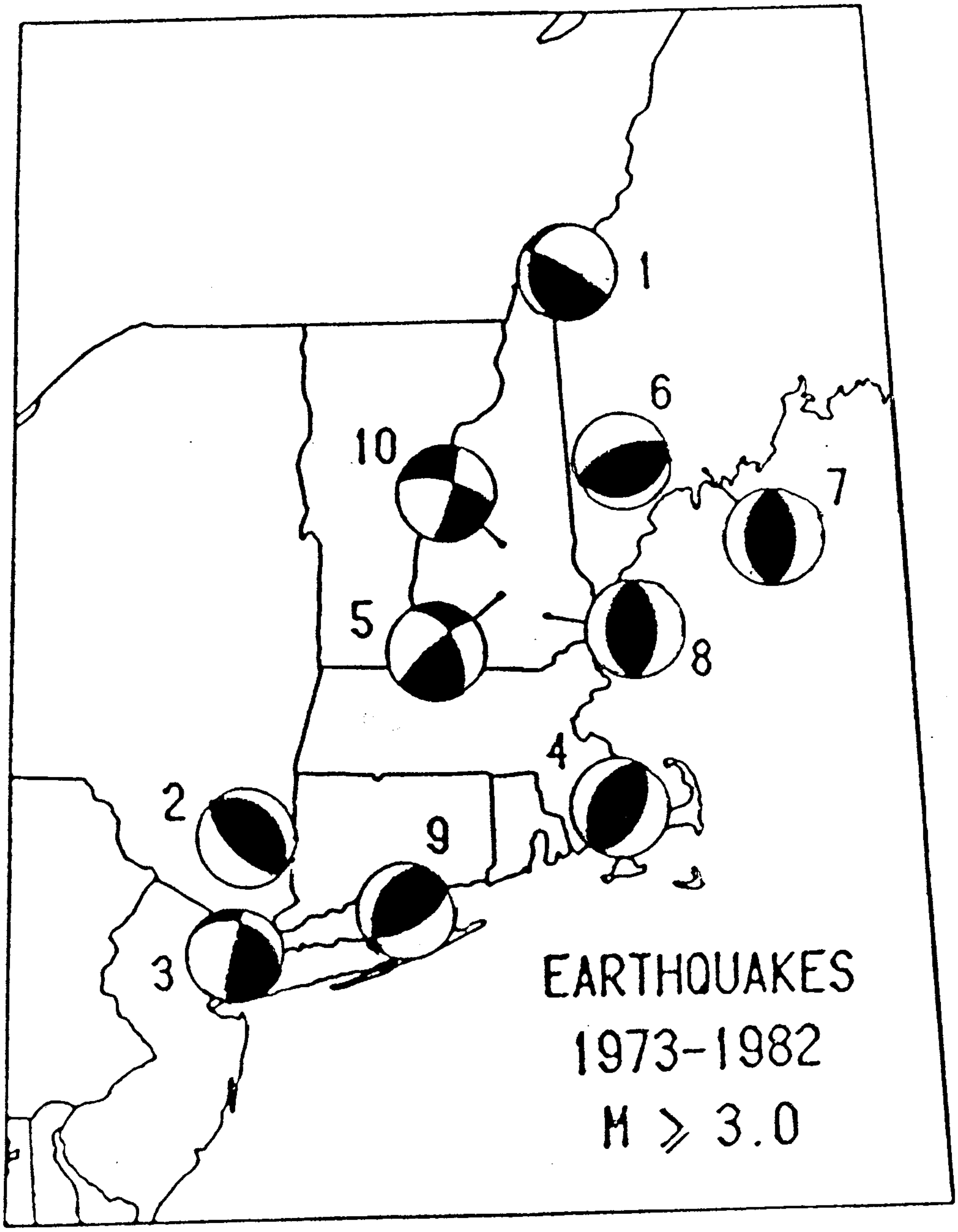


Figure 2.5 Focal mechanisms for ten recent earthquakes in New England and vicinity used for determining regional stresses. These are lower-hemisphere plots; shaded quadrants indicate compressional first arrivals. Data show considerable variation, except for mechanisms 7 and 8, which are identical (From Gephart and Forsyth, 1985).

Sbar et al., (1972) suggest that the deeper events may represent an extension of movement on a new shallow fault and its deflection down through an existing weakness, and conclude that the local principal stress is compressive, horizontal and trends nearly east-west.

The 1983 Goodnow earthquake in the central Adirondacks (M 5.3) occurred on a NNW striking fault dipping steeply to the west (Figure 2.6, Seeber and Armbruster; 1989). The surface extrapolation of this 8 km deep rupture roughly coincides with the location of the 15 km long Catlin Lake Lineament (Isachsen and McKendree; 1977), however, Grenville age markers have not been displaced more than a few tens of meters across the fracture zone along the Catlin Lake Lineament (Seeber and Armbruster; 1989). The Wilmington Notch Fault described earlier has been observed to hinge out into a zone of severely fractured rocks which posses no apparent offset (Putman, personal communication), resembling the zero displacement crackle zones described by Isachsen et al. (1981). This observation suggests that old (intraplate) faults may be the locus of moderate seismicity without actually displaying surface faulting effects. This is a recurring theme in the Northeast, and composite fault plane solutions have thus far only been able to infer movement on mapped faults (Sbar and Sykes, 1977). In many cases, resolved seismogenic faults strike at large angles to the trend of mapped structural features, (faults, lineaments) along the seismic zone (Johnston, et al, 1985; Seeber and Armbruster, 1989). Nonetheless, intraplate earthquake data can show that mapped features often control

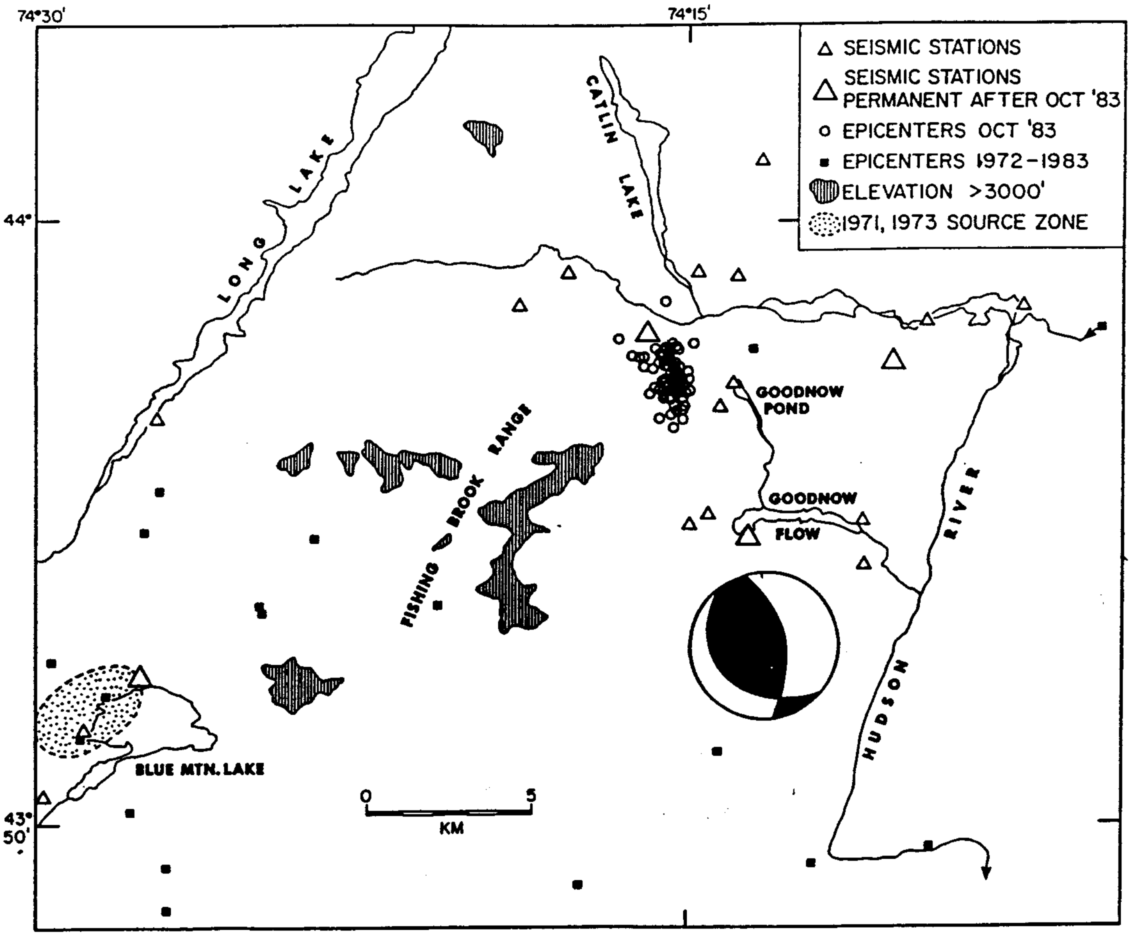


Figure 2.6 Blue Mountain Lake-Goodnow area of the Central Adirondacks. Three sets of epicenters are indicated: open circles = aftershocks of the 1983 Goodnow earthquake (October 7-29, 1983; data from the temporary seismic stations and fault plane solution shown); stippled area = epicenter zone of the 1972-73 Blue Mountain Lake swarms; solid squares = 1972-83 epicenters located by the regional seismic network. Long Lake, Catlin Lake and the Hudson River mark geomorphic features that reflect zones of fractures, possibly brittle faults. The north-northeast trend is dominant in the morpholory of the Central Adirondacks, but the north-northwest trend seems to reflect the seismogenic faults (From Seeber and Armbruster, 1989).

the spatial distribution of seismicity, in agreement with the reasoning that pre-existing faults are a zone of weakness (Ratcliffe, et al., 1986). Since the McGregor fault system is a likely zone of weakness, its propensity for failure in the regional stress regime must be addressed. Considering that the compressional stress field (at shallow depths) is oriented optimally to promote thrust and reverse faulting on planes striking NW-SE, the McGregor fault (striking NNE, on average) is not ideally situated for failure (Figure 2.7A).

However, an important issue that is poorly constrained at this time is the depth to which this reverse faulting stress field is acting. Zoback and Healy (1984) observe that shallow (upper 1 km or so) reverse faulting is common in compressional tectonic regions, but such stresses are not necessarily reflective of the stress field orientation at depth (Woodward-Clyde, 1986). For example; near the San Andreas Fault the near-surface (1 km) reverse faulting stress field like that shown in figure 2.7A becomes a strike slip stress field (Figure 2.7B) at greater depths (Zoback, et al., 1980). This could be a result of lithostatic loading at depth causing σ_2 and σ_3 to exchange their orientations (the vertical principle stress is intermediate in magnitude between the horizontal principal stresses). If this mechanism is occurring at depth in the Northeast, the greatest potential for seismic failure would theoretically favor steeply dipping fault planes striking NNE (such as the McGregor) or ENE in a strike-slip fashion. In South Africa a near-surface reverse faulting stress field has actually been shown to become an extensional (normal faulting) field below 2 km depth (McGarr and Gay, 1978). Whether or not a similar phenomenon could be occurring at depth in the Northeast is unknown.

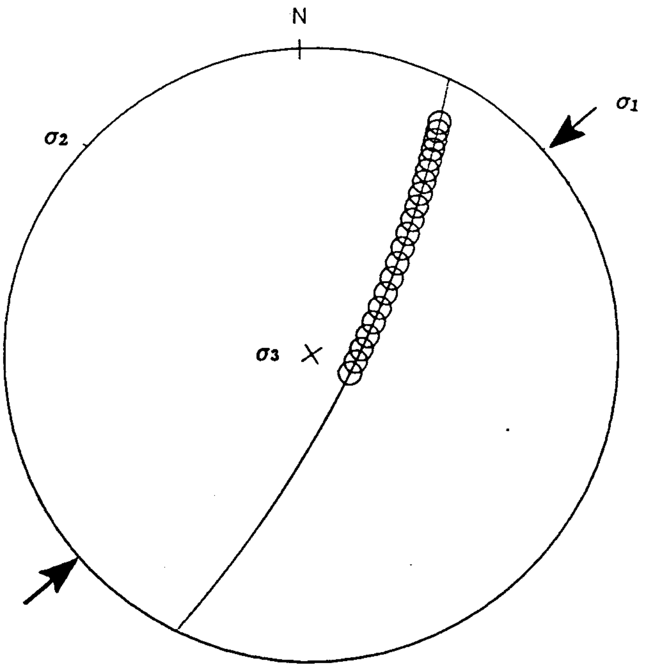


Figure 2.7A A lower hemispheric projection of the McGregor Fault with an average NNE strike and an assumed minimum dip of about 74° E. The small circles represent the ideal slip vectors which are possible given a failure of the fault in the shallow, compressional stress field given by σ_1, σ_2, and σ_3 (calculated after Bott,1959). A combination of reverse and strike-slip or reverse type faulting is possible.

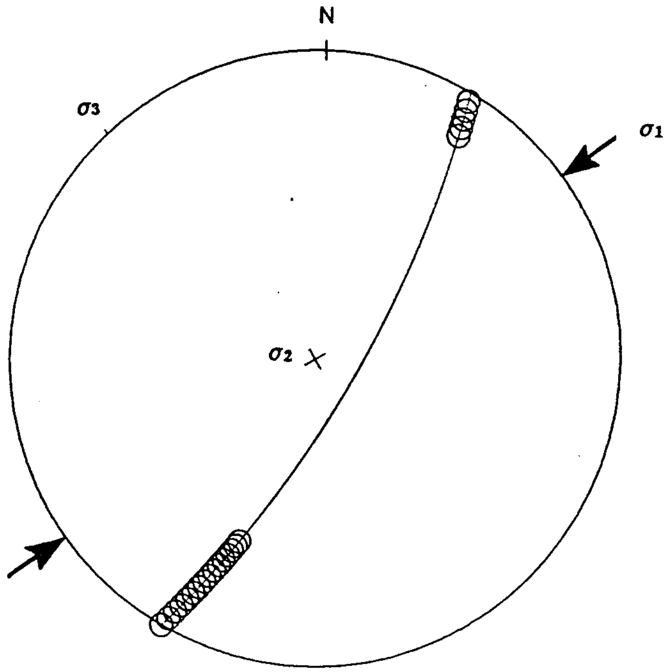


Figure 2.7B A theoretical reconfiguration of the principal stress axes at depth (greater than 1km) due to lithostatic loading. The suggestion is that strike-slip or a combination of strike-slip and reverse type movement potential may exist on the McGregor Fault.

The real problem encountered with a compressional regional stress field reactivation line of reasoning is reconciling it with the ongoing uplift of the Adirondack dome. Indirect evidence suggests that the ancient movement of the McGregor was normal - its fault plane dipping steeply to the east. If this is true, the rising Adirondacks west of it must be the result of extensional displacements, where σ_1 has some approximate vertical orientation. Therefore, one is forced to conclude that if the McGregor fault is accomodating the Adirondack uplift there must be a differently oriented local stress field responsible, perhaps it exists at depth, beyond the reach of bore hole measurements or even deep earthquake analyses. But while there is no evidence to suggest that focal mechanism solutions for earthquakes located in the Adirondacks consistently describe a differently oriented stress field for deep vs. shallow earthquakes, this is not too surprising given that the deeper events (greater than 8 km) are rare, typically not more than 20 km deep (still well within the crust), and the technology has limitations.

2.1 POSTGLACIAL FAULTING IN THE NEW YORK AREA

Evidence for recent (small scale) bedrock faulting dated by offset glacial striations can be found in outcrops of Paleozoic shales of the Hudson and Champlain Valleys and into Quebec (Adams, 1989; Oliver et al. 1970; Woodworth, 1907; Chalmers, 1897). They occur typically as groups of scarps and display similar features. Displacements usually range from 5-25 mm and nearly always lie in planes of cleavage. In nearly all of the occurrences, these scarp sets have their S or E side upthrown, striking between 9 and 70 degrees with very steep dips to the N or E where known. Groups of parallel scarps separated by a few meters can be found within a large outcrop so as to suggest that the local cumulative displacement may be significant (Oliver et al. 1970). Chalmers (1897) and Woodworth (1907), were the first to show that a pattern exists by scarps found in Paleozoic shale belts in the Hudson and Champlain valleys. The strikes of the scarps generally parallel the trend of the belt which has been traced down the St. Lawrence river valley as far as St. Georges (Oliver, et al. 1970). The absence of any preferred proximity of these scarps to slopes and the regional trend seem to argue against shallow causal mechanisms such as slumping or frost heaving (Woodworth, 1907; Oliver et al., 1970). Examples of these small scarps can be found in deformed Ordovician shales in the Mill Creek stream bed near 3rd Avenue in the City of Rensselaer, New York (Figures 2.8A and B). They are similar to the scarps described above except that they do not offset glacial striations. The suggestion that they may be recent (emplaced by faulting) is only supported by the observation that they have not been planed flat by stream erosion and are oriented (roughly N-S) in a direction which would not favour glacial plucking as a causal mechanism.

Figures 2.8 A and B. Small scale scarps in Ordovician shales in the Mill Creek stream bed, Rensselaer, which have not been planed flat by stream erosion. Discontinuous "offsets" up to 9cm (about -5cm on avg.) occur on anastomosing cleavages or non-planar shear cleavage surfaces (probably of Taconic origin). Looking approximately north, figure 2.8A shows two nearly parallel scarps from the upper left to the lower right corner of the photograph, where the distance between them is about 1m. Figure 2.8B shows a discontinuous scarp trending from the upper right to the lower left; width of field of view is about 8m. The scarps tend to trend between 330 and 355 degrees and have their SW side downthrown, dipping very steeply to the east where evident.

These scarp features are not exclusive to Paleozoic shales, and others of longer lateral extent have been observed in the Oriskany sandstone (Devonian) of the Helderberg Escarpment near Thompson's Lake road. Goldring (1933) described the offsets as "step faults" across glacial scratches, upthrown to the north. A similar fault, with a throw of about one foot, occurs in the Oriskany sandstone of the Onesquethau Creek, about two miles east of the Town of Clarksville, which has not been eroded by the stream (Ruedeman, 1928).

Adams (1989b) has observed many of the tiny (10-100 mm) bedrock scarplets in a broad arc from Western Ontario to Newfoundland, oriented tangentially to the margin of the ancient Laurentian ice sheet. He believes they represent the postglacial release of stresses caused by flexural deformation of the upper crust, partly because the individual outcrop offsets are consistent with stresses implied by the postglacial tilt of nearby lake shorelines.

A probable relative of these structures are pop-ups (Figure 2.9), linear brittle deformation uplifts of a meter or so (these include elongated domes which have been observed to form on quarry floors; Saull and Williams, 1974). It is unlikely these "ridges" of broken rock could have survived being overriden by the Wisconsin ice sheet, and Adams (1989b) suggests that they formed soon after deglaciation because the edges of the axial fissure is often as severely weathered as the surface of the bedrock itself.

The question of whether these small scale scarps have formed by faulting in response to glacial loading and/or unloading, tectonic

Figure 2.9 Pop-up in Potsdam sandstone near Alexandria Bay, New York.
(From Coates, 1975)

stresses, or some other effect remains unresolved (no known earthquake activity has been associated with their appearance). However, it is important to note that these structures represent undeniable evidence of actual Holocene deformation events.

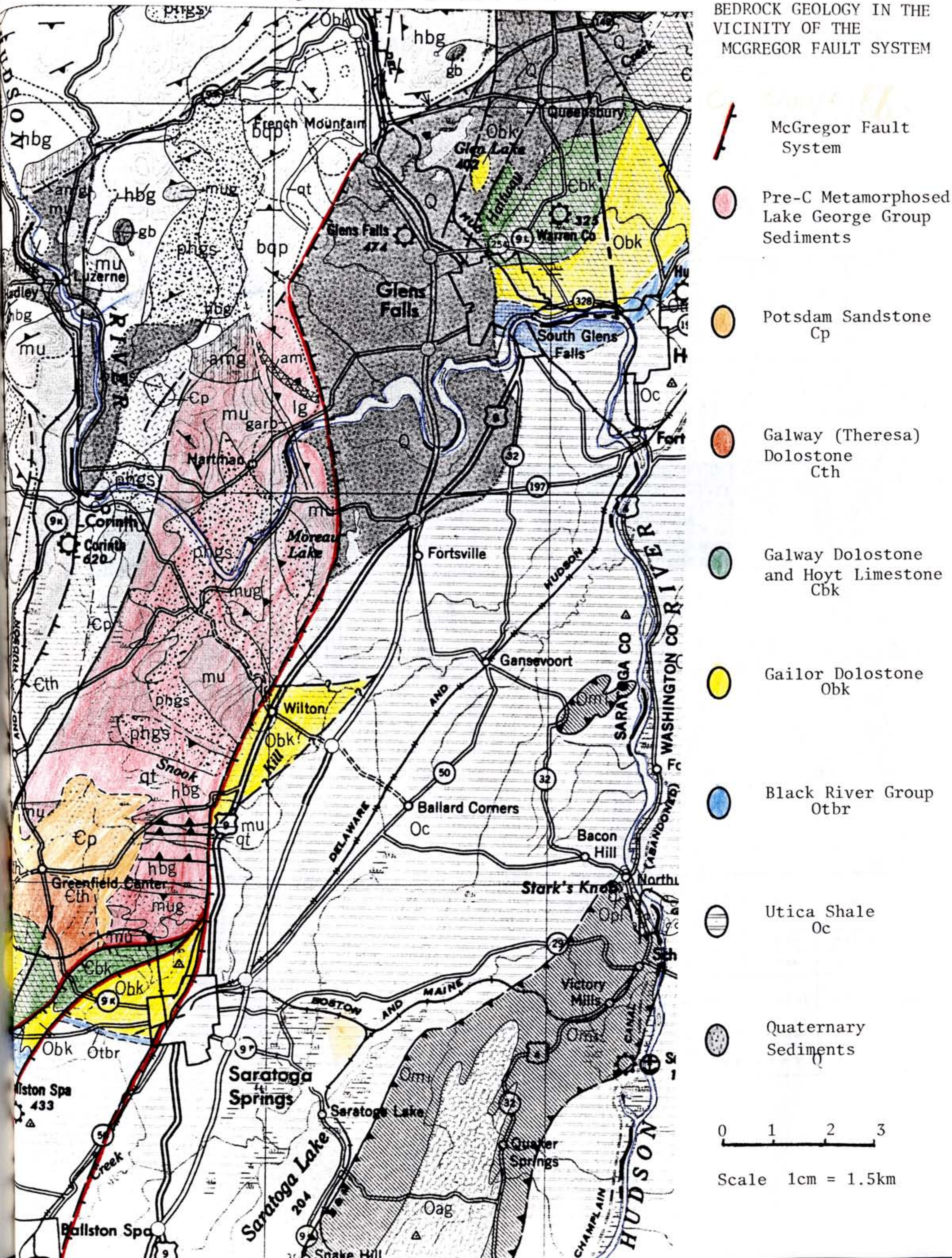
3.0 BEDROCK STRATIGRAPHY

An account of the stratigraphy of the areas that are associated with the McGregor-Saratoga-Ballston Lake fault system is prerequisite to understanding the age and nature of the fault relationships therein. The reader is referred to Figure 3.1, a portion of a map by Fisher, et al (1971), Figure 3.2, a collection of correlative stratigraphic sections of strata in New York (Fisher, 1984), and Figure 3.3, a generalized stratigraphic column of the Saratoga area. It should be noted that with the exception of the Wilton area, Paleozoic strata proximate to the fault system are only exposed south of the Saratoga Region, and the Southern Champlain Valley equivalents toward the north (Figure 3.2) are deeply buried under quaternary sediments. I have described the lithologies with greater detail than would seem warranted for fault control relations, and I emphasize that more lateral variation exists in these units than implied. Nevertheless, I found some value in the detail when reviewing bore hole log descriptions of bedrock known only from cuttings.

Precambrian Basement

The Precambrian Adirondack basement rocks are characterized by metamorphic textures and mineral assemblages. The rocks of the medium to high grade Adirondack 'core' (about 65 miles north of Saratoga Springs) are a mix of hornblende-(biotite) granitic gneiss, charnockitic gneiss, and meta-anorthosite, with some metasediments. Only the metasedimentary rocks have compositions diagnostic of specific protoliths, such as marble, calc-silicates, quartzites, and pelitic paragneisses. North of Saratoga Springs and west of the Glens Falls area, basement rock outcrops are mainly

Figure III-1 (Fisher, 19?1)



CORRELATION —

AUTOCHTHONOUS CAMBRIAN-ORDOVICIAN STRATA IN NEW YORK

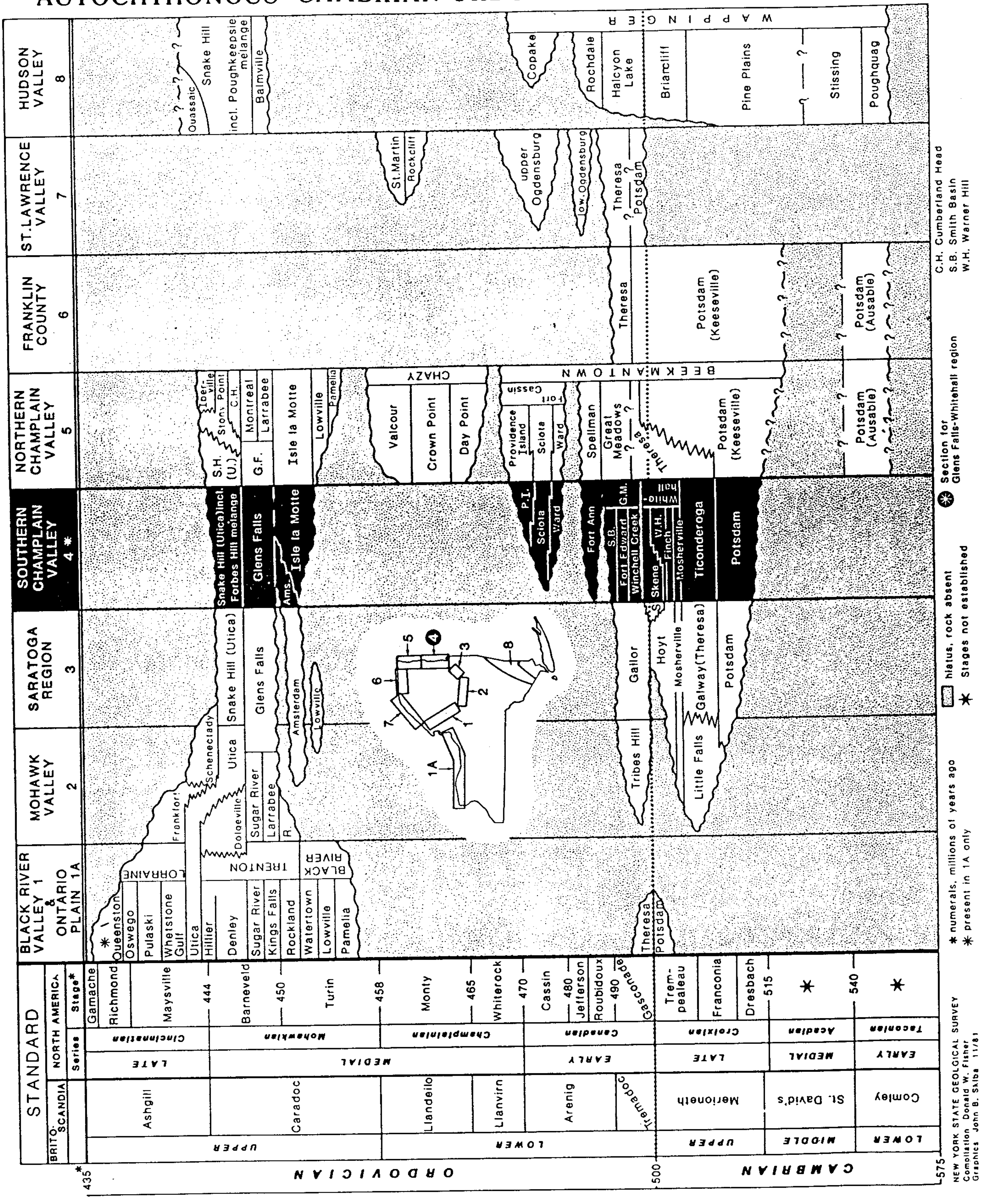


Figure 3.2 Fisher, 1984.

GENERALIZED STRATIGRAPHY OF THE SARATOGA AREA

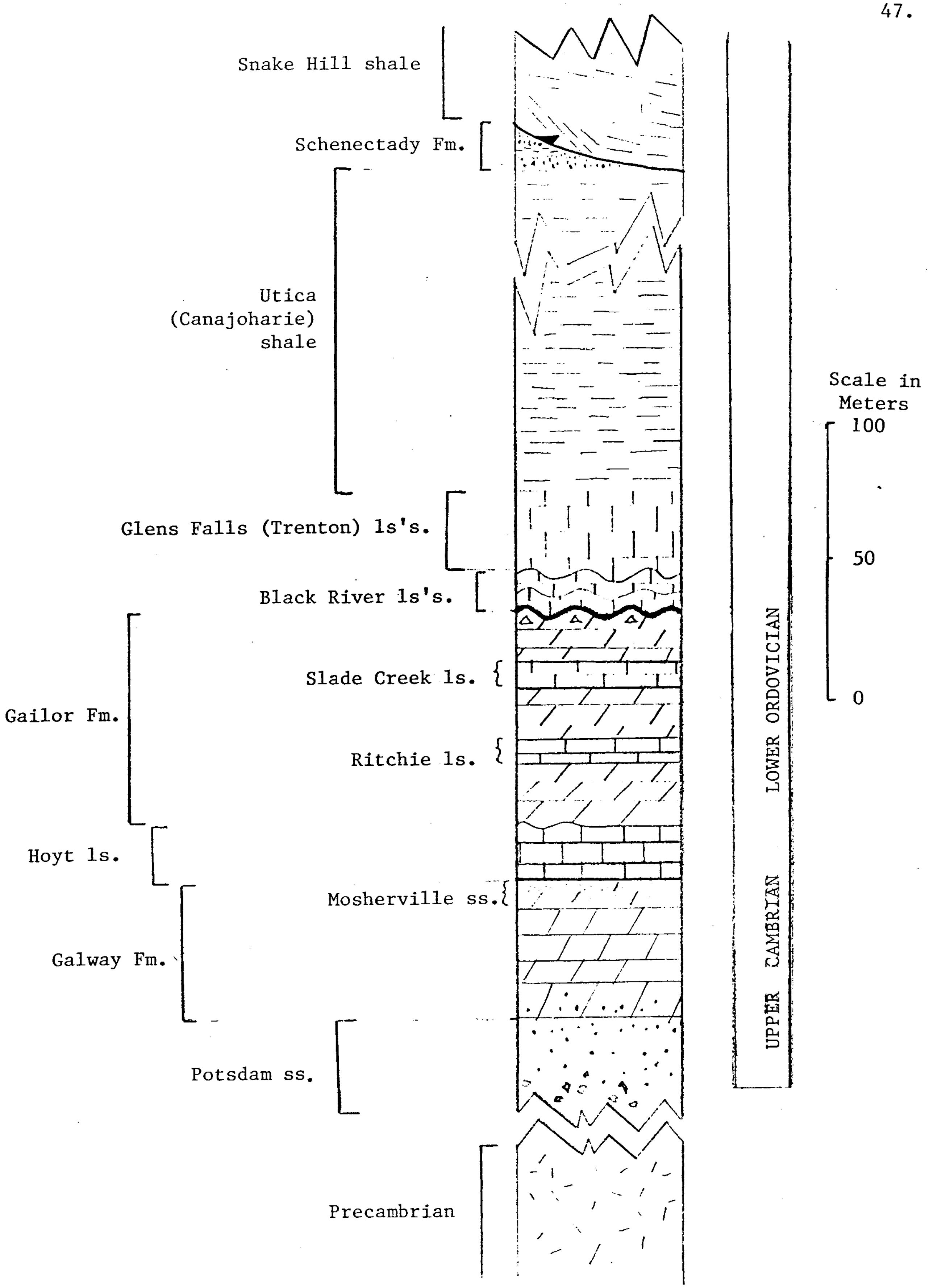


Figure 3.3 Modified from Mazzullo et. al., 1977 and Fisher, 1985

the metasedimentary rocks of the Lake George Group. These consist of quartzites, graphitic schists, garnet-sillimanite gneisses, and coarse calcitic marbles (often with graphite flakes).

The Lake George Group sediments accumulated in a shallow ocean more than 1,154 million years ago (Metzger, Bohlen, and Hanson, 1988), the age of their metamorphism during the Grenville Orogeny. During this orogeny, the rocks were folded, refolded, sheared, intruded by igneous rocks, and heated to about 750 °C at about 8 kilobars (Ree, 1991; Fisher, 1984; Putman and Sullivan, 1979).

For the rest of the Late Proterozoic, the terrane experienced diabase dike intrusions and high angle faulting as a result of the rifting event that formed the Iapetus Ocean. Little rock record exists for this interval, suggesting shallow sea levels and extensive erosion, and a prolonged uplift of the "basement rocks".

Potsdam Sandstone

Resting unconformably on the Precambrian basement, the Potsdam sandstone is a massive, bedded orthoquartzite composed of well rounded quartz grains cemented by quartz. Crossbedding is commonly well developed and the Potsdam is traditionally interpreted as a prograding beach sand. The bulk mineralogy is monotonous except in the basal portions (the lowermost 6 meters) where much of the sandstone is arkosic, and stream channel deposits are common at the lower contact.

The Potsdam surrounds and forms an on-lap facies to the Adirondack terrain, portions of which survived as low relief peninsulas and islands in Cambrian and Early Ordovician time (Mazzullo and Friedman, 1975). Though the Potsdam is largely Late Cambrian in age, lowermost units may be early Cambrian or even Pre-Cambrian (Fisher, 1977). In the Saratoga Springs area it outcrops west and northwest of the city and is approximately 30 meters thick (Fisher and Hanson, 1950; Cushing and Ruedemann, 1914).

Galway Formation

The sandstones of the Potsdam Formation are transitional upwards to the Galway Formation, a cherty, medium to thick bedded, dark to medium gray, medium to coarsly crystalline quartzofeldspathic dolostone. The lowermost portions are an alternating series of orthoquartzites very similar to the Potsdam sands below, and quartzose dolostones, in beds which average 3.0 meters in thickness. The quartz content of the dolostones and the thickness of the quartzite members decrease upward in the formation. Vugs with dolomite and calcite inclusions (usually white but occasionally rose colored) are found throughout the formation and minor pyrite (more rarely, sphalerite), is interspersed in some thin, darker, carbonaceous dolostone members.

The measured thickness of the Galway Formation in the Saratoga area is about 47 meters (Mazzullo, 1978) but is believed to increase rapidly toward the west to as much as 92 meters in the Galway area (Young, 1980). The top 7 meters of the formation consists of interbedded light gray, medium crystalline dolostone interspersed with four prominent white chert horizons which range between 3.5 cm to 37 cm in thickness. The cherts are a distinctive mottled variety known as novaculite, a very useful marker horizon in the area. At 7 meters below the top is a very vuggy zone filled with sparry calcite or dolomite, and occasionally, well terminated quartz crystals (Young, 1980). This zone may correlate with one further west in the Tribes Hill Formation which contains doubly terminated clear quartz crystals renowned as Herkimer "diamonds".

Though most of the formation is unfossiliferous, there are trilobite fragments, brachiopods and stromatolites in a thin (20 cm thick) shaly dolomitic sandstone layer at the type locality (on New York Route 9N, 3 miles west-northwest of Saratoga Springs) (Fisher and Hanson, 1951). Apparently originally deposited as lime muds and sands in a low energy subtidal to supratidal marine environment (Mazzullo, 1978), the Galway was subsequently pervasively dolomitized, wiping out much of the primary sedimentary and biogenic structure. Fisher and Hanson (1951) claim that faunal evidence supports a lower Franconian (upper Cambrian) through lower Trempealeau (uppermost Cambrian) age for the Galway.

Hoyt Limestone

In the Saratoga Springs area, the Hoyt limestone consists of 42 1/2 feet (Fisher, 1984) of medium bedded, dark-gray, fine to coarse textured limestone and minor dolostone with some quartz-sandstone layers. It is exceedingly fossiliferous compared to the other units in the area, with abundant cryptalgalaminites, biogenic oolites (Taylor and Halley, 1974), stromatolites, and some trilobites. Fisher and Hanson (1951) and Taylor and Halley (1974) claim the fauna are Trempealeauan. Locally there are ripple marked calcareous arenites, flat-pebble conglomerates, shallow channels, and small scale cross-beds. The well preserved Cryptozoon reefs are best exposed at Lester Park and the Petrified Gardens near the Pompa Quarry, west of Saratoga Springs. The Hoyt limestone is restricted in geographical extent due to truncation by an unconformity at its summit which wedges out to the west and/or structural isolation by repeated normal faulting in the area.

Owen (1973b) describes the paleoenvironment as largely intertidal or subtidal (pelletal limestones and oolitic packstones), and occasionally supratidal (fine grained dolostone lithologies). He continues,..."an epeiric-sea margin with northeast-trending oolite shoals...". "Lagoons behind the shoals provided local relief and a spectrum of energy conditions for (the) morphologic variation in the cryptalgal structures." Studies by Logan (1962) in Shark Bay, Western Australia demonstrate that living Cryptozoons thrive in an elongated bay with somewhat higher salinity than the open ocean. Adjacent land of low relief and slight rainfall supplies little detritus. The Hoyt facies was apparently similarly restricted and has not been found outside of the Saratoga area.

<u>Gailor Formation</u>

Unconformably overlying the Hoyt Limestone, the Gailor Formation consists of 77 meters of medium to thick bedded, light-medium gray colored, fine to medium crystalline, slightly cherty dolostone, which includes two prominent limestone members. The upper Slade Creek limestone has only been identified in core samples taken from the Pallette Quarry, Saratoga Springs, never in outcrop (Mazzullo, et al., 1978). It is virtually indistinguishable from the Ritchie limestone below, which Zenger (1971) suggests is also an intercalated member within the Gailor. The Gailor dolomite is very similar to the dolomites of the underlying Galway formation (though not always so coarsely crystalline) and distinguishing the two can be very difficult in areas of poor outcrop. Also, there can be rapid lateral facies changes to dark gray, silty and sandy dolostone within distances of only a few kilometers (Mazzullo, et al., 1978).

The Gailor Formation is the rock unit which is the primary host of the carbonated waters of Saratoga Springs. The intercalated limestones are more soluble in carbonic acid, and the waters are slowly enlarging fractures, bedding planes, and joints-cavities for mineral water storage in the dolostone as well. The Saratoga fault acts as a conduit that channels these waters closer to the surface, as at High Rock and Old Red "Springs".

Based on fossil samples, Fisher and Hanson (1951) considered the Gailor to be Lower Canadian and Zenger (1971) and Fisher and Mazzullo (1976) place the Ritchie limestone member in the Canadian. Mazzullo, et al., (1978) believes the lower Gailor may be Trempealeauan.

Mazzullo, et al., (1978) suggest that the depositional environment (of both the Galway and the Gailor) was very shallow based on "the abundance of reworked intraclasts in these rocks, as well as the rapid vertical transitions from marine to intertidal (and supratidal) lithofacies...". He continues, "The interbedded grainstones (sandy intrasparites) in these rocks may have been deposited in tidal channels or perhaps as submerged shoals and bars between the intermittently exposed tidal flats". Sand and silt sized detrital quartz and feldspar attest to the proximity of source rocks, the ancestral Adirondack Mountains (quartzites, gneisses) and/or reworked Potsdam sandstone.

Black River Group

Subsequent to the deposition of the lower Ordovician Formations (Beekmantown carbonates) much of the eastern North American shelf was uplifted (the upper Gailor locally displays paleokarst effects of solution-collapse breccias, which can be found on the northern wall of the Gailor Quarry) and eroded, creating the Knox unconformity. This feature is defined as the upper limit of the Gailor Formation and its Mowhawk Valley equivalent, the Tribes Hill Formation. The gap in the rock record is thought to span about 5 million years (Fisher, 1980), during which time considerable erosion took place.

Above the Knox unconformity, the Lowville Limestone is sometimes - but not always present, and where it appears in the Saratoga Springs area it is less than .5 meters thick (Fisher, 1951). An extremely fine grain size, conchoidal to subconchoidal fracture, tan-gray color, to a chalk white weathered color make it a very distinctive rock. Above the thin Lowville limestone, or the Knox unconformity, the Amsterdam is a lumpy bedded, blocky, dark gray to black, calcilutite to calcistiltite which weathers grayish white. It is typically about 1.3 meters thick (Fisher, 1951), and where the Lowville is not present, its basal bed is a thin mottled conglomerate containing pebbles of the Gailor dolomites (Cushing and Rudeman, 1914).

Based on fossil evidence, Fisher (1966) describes sub facies within the Lowville suggesting intertidal (vertical worm borings) zones followed by inner shelf reef deposits (very abundant tabulate coral colonies), and infratidal shelf fauna typify the middle and upper Amsterdam. He assigns a Wilderness (Caradocian) age to the group.

Glens Falls Limestone

For the limestone intervening between the Amsterdam below, and the Utica (Canajoharie) Shale above, in the Mohawk Valley and in the Saratoga-Glens Falls region, Ruedemann (1912) proposed the name Glens Falls Limestone and assigned a Trenton "age" to it, as did Kay (1937) who formally divided the Glens Falls Limestone into the Larrabee, a lower, massive, crystalline member, and the Sugar River (Kay, 1968), an upper, thin-bedded, fine grained member. Fisher (1984) ascribes a late middle Caradoc age to them.

The Larrabee Limestone is a light to medium gray, thin to medium bedded dark gray weathered calcarenite to calcisiltite. From a drill core taken north of Geyser Road, Fisher (1984) found it to be 63 feet (19 meters) thick. Locally, (at Rock City Falls, for example) there is a basal, lumpy bedded somewhat conglomeratic argillaceous calcilutite, about 2.5 meters thick. The Larrabee is very fossiliferous with brachiopods being very numerous, along with crinoid debris, snails, and some trilobites. Fisher (1966) concludes that the Paleoenvironment was of the inner shelf zone, with high depositional energy but a prolific marine bottom fauna.

The upper Glens Falls Limestone is not exposed, but is known from cuttings in the mineral wells drilled throughout the area. Fisher (1984) calls it the Sugar River Limestone (a Mohawk Valley equivalent) and describes it as an alternating sequence of medium to dark gray fine grained and light gray shelly limestones. Titus and Cameron (1976) describe a decreasing diversity of fauna, occupying a transitional shelf to basin facies in a "quiet, deep water environment on a muddy substrate".

Utica Shale

The Glens Falls (Trenton) Limestones (including the Isle LaMott limestone which intervenes as part of the "Trenton" in Glens Falls) pass upward into the Utica, or Canajoharie shales, the latter a name coined by Ruedemann (1912) that is going out of style. The Canajoharie is considered to be an equivalent of the lowermost Utica shales, and it is the lower Utica then, that is the (poorly exposed) surface rock in much of the South-eastern Saratoga quadrangle. It is a fissile and splintery soft black shale which weathers grayish brown, becoming spotted whitish when somewhat calcareous. The basal parts are argillaceous limestone, and calcium content decreases upward in the sequence (Delano, unpublished data). Frequently intercalated within the soft shales are harder, bluish-gray mudstones indurated by calcite which are 7 to 15 cm thick, but occasionally 1 to 1.2 meters thick, which fracture concoidally. Much of the shale contains pyrite, which oxidizes and leaches into cleavage and bedding planes as rusty brown colored limonite. Several K-bentonite horizons are common among exposures in the Mohawk Valley.

Because the fissile shale is easily eroded its occurrence in outcrop is very rare. The best exposures occur along the Northway (Route 87) north of Exit 12, along the Kayderosseras Creek in Ballston Spa, and along Geyser Creek just south of Saratoga Springs. Its thickness approaches 335 meters in the Eastern Mohawk Valley, thinning westward to 110 meters over the Adirondack Axis and thickening again westward to 200 meters (Fisher, 1966).

The fauna represented is largely pelagic, of very low diversity, consisting chiefly of graptolites, with scant appearances of benthonic scavengers (Triarthrus Trilobites, for example, Titus and Cameron (1976) and Fisher (1966). The lack of burrowing traces, the black color (carbonaceous matter) of the sediments and pyrite content strongly suggest an euxinic deep water depositional environment (Nickolds, et al. 1979). Fisher (1984) places it in the Barneveld Stage, or Late Caradoc (Late Middle Ordovician) in age, becoming diachronously younger toward the west.

Schenectady Formation

The Schenectady Formation is a very thick (approaching 790 meters maximum, Fisher, 1966), randomly interbedded collection of silty gray and black shales, greywackes, and argillaceous sandstones. Many turbidite depositional features exist, such as sole markings and flow rolls. A product of both vigorous erosion and rapid sedimentation, the formation accumulated in a sinking basin on the continental slope, in front of the rising Taconic Mountains (Rowley and Kidd, 1981). It conformably overlies the Utica Shale and occurs west of Saratoga Springs (where down faulted east of the Hoffmans Fault) near the Town of Galway. Fisher (1984) considers it to be late Mohawkian or Caradoc (Latest Middle Ordovician) in age.

Snake Hill Shale

The type locality for the Snake Hill Shale is an outcrop in the eastern most portion of the Saratoga Quadrangle, near Saratoga Lake. It is a dark gray, silty, micaceous shale with intercalations of thin (occasionally thick) bedded siltstones and fine sandstones. More rarely, these are calcareous.

The formation is severely deformed (and often faulted) everywhere it is exposed, and occurs as discontinuous allochthonous sheets from Glens Falls south through Albany. It is slightly older than its autochthonous westward equivalent - the Utica Shales, and was emplaced by westward directed overthrusting during the Taconic Orogeny. An estimate of its thickness would be difficult if not impossible, however, Ruedemann (1930) assigned the unit a minimum thickness of 915 meters.

3.1 OVERBURDEN STRATIGRAPHY: A GLACIAL HISTORY

The records of the Wisconsin Ice Age are recent and widespread, so that its history has been worked out in great detail. However, the particulars of the event were not deciphered without difficulty, especially on a local scale, and in the last 30 years researchers have taken advantage of information (such as aerial photographs and well logs) not available to the earlier workers.

The effects of the ice sheets can be exceedingly complex because the ice did not advance on a simple front but pushed out in lobes that initially followed the lower terranes. In its retreat it paused repeatedly and locally readvanced (especially in later stages), leaving a cryptic array of marginal moraines, and ice override/retreat features.

About 120,000 years ago (Mickelson, et al. 1983), the Laurentide ice sheet centered itself over the Hudson Bay lowlands and began radiating outward. The ice entered New York in two major lobes; the St. Lawrence-Ontario Lobe and the Hudson-Champlain Lobe (Figure 3.4). The physiography of the state controlled the ice movement which tended to flow along the axes of major preglacial lowlands and diverge around highlands, so that the Hudson-Champlain Lobe became divided into the 1) southward-flowing (down-valley) Hudson Lobe, the 2) Mohawk Lobe which flowed westward up the Mohawk Valley, and 3) the Adirondack Lobe that followed the northeast/southwest structural and topographic trends of the Adirondack Mountains (Dineen and Hanson, 1985).

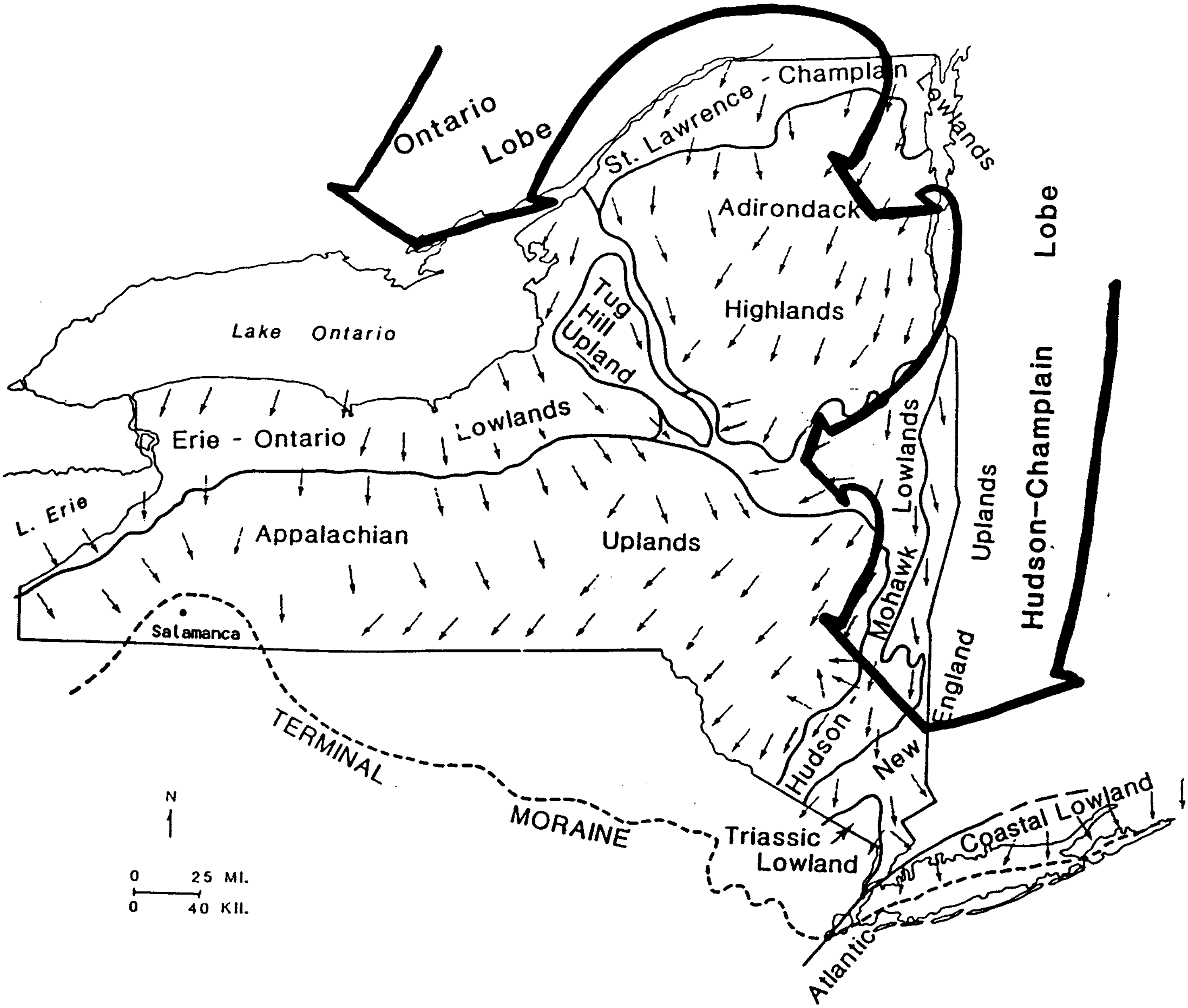


Figure 3.4 Physiographic provinces and regional ice flow movement directions (small arrows) in New York State. The Wisconsin glacial ice sheet came from the Laurentian Mountain region of Quebec and was initially divided into two main lobes (heavy large arrows) which flowed around the Adirondack Highlands. The Hudson-Champlain Lobe was further divided into the southward flowing Hudson Lobe, the westward flowing Mohawk Lobe, and the Adirondack Lobe which followed the structural and topographic trends of the Adirondack Mountains. Eventually nearly all of New York was covered, and the terminal moraine delineated indicates the maximum southern extent of the glacier.
(After Dineen and Hanson, 1985)

The evidence indicating the directions of glacial advance include the orientations of drumlins and striae, and even the lithologies of rocks deposited in tills and stratified drift. At its peak, about 22,000 years ago (Mickelson, et al, 1983), the Wisconsinian ice covered New York State except for the Salamanca Re-entrant in western New York and the southern edges of Staten Island and Long Island (Denny, 1956; Flint, 1971; Figure 3.4). The depositional characteristics resulting from the advancing Hudson Lobe along the McGregor escarpment often begin with a basal lodgement till and/or marginal moraines, usually less than 10 feet thick (Connally, 1975). Down-valley movement typically deposits lodgement till upon bedrock. Up-valley movement can impond rivers so that lodgement tills are subsequently deposited on proglacial lacustrine silts and clays (Figure 3.5). At Sherman Island (west of Glens Falls) where the Hudson River flows over the McGregor Escarpment, the southward flowing ice first plugged and then invaded that valley. Overburden exposed on Corinth Road south of the Luzerne Mountains (Figure 3.9) displays a black varved clay with drop pebbles (siltstones) which is often deformed where the upper margin shows a contact with a compact black lodgement till. At higher elevations the black clays develop extremely thin seams or "microvarves" which can occur in lakes starved for sediment. The probable path of the ice advance as suggested by drumlins and glacial striations (DeSimone and LaFleur, 1985) is depicted in figure 3.9. The Palmertown Range bore the greater brunt of the glaciers' impact and as a result experienced greater erosion on the northeastern flank. This is a recurring feature of steep tributaries on the western margin of the Hudson River Basin (Kings Station; Daniels Road; Palenville and Woodstock are other examples).

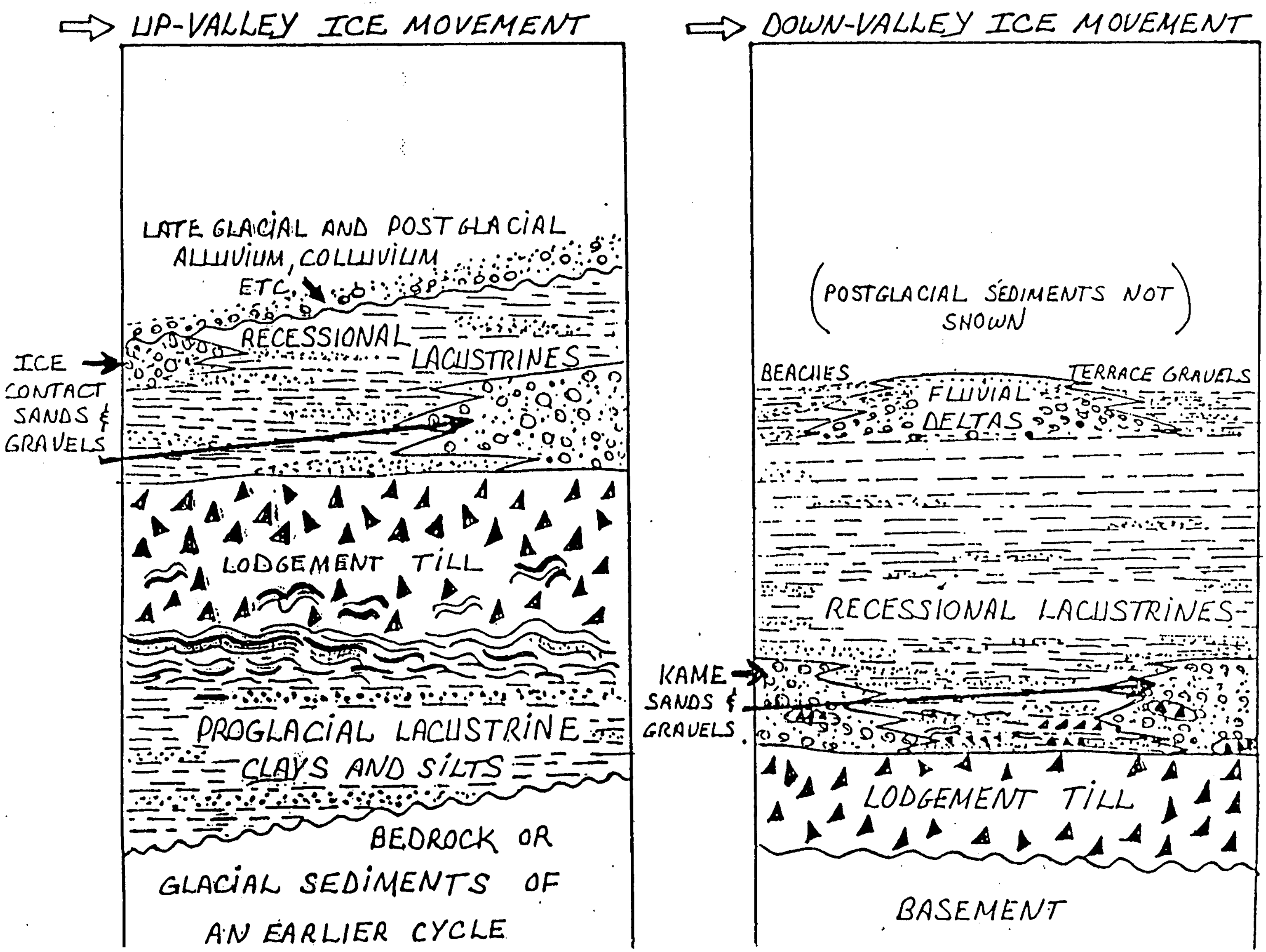


Figure 3.5 DeSimone, Unpublished Work.

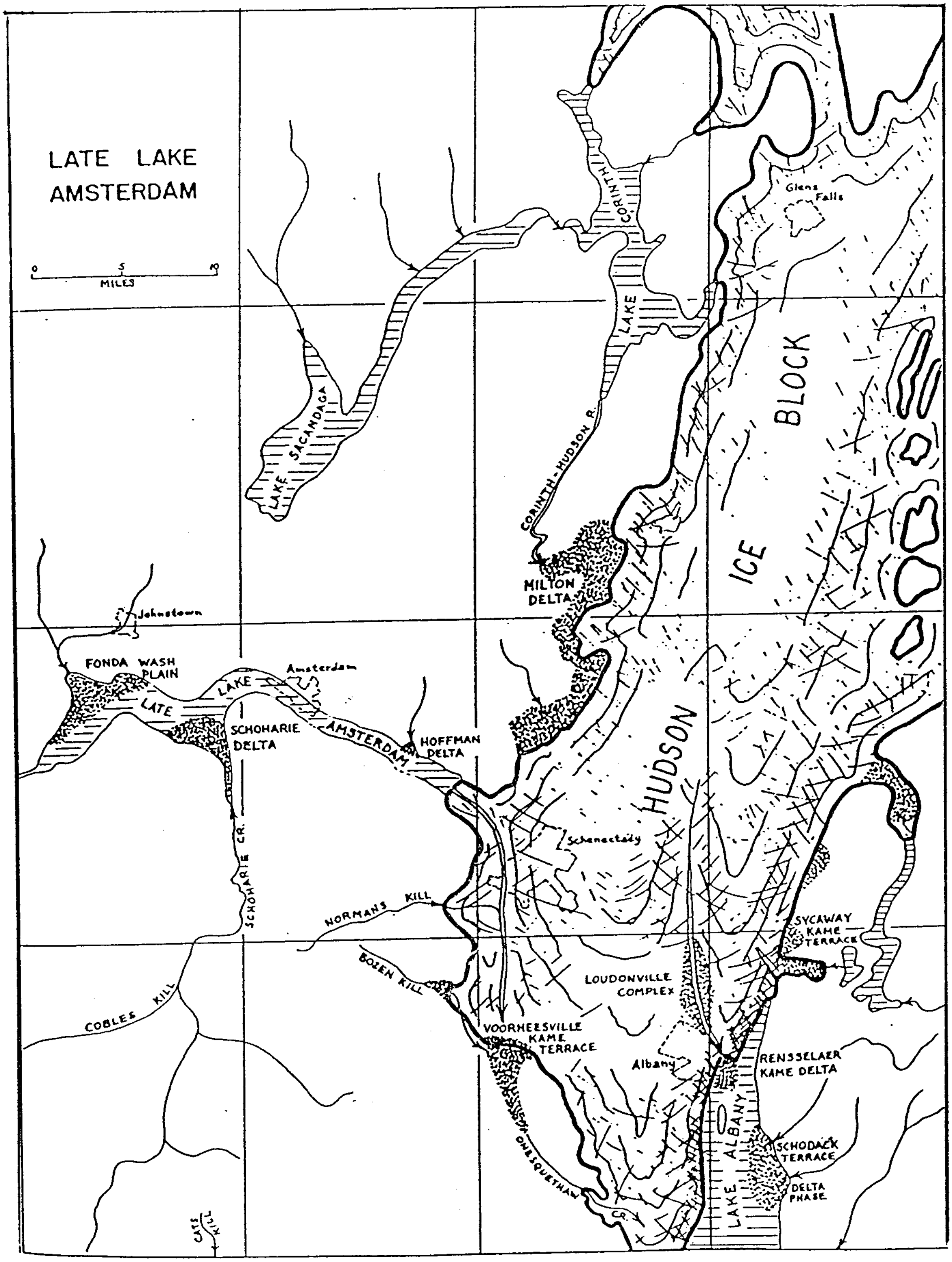


Figure 3.6 A sketch map interpreting conditions during the waning Late Lake Amsterdam Stage of the Wisconsin Ice Age approximately 14,700 yrs. ago (DeSimone and LaFleur, 1985) in Eastern Central New York State. The retreating Hudson Ice Block occupied much of the lower Hudson Valley. (From LaFleur, 1965)

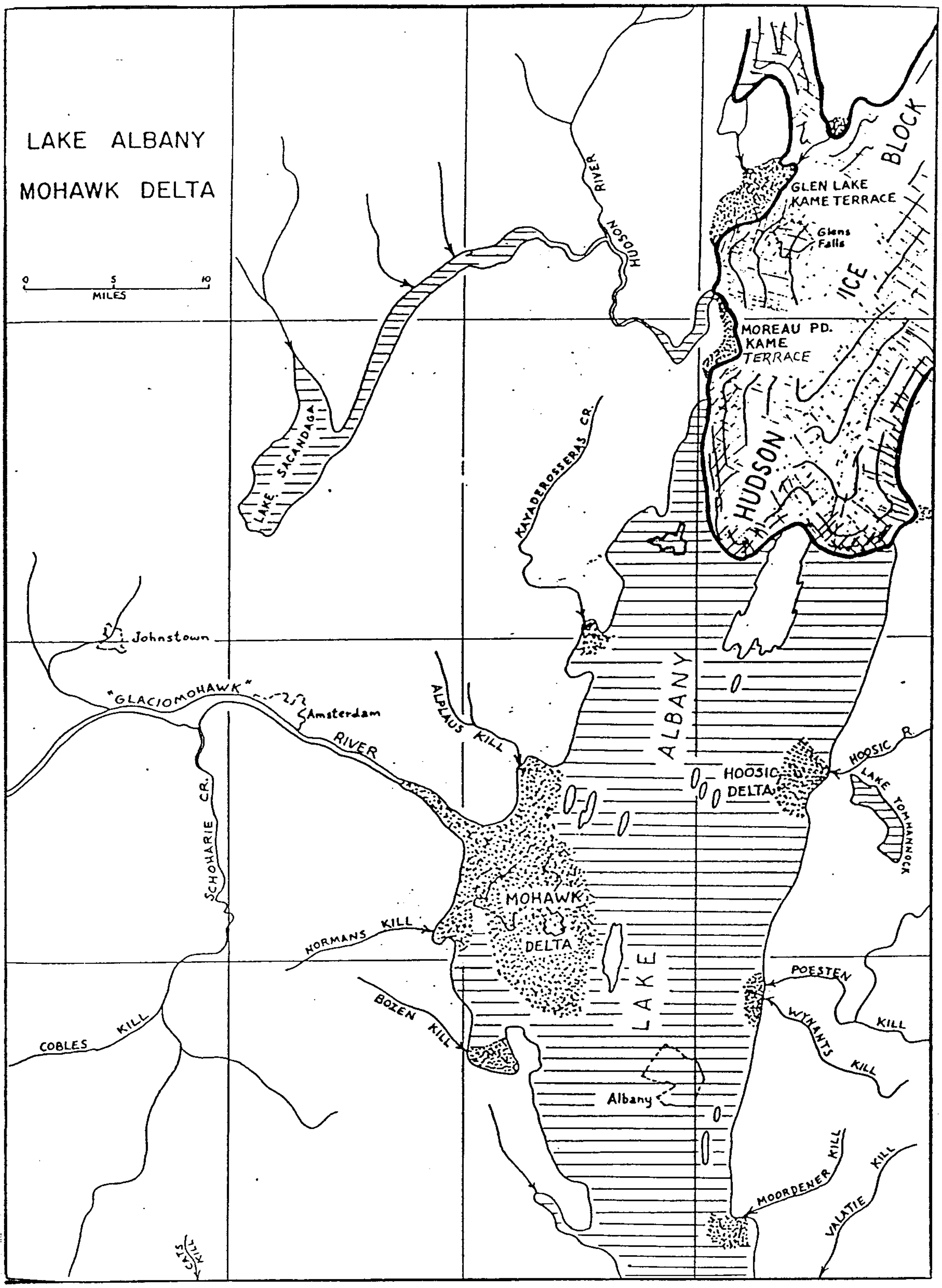


Figure 3.7 No longer blocked by the retreating Hudson Ice Block, rivers fill ancient glacial Lake Albany and deposit deltas. Near ancient ice margins, kame deposits are common. The Moreau Pond Kame Terrace/Kame Delta is an example near the base of the Pamertown Range/McGregor Escarpment, where sediments are deposited up to an elevation of 420 feet. (From LaFleur, 1965).

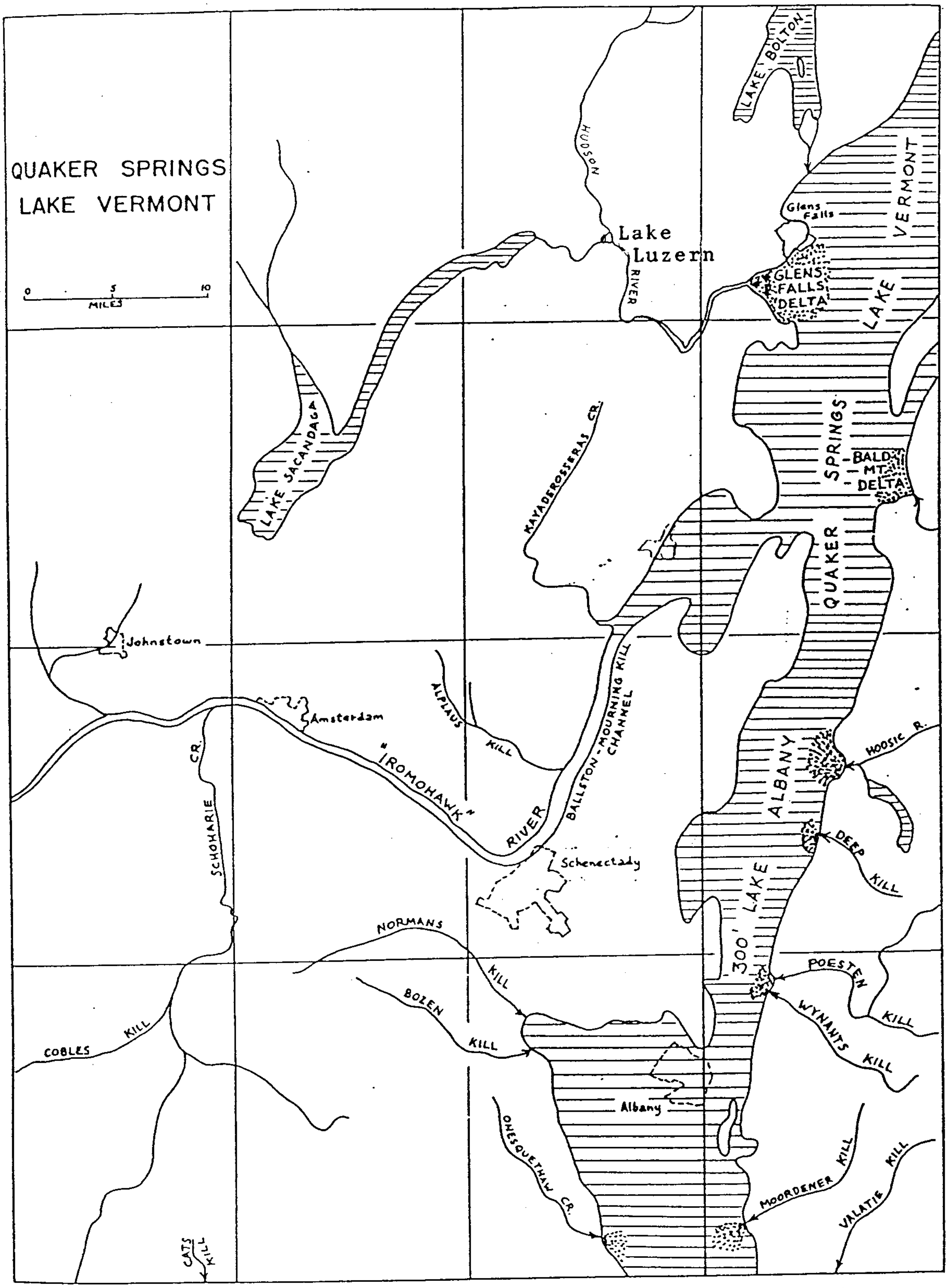


Figure 3.8 The elevation of ancient Lake Albany has fallen (in stages) to the 310 foot Lake Quaker Springs level (this stage began about 13,700 years ago, DeSimone and LaFleur, 1985). The 310 foot Glens Falls Delta is deposited. (From LaFleur, 1965).

The manner of the deglaciation in the Hudson lowlands involved the creation of a series of lakes which formed as meltwater was dammed toward the south by glacial debris (Figures 3.6, 3.7 & 3.8). Uplands were exposed early in the warming trend, due to the thinner ice cover. Evidence indicates that except for a couple of minor readvances (Dineen, et al., 1983; DeSimone, 1985), there was a continuous episode of ice stagnation through the area north of the Catskills, the easternmost Mohawk, and northern Hudson Lowlands, and that a lacustrine environment frequently accompanied the retreat (DeSimone and LaFleur, 1985; LaFleur, 1979; LaFleur 1965). The melting Hudson Lobe contributed a great deal of sediment to the ice-marginal Lakes Albany and Quaker Springs and a characteristic lacustrine sequence was deposited. Typically, a basal facies of interbedded gravel and sand (occasionally with turbidites and minor flow till) grades upward into a middle facies of fine clay and silt rhythmites (varves). The rhythmites grade upward into an upper facies of shallow water silt and sand beds.

The margins of the Hudson River basin are often found to contain ice-contact stratified drift (kame terraces), usually deposited by meltwater in close proximity to stagnant ice; slump structures and kettles are common. They contain chaotically to regularly bedded sands and gravels with minor silts and clays. But as the glacier continued to melt away from valley walls, unimpeded rivers began depositing sediments into the ice-marginal lakes. In the Sherman Island area these deltaic sediments first appear at about 410 feet (Figures 3.9 and 5.1), roughly coinciding with the highest stable level of Lake Albany, about 14,700 years ago (DeSimone and LaFleur, 1985). Bore hole logs north of the river record 15-18 feet of lag deposits of boulders and coarse sand above over 75 feet of coarse to fine sands (Empire Soils, 1991), well stratified with

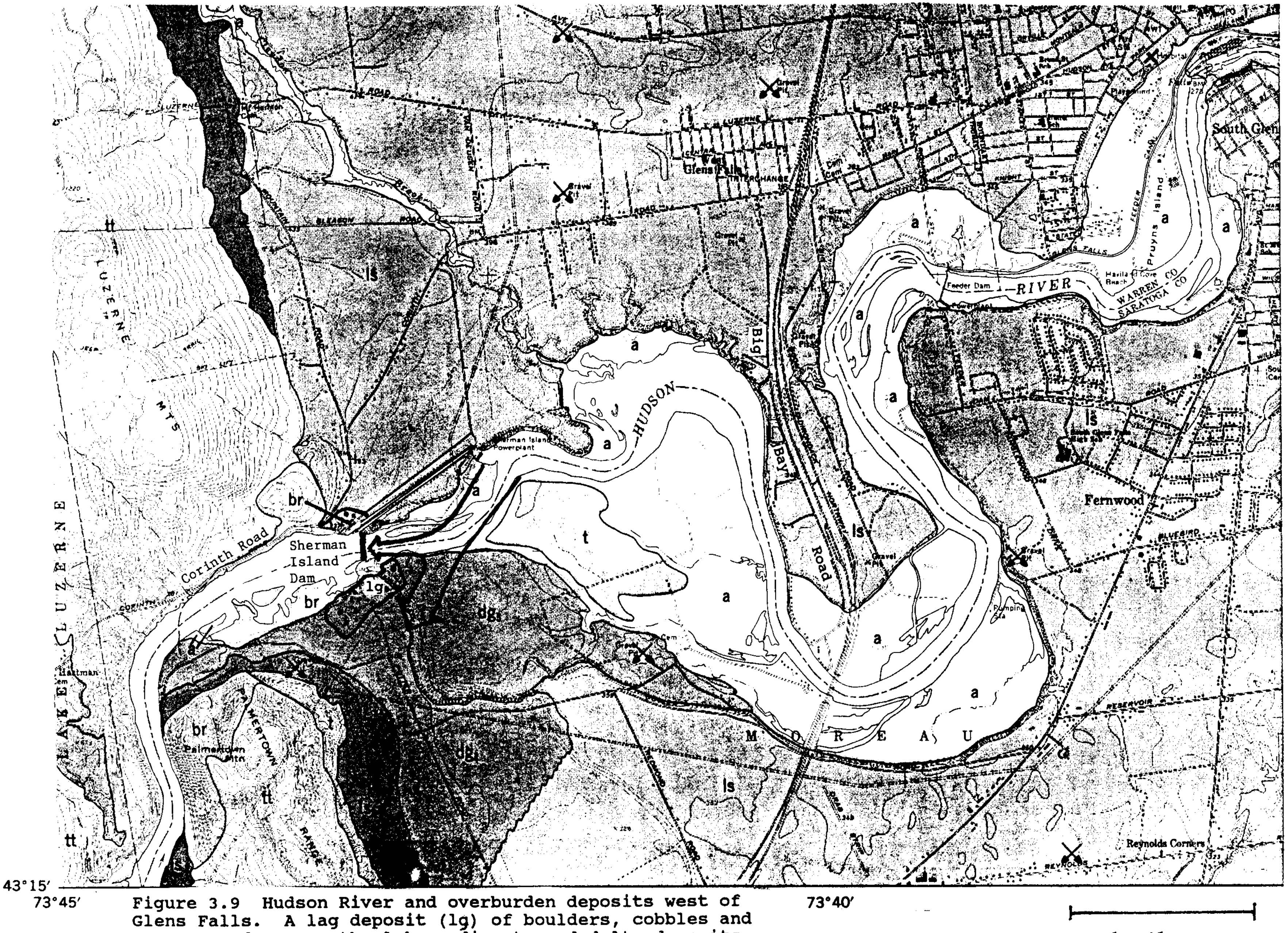


Figure 3.9 Hudson River and overburden deposits west of Glens Falls. A lag deposit (lg) of boulders, cobbles and coarse sand covers the lake sediments and delta deposits immediately N & S of the river near the damsite. Arrows denote ancient glacial ice flow paths. Modified after Connally, 1975.

occasional clay seams (bottomset beds). Connally (1975) describes the deposits south of the river as delta gravels; "...fine pebbles to boulders with up to 50% sands and silts exposed as large scale foreset and thick topset beds". About 13,700 years ago Glacial Lake Albany fell to 310-315 feet and stabilized as Lake Quaker Springs, an elevation in the middle of a sequence of falling lake levels, (DeSimone and LaFleur, 1985; Stewart, 1961; Figures 3.7 & 3.8). Shortly thereafter the Hudson Ice Front retreated north of the Glens Falls area (Connally and Sirkin (1969) dated bog deposits from near Lake Luzerne at 12,400 yrs.) and the Glens Falls Delta (at 315') began to be deposited. With subsequent lake level recessions (and extinction) the Hudson River began eroding a channel through this delta also, to its present day elevation of about 290 feet.

Seismic refraction data (Fine Hsu; personal files) has shown the existence of a buried pre-glacial bedrock river channel, carved into the McGregor escarpment to an elevation of 168 feet at the Sherman Island damsite. This suggests that at this location, an ancient 'Hudson River' once flowed over the edge of a pre-glacial hanging valley, which was subsequently buried under syn- and post-glacial deposits. It is obvious that much erosion has occurred here producing the large gap between the Luzern Mountains and the Palmertown Range, bearing a direct relationship with the existence of the Glens Falls Delta. This observation is apparently in conflict with the speculation that the escarpment may be recently (within Neogene time) uplifted - no other rivers or streams have incised so deeply. This conflict may be reconciled if we admit to the possibility that the ancestral Hudson existed here as a result of a pre-existing weakness in the bedrock. It is known, for example, that much of the course of the upper Hudson is fault controlled. Fisher (1984) has tentatively placed a fault striking roughly E-W under the Hudson at this location. Also, the Wisconsin Ice Age is only the most recent of four Pleistocene glacial advances (Nebraskan, Kansan, Illinoisian). It is surprising that there are not more examples of extensive escarpment erosion. Possibly no significant ancestral drainage existed other than the Hudson.

Of the overburden deposits proximal to the McGregor fault system, the horizontally stratified delta deposits (bottomset beds especially) of the Glens Falls Deltas (Figure 3.9) appear to offer the best opportunity for a paleoseismological excavation. The area is largely undeveloped, so sediments are unlikely to have been disturbed except for shallow agricultural purposes, and much of the terrain is second growth forest. Another advantage of the locality is that the fault scarp displays progressively increasing displacement toward the north, consequently, greater geologically recent movement may have occurred there.

4.0 PALEOSEISMOLOGY: EVALUATING SEISMICITY

In this section I hope to demonstrate that the Quaternary geologic record is a potentially valuable source of information regarding seismic hazard that has gone largely untested in the Northeastern United States. In many instances recurrent Paleozoic or Mesozoic movement on a fault can be documented, but recognition of subsequent displacement is difficult due to the absence of intact, preserved, or suitable younger stratigraphy. Where indirect evidence suggests that more recent faulting activity has occurred, suspicions should be followed up by investigating the Quaternary sedimentary stratigraphy.

The brief (400 yr.) historic earthquake record in North America should be used with extreme caution in estimating possible future seismic activity. Historical records of intraplate earthquake activity in China go back over 2000 years and show that events do not always recurr in the same location over time, and that the recurrence period can be even longer (Wu, 1988). It should also be noted that occasionally very powerful earthquakes have apparently occurred on innocuous or unidentified faults (the Charleston, South Carolina earthquake of 1886 for example, Obermeier et. al., 1989).

The information in recent sedimentary deposits can potentially take us several thousand years beyond the historical record, and contribute to the understanding of long-term seismicity. This is critical to the siting and design of safe structures, and to the establishment of realistic building codes (i.e. appropriate risk appraisal).

Paleoseismology has been defined as "the identification and study of prehistoric earthquakes " (Wallace, 1981). Its focus is to identify and study active faults, and hopefully to constrain the potential for, and the timing of, future seismic events. The techniques involved include analysis of microstratigraphic relations along faults (Engelder, 1974), fault-scarp and fault-trace geomorphology (Wallace, 1980), seismically induced sedimentary structures (Clark, 1972; Seih, 1978), and river and marine terraces related to uplift and faulting (Plafker and Rubin, 1978; Madole, 1988). Excavations specifically intended for paleoseismological studies first began to become popular in the early 1970's (Clark, 1972; Malde, 1971). In trenches excavated across the Coyote Mountain Fault in California, Clark (1972) found clear evidence that fault displacements had occurred more than once (Figure 4.1). Later in 1978, Kerry Sieh dated (using Carbon-14 analyses on peat layers) a series of prehistoric offsets on the San Andreas fault at Pallett Creek, California. Each major displacement occurred at intervals averaging 140 years, going back to the 6th century A.D. It must be emphasized, however, that although excavations can provide a great deal of valuable data about prehistoric earthquakes, this will only happen under the circumstances where indicative sedmentary structures exist near a potentially active fault, and datable materials are found. A problem then, is that the locations chosen for this type of study may not represent an area where the maximum amount of recent movement has occurred on a fault, and are limited in that only areas covered by stratified sediments are likely to yield any meaningful results.

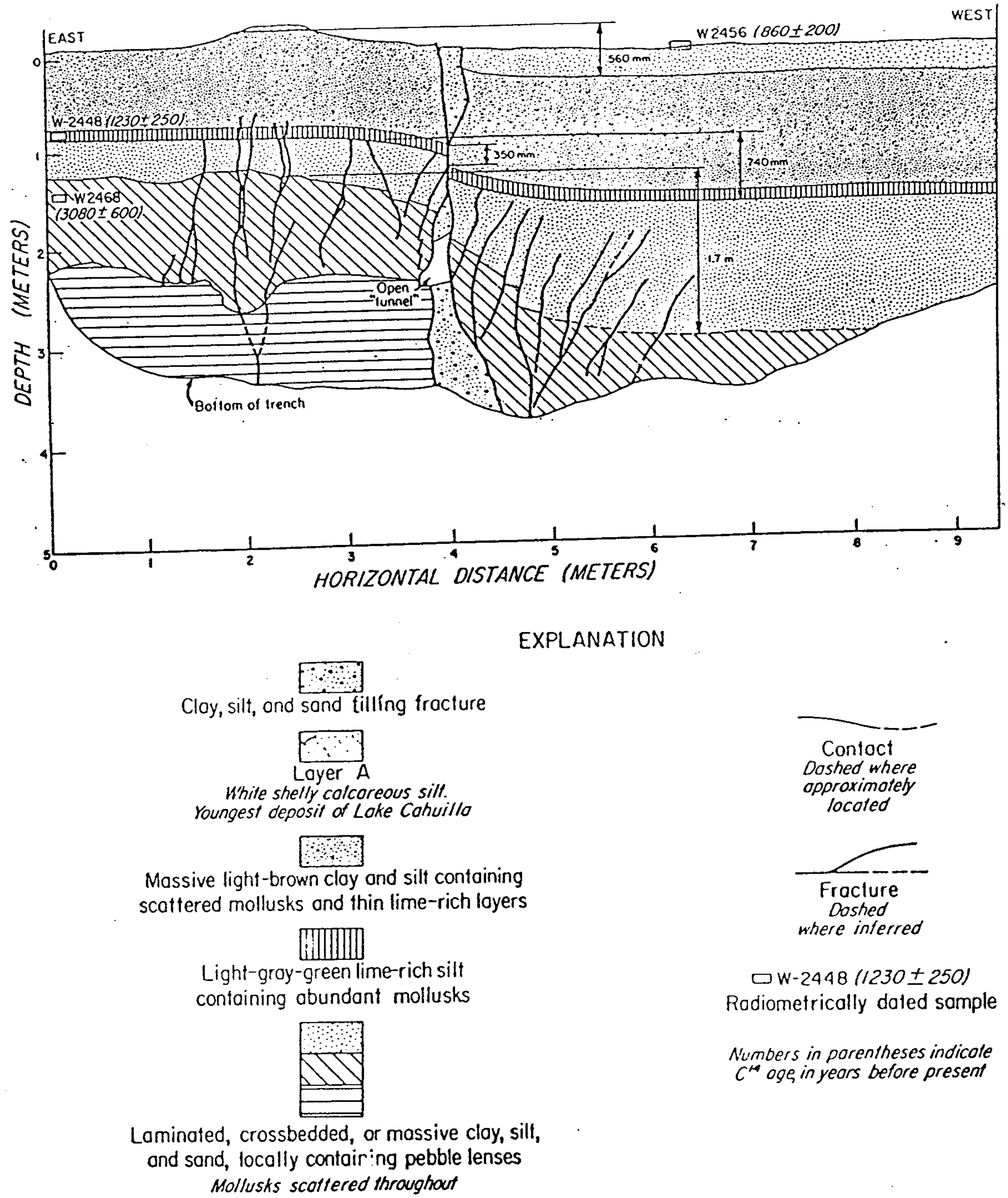


Figure 4.1 A sketch of a profile from a paleoseismic excavation in ancient lake sediments proximate to the Coyote Creek Fault, about 40 km east of San Diego, California. Shows progressively greater offset of older strata and bending (drag) of strata across the break (From Clark et. al., 1972).

The excavation of sediment has become an extremely valuable paleoseismological technique for earthquake study over the last two decades. The method simply involves excavating trenches in the overburden above a known or suspected fault locality, and examining the soil profile in situ. Abrupt, prominent (greater than 5 cm) offsets in loosely consolidated sediments are very suggestive that faulting was associated with tectonic earthquakes (Clark, 1972; Allen, 1974; Sieh, 1978), and the additional exposure of faulted bedrock (with fault gouge and slickensides) lends further support (York and Oliver, 1976; Sieh, 1978). Other good paleoseismic indicators include the effects of soil liquefaction, such as sandblows (Figure 4.2), sand or clay dikes or diapirs, and convoluted layering (Scott, 1973; Coates, 1975; Sieh, 1978). Also, folded soil horizons, or those that are found to be steeply dipping parallel to the underlying bedrock and were folded or tilted with that bedrock (Vanarsdale, 1986) can be considered to be the result of seismic activity. In the most favorable circumstances, the presence of contemporaneous buried organics (soil humus, peat, or snails for example) can provide Carbon-14 ages that constrain the time of faulting.

A study by John Sims (1975) involved deformational structures in lacustrine, ancient lacustrine, and glacio-lacustrine sediments proximate to active seismic zones. The correlation with seismic events was based on several criteria including a similarity to structures formed experimentally (Figure 4.3) and their restriction to single stratigraphic intervals over large areas of a lake bottom. The structures include small scale folds, load casts and evidence of liquifaction.

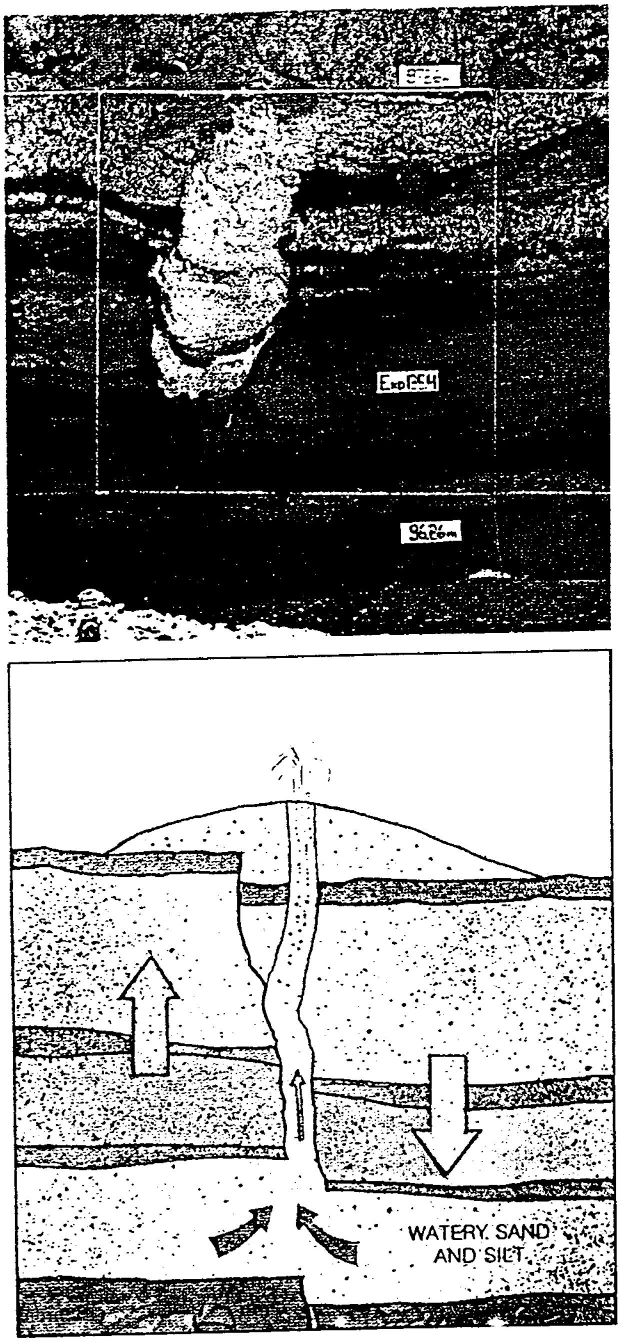
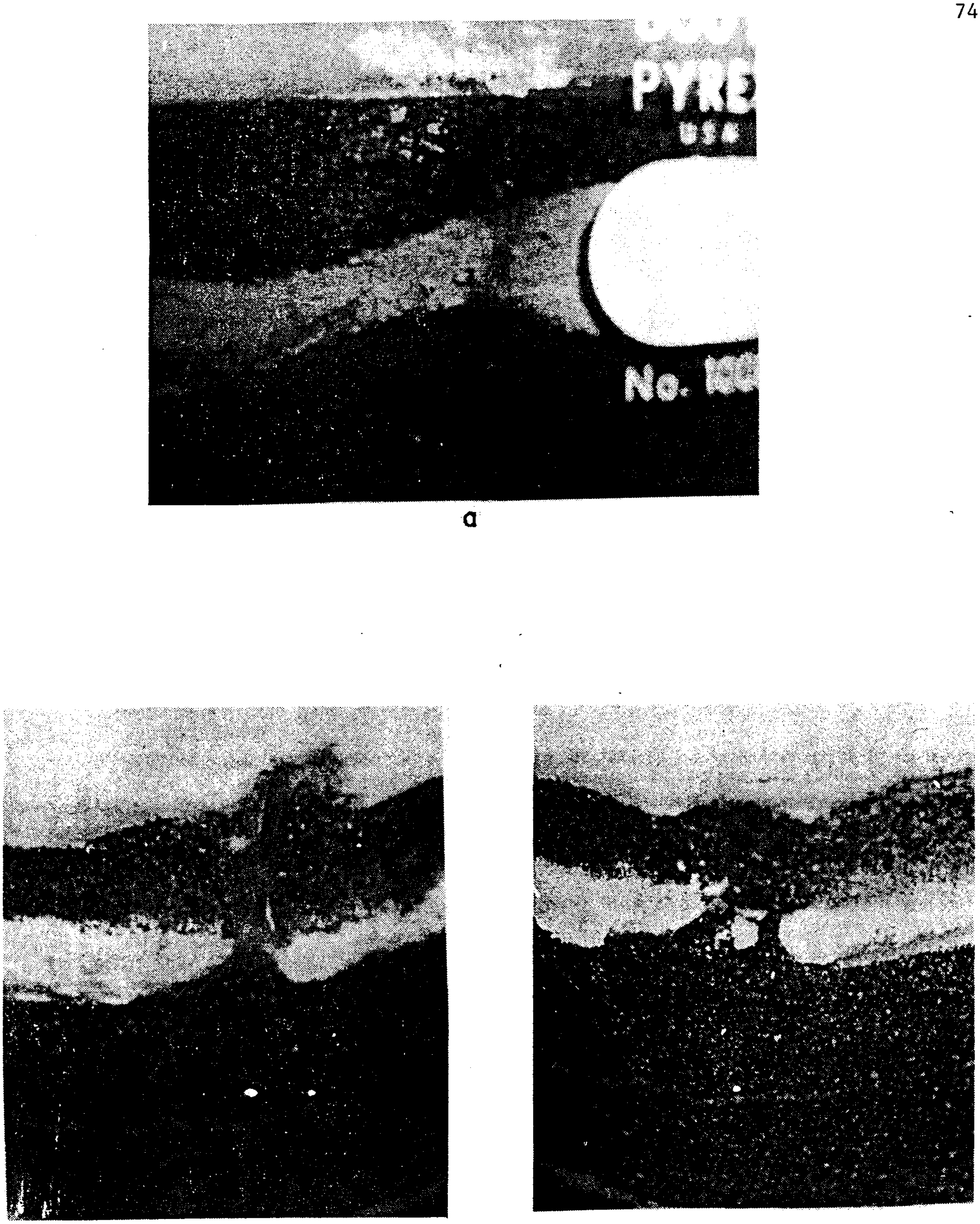


Figure 4.2 Remnants of a sandblow - a spouting of sand and water caused by moderate to severe earthquakes - can be seen in the photograph of a cross section of an old California stream bed that was rocked by a quake around 1700. Sandblows occur, as illustrated in the drawing at bottom, when a layer of subsoil takes on liquid characteristics during tremors. Pressure drives the watery sand and silt up through a fissure, leaving a mound of sediment on the surface that geologists can use to identify and date the earthquake (From Walker, 1982).



a

b

c

Figure 4.3 A series of photographs showing sediment changes during shaking table experiments. Layers of medium sand were sandwiched between layers of kaolinite (white), then saturated and shaken (not stirred). The sequence shows a) the formation of a fluid beneath a kaolinite layer, and up-arching, b) penetration through a kaolinite layer and ejection of sand, and c) settlement.
(From Coates, 1975).

Under the right conditions, sudden uplift or subsidence can be recognized in fault stranded stream deposits. Particularly, localized slopewash and fan alluvium deposited on the downthrown side of a fault shortly after the time of faulting (Figure 4.4).

Neotectonics is a technique which grew from field observations of landmarks that were displaced by repeated earthquakes on a known fault. The lateral offset of streams is ideal for detecting strike slip fault displacements, for example, and steep vertical offsets where little erosion has occurred in the upthrown block can be suggestive or reverse or normal faulting. Where man-made structures are involved (buildings, roads), an average slip rate can be calculated by dividing the distance of movement by the age of the feature.

The application of paleoseismic excavation investigations in the Eastern United States has yielded only limited success in identifying earthquake induced structures in overburden. The recognition of seismically generated liquifaction events has proven to be a useful line of inquiry but only where specific conditions exist. Liquifaction effects have been demonstrated and explained experimentally (Coates, 1975; Dobry, 1989; Figure 4.3) by cyclically shaking a stratified sample system. Occluded layers such as a bed of sand underlaid and topped by thin layers of clay possess differences in porosity and permeability so that pore pressure will accumulate and dissipate at different rates. Saturated sand layers can behave like a fluid, "exploding" upwards, characterized in the occurrence of sand blows. Sand blows are of two types: craterlets or vented-sand volcanoes (Obermeier, 1987). A craterlet is a hole created when liquefied sand vents to the surface and

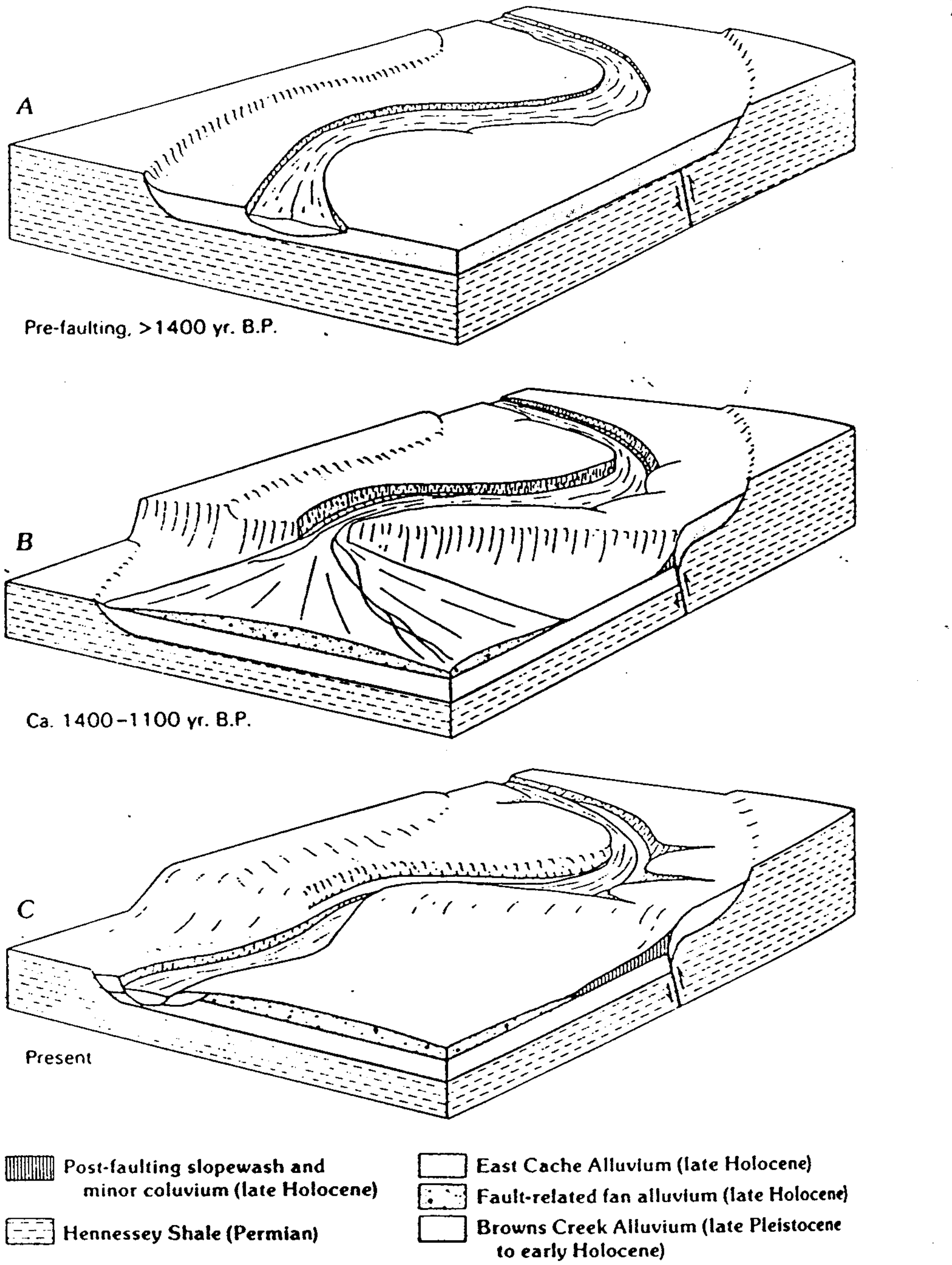


Figure 4.4 Block diagrams showing relations between the Meers fault and upper Quaternary alluvial deposits. A. Unfaulted Browns Creek Alluvium. B. A reverse fault that dips steeply northeast formed a scarp that is as much as 5 m high. Most of the scarp height was formed by warping rather than by stratigraphic displacement. Fault movement, 1400-1100 yr B.P. caused stream incision on the upthrown side of the fault and deposition of fan alluvium and slopewash on the downthrown side. The alluvium buried charcoal and soil humus that provide maximum C dates for the time of faulting. C. Subsequent episodes of channel deepening alluviation, and incision during late Holocene time produced crosscutting stratigraphic relations, cutbank exposures, and C-datable materials that provide minimum dates for the time of faulting (From Madole, 1988).

moves aside surface (soil) material. The craterlet, commonly 1-3 m wide and 1 m deep, is subsequently filled in with adjacent material. Those produced in the New Madrid seismic zone by the 1811-12 earthquakes are cones, 0.5-1.0 m high and 15-60 m in diameter. Much of the overburden consists of a thick, clean, medium-grained sand overlain by a 2-10 m thick clay layer. The clay cap is shattered by extruded sand which contains stratigraphic characteristics indicating flowage, away from a central vent system, and becomes finer grained upward. The following account of an event during the earthquakes of 1811-1812 in New Madrid, Missouri is typical; "Great amounts of liquid spurted into the air. It rushed out in all quarters, bringing with it...carbonized wood, reduced ... to dust, which was ejected to the height of from ten to fifteen feet, and fell back in a black shower, mixed with the sand which its rapid motion forced along;" (Penick, 1981). A third type of sand blow is described in the literature as a sand boil, which was originally defined as a spring that bubbles through a river levee accompanied by an ejection of sand. It's use is commonly applied to smaller scale (down to about 10 cm), less 'explosive' sand volcano features (but the terms are interchangably applied by some authors) and apparently are a result of deposition over a longer time interval. Sand boiling was observed to occur over a 20 minute interval shortly after the end of the 1978 Miyagiken-Oki earthquake in Japan (Tohno and Yasuda, 1981).

Sand blows are typically connected at depth to a vertical or high angle sand-filled feeder dike (pipes are less common), which may be part of a network of dikes and sills, many of which pinch out laterally or vertically before reaching the surface. In some places, sand blows will form close together along linear features that can be hundreds of meters long (Saucier, 1989). The sandy source layer(s) are typically highly contorted laminated sediments.

Soft-sedimentary structures indicative of seismic events (M 5.0) include roughly planar sand-filled dikes that are vertical to steeply dipping and that widen downward and connect to a sediment source at depth. Typically, the dikes cut through a layer of low permeability such as silt and/or clay which overlies the source strata of silty to gravelly sand. It must be kept in mind that liquifaction can be caused by any system that generates waves, such as storms or seiches and that other mechanisms that produce these features, including permafrost effects, artesian springs, landslides, de-watering structures or collapse structures (due to glacial ice block melt-out, for example), might masquerade as earthquake induced features. These can be rejected if the dikes have the following aspects: i) they widen downward; ii) they are strongly aligned in local areas; iii) they vented to the original surface; iv) material in the dikes fines upward and was transported upward; v) bedding in the source beds is homogenized and contacts with overlying fine-grained material are highly convoluted; vi) the dikes occur in flat and topographically elevated landforms; vii) the size of the dikes decreases with increasing distance from a central area of large dikes (Obermeier, 1991). It is important to note that liquefaction features are not necessarily accompanied by offset features and vice-versa. Also, while earthquake induced liquefaction effects may not always produce dike-like features, the presence of wide-spread thixotrophy and diapirism of many units (mostly laminated silts and clays) can sometimes implicate earthquake shaking as the responsible mechanism, especially when accompanied by minor faulting and complex folding (Coates, 1975).

The smallest regional (historic) earthquake known to have induced liquefaction features was a (body wave) magnitude 6.2 (north of New Madrid, Missouri; Obermeier, 1988), while paleoseismic investigations suggest that prehistoric earthquakes as small as magnitude 5 to 6 have produced evidence of paleoliquefaction (in the eastern United States; Coates, 1975; Crone, 1987; seeber and Tuttle, 1989; Amick and Gelinas, 1991). While there is a great deal of liquefaction evidence in support of large (6.0 M), recurrent prehistoric earthquakes in the New Madrid area (Wesnousky, et al., 1989; Saucier, 1989; Russ, 1979), in the Wabash River Valley, (Obermeier, et al., 1991) and in South Carolina (Amick and Gelinas, 1991; Obermeier, et al., 1989), in only one case has strong evidence been unearthed in support of an earthquake induced liquefaction event in the Northeastern United States. The discovery was made with the help of historical reports of effects of the 1727 Newbury Earthquake, Newbury, Massachusetts, and deformed sedimentary structures were attributed to it and to a quake approximately half a millenium before that event (Tuttle and Seeber, 1989; Figure 4.5).

The reason offset is not observed in paleoseismic excavations in the Northeastern United States is because surface faulting offset effects are extremely rare in intraplate areas. There are only ten documented examples of surface rupture (of bedrock) from naturally occurring historic earthquakes in the more stable, continental, intraplate regions of the earth (Johnston, 1990; excluding earthquakes triggered by surface quarrying, Pomeroy, et al., 1974; Adams, 1982; or hydraulic mining, Fletcher and Sykes, 1977; and the like). Only one of these occurred in Eastern North America, the Ungava earthquake (magnitude= 6.3) of 1989, which occured in Northern Canada (Adams, et al., 1991). That earthquake

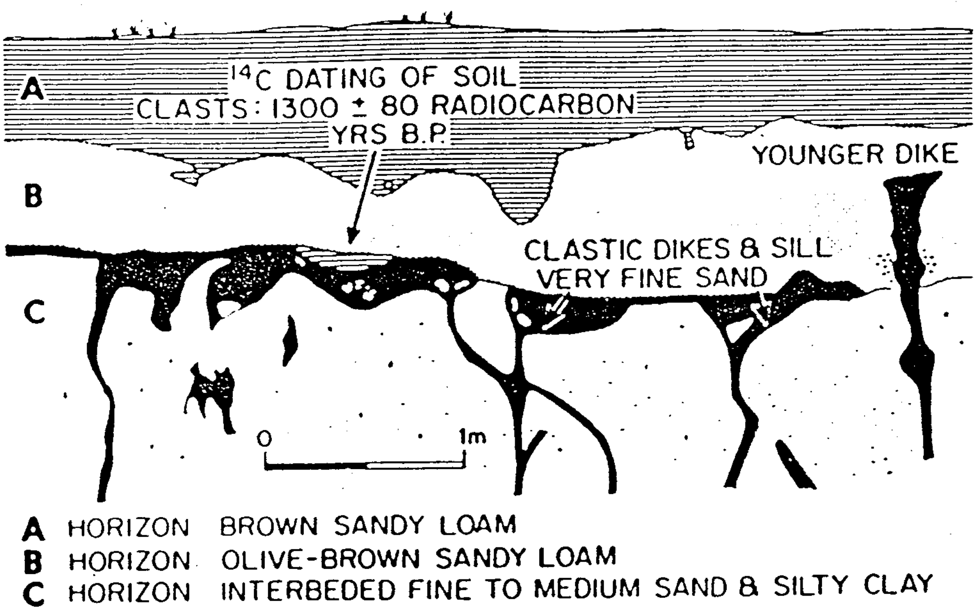


Figure 4.5 Clastic dikes and sills exposed in a trench of a low-relief dome in Newbury, Massachusetts. The incorporation of A and C horizon material in the sill indicates that sand was forcefully injected from below. A radio-carbon age of about 1300 years on clasts of the A horizon in dike sand and considerable soil development since the injection event suggest that these features predate the 1727 earthquake (Obermeier, 1989).

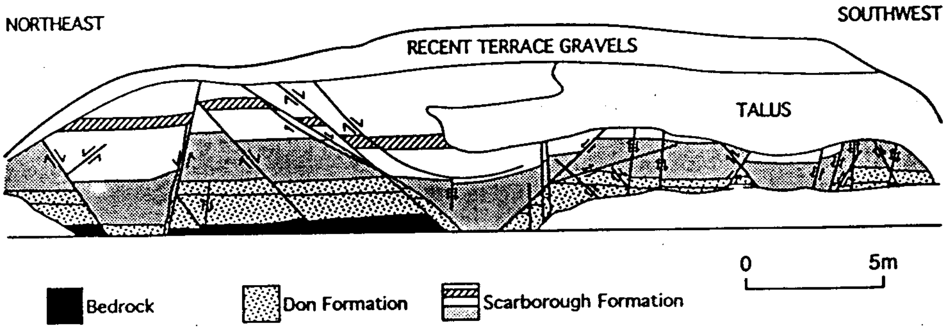


Figure 4.6. Faulted strata along the Rouge River, Toronto. The bedrock is Utica Shale. The Don Formation is Quaternary sandstone and mudstone about 125,000 years old. Note that the offset structures have no surface expression.
From Mohajer et. al., 1992.

was shallow enough to produce a surface rupture (centered at 60.12 N, 73.60 W) striking NNE, 8.5 km long with up to 1.8 m of reverse faulting offset. It was inferred by Adams, et al., (1991) that the faulting was controlled by pre-existing bedrock structures.

Significant paleoseismic fault offset has only been observed in one instance in Eastern North America, where steeply dipping normal faults offset both bedrock (Utica Shale) and overlying Quaternary sediments along the Rouge Valley in metropolitan Toronto (Mohajer, et al., 1992). The maximum displacement is about 1.25m and slickensides indicate dip slip motion (Figure 4.6). Based partly upon the magnitude of the offset, Mohajer et al. (1992) suggest that the faults represent the effects of uplift from glaacioisostatic rebound aided by high horizontal compressive stress and conclude that southern Ontario should not be assumed to be a region of negligible seismic risk.

Investigations in the southern midcontinent have turned up similar evidence. Examples include the Meers fault in Oklahoma (Crone and Luza, 1990; Madole, 1988), the Reelfoot scarp in eastern Tennessee (Russ, 1979) and the Kentucky River fault system (Vanarsdale, 1986). A result of these studies suggests that in addition to the New Madrid, Missouri seismic zone, much of the midcontinental United States is not tectonically stable (Crone and Luza, 1990), and that potentially hazardous faults may be aseismic for long periods of time and can only be identified by geological paleoseismological studies (Allen, 1975). In addition, these studies have shown that earthquake recurrence intervals can be on the order of 500 to 1000 years or more, suggesting that the historic earthquake record may be of limited value in earthquake hazard assessments.

5.0 LOCATING THE MCGREGOR FAULT BENEATH OVERBURDEN

Mapping, Well Logs, and Geophysical Methods

The stratified bottomset beds of the Glens Falls Delta at Sherman Island appear to offer a good opportunity for a paleoseismic excavation. However, the McGregor fault in this area is deeply buried by these deposits, and methods to accurately locate it at depth must be employed to adequately perform this investigation. There is much evidence suggesting that fracture traces and natural lineaments visible on aerial photographs are the surface manifestation of joints or faults in the underlying bedrock (Lattman and Parizek, 1964; Siddiqui and Parizek, 1971). Aerial photographs of the Sherman Island area were accordingly examined under a low-magnification stereoscope and one faint lineament was noted at a location and orientation (about 356°) suspiciously coincident with the inferred fault location as determined from bore- hole and gravity surveys (Figure 5.1). Excursions in the area have determined that this lineament is not the result of human activity (pipeline, fencing, etc.). Therefore, this feature may represent the McGregor fault trace, or the trace of a fault splay near the McGregor fault.

At a few localities to the south, the fault trace can be fairly well constrained by outcrops of brecciated or drag-folded Paleozoic rock close to Pre-Cambrian Gneisses (near Wilton and Kings Station, for example; Figure 1.3). These outcrops consistently appear at the break in slope between the steep escarpment face and overburden deposits of rather low relief. Several deep boreholes north of the field area immediately east of this slope break, never struck bedrock, and logs

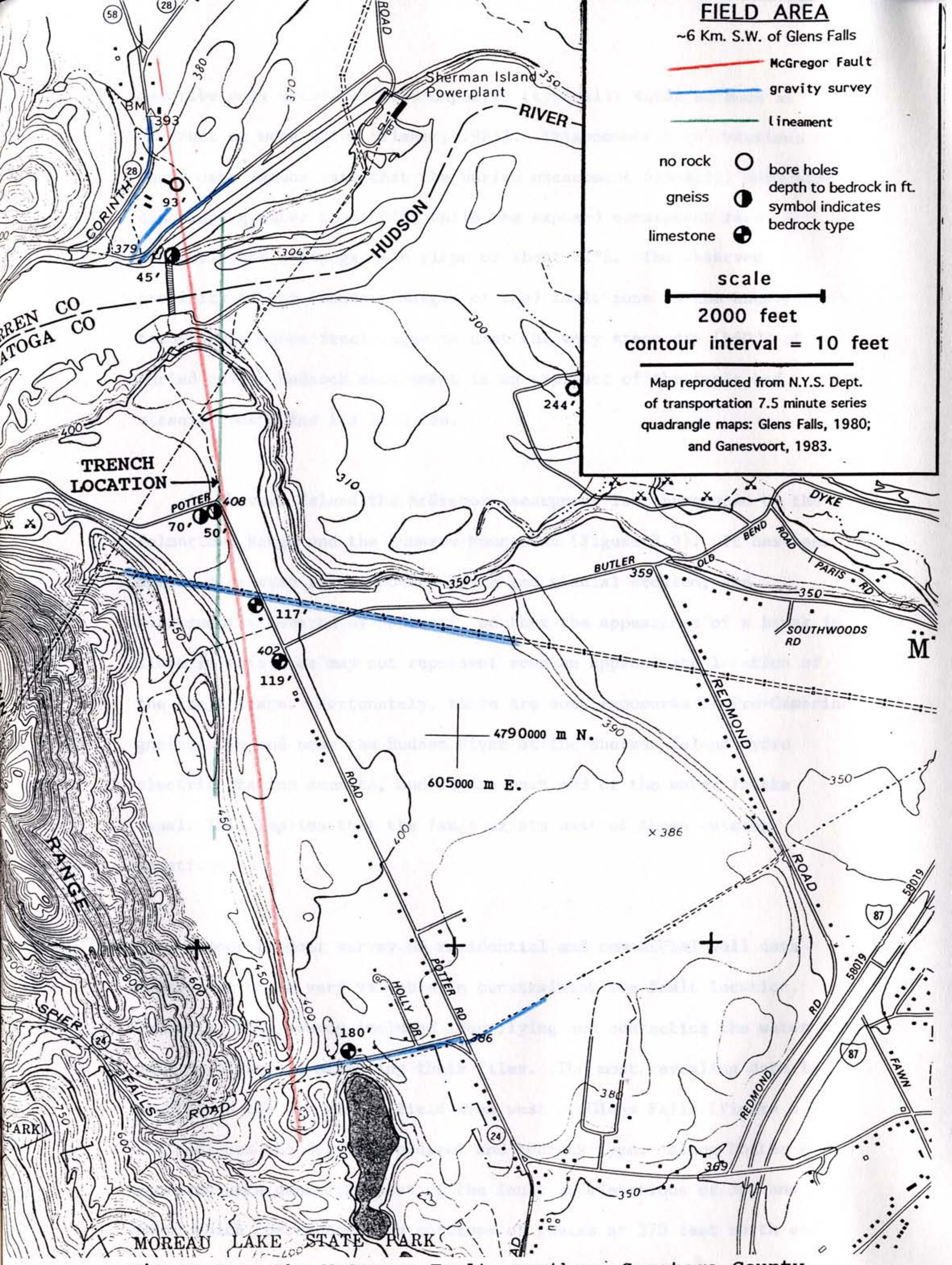


Figure 5.1 The McGregor Fault, northern Saratoga County, New York. Trench location indicated.

describe only unconsolidated deposits (typically sand) as much as 427 feet or more thick (Fisher; 1985). Trigonometric calculations with these data demonstrate that the buried escarpment (locally) must have a dip slope greater than 60°E; while the exposed escarpment face typically has a topographic slope of about 22°E. The observed proximity of the (eastern margin of the) fault zone to the escarpment's topographic slope-break suggests that the very steep dip (>60°) of the buried gneiss bedrock escarpment is an artifact of the fault and closely constrains its location.

At Sherman Island the McGregor escarpment is represented by the Palmertown Range and the Luzerne Mountains (Figure 3.9). It has been deeply dissected by the Hudson River and glacial erosion, and subsequently covered by sediment, so that the appearance of a break in slope in this area may not represent even an approximate location of the fault trace. Fortunately, there are some exposures of Pre-Cambrian gneisses in and near the Hudson River at the Sherman Island hydro electric station damsite, and at the west end of the water intake canal. This implies that the fault exists east of these outcrop locations.

A door to door survey of residential and commercial well data turned out to be very valuable in constraining the fault location. Commonly, this source included identifying and contacting the water well drillers and accessing their files. The most revealing data is plotted on the map of the field area west of Glens Falls (Figure 5.1), and includes the depth to bedrock and bedrock type. Along Potter Road, gneisses were described west of the fault at elevations of 360 and 345 feet, which correspond with outcrops of gneiss at 370 feet north and south of the Hudson River - just west of the dam (Figure 3.9). A borehole immediately north of the dam however, strikes gneiss at an elevation of 315 feet, demonstrating that some erosional retreat of the scarp has occurred. An outcrop of gneiss in the Hudson River channel east of the dam (at an elevation of 290 feet) is believed to be very close to the fault. Along Potter road well logs east of the fault describe limestone at elevations of about 380, 381, and 210 feet. I suspect that the cuttings described by the well drillers as "black-lime", a dark gray, fine grained limestone, probably represent the uppermost Glens Falls/Sugar River Limestone, or the basal Utica Shale, which is a black argillaceous limestone (these units are exposed in the Portland Cement Quarry, south of the Hudson River in Glens Falls). According to the driller, these wells were bored at least 2 feet into the limestone in order to confirm that they sampled the bedrock rather than buried glacial erratics. Boreholes still further east of the fault typically run very deep without striking bedrock at all (down to an elevation of 56 feet near Old Ben Road, for example; Figure 3.1). This evidence makes it likely that the limestone encountered represents the presence of a fault horse (a displaced sliver of rock caught between the walls of a fault) or a fault splay. Both features have been observed to be commonly associated elsewhere with this fault and are typical of steeply dipping faults with a large throw. The bedrock outcrop near the damsite and borehole data constrains the strike of the fault in the northern section of the field area to between 348 and 355 degrees, assuming it maintains a constant strike across this area.

In addition to direct field evidence and borehole logs, economical, noninvasive, geophysical techniques exist that are capable of profiling subsurface bedrock structures and/or stratigraphy. A gravity survey was chosen because the equipment is highly mobile, allows good depth of "penetration", and the technique is not labor intensive. Also, a gravimeter is not susceptible to electrical (or electromagnetic) interference from high tension transmission lines, which are prevalent in the area and make ideal ground survey routes.

The gravity method, as applied here, determines very small differences in the gravitational field of the Earth and attempts to use these observations to resolve local variations in the density of rocks near the surface. The quantity being measured and recorded is not the actual magnitude of the force of attraction, but rather the variation of that quantity at surveyed locations (stations) along the earth's surface. The principal purpose of the application in this instance is to determine the location of the maximum density gradient between Proterozoic gneiss near or at the surface (west of the fault) and the basement gneiss beneath Paleozoic shale/carbonate rocks covered by sand (east of the fault).

The gravity meter used is extremely sensitive, capable of measuring variations in acceleration of about 10^{-5} gals. (or 0.01 milligals; one gal. = 1 cm/sec), which represents a precision of one part in 10^{8} (Telford, et al, 1976). However changes in gravitational accelerations resulting from anomalous density variations (across a fault, for example) are not abrupt, but gradual and often rather small in magnitude (Figure 5.2). Corresponding gravity anomaly graphs are typically smooth profiles that cannot be interpreted unambiguously, but

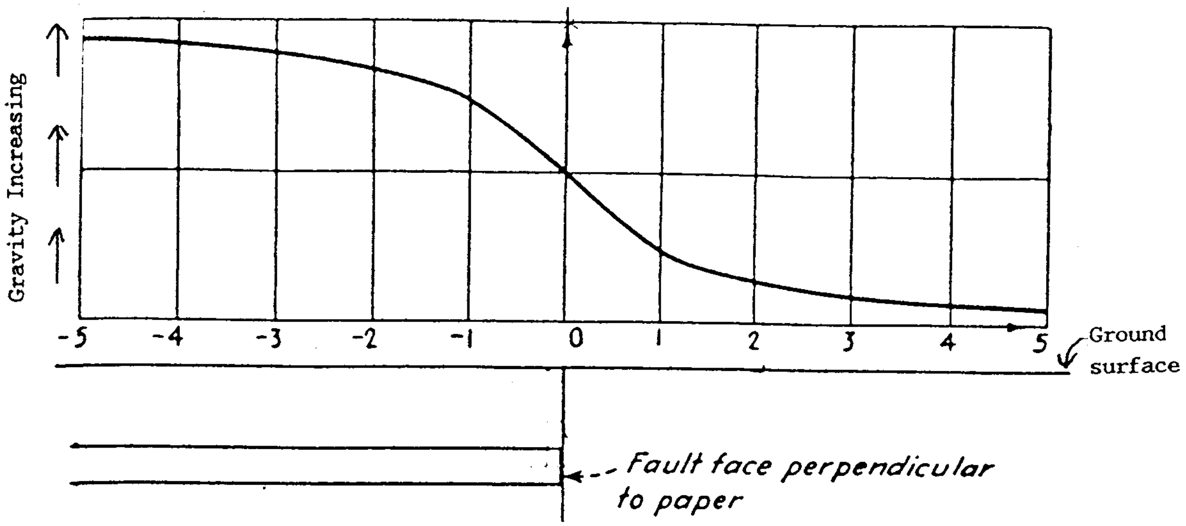


Figure 5.2 Theoretically determined gravity profile over a faulted slab, downthrown side assumed infinitely deep (Modified from Dobrin, 1976).

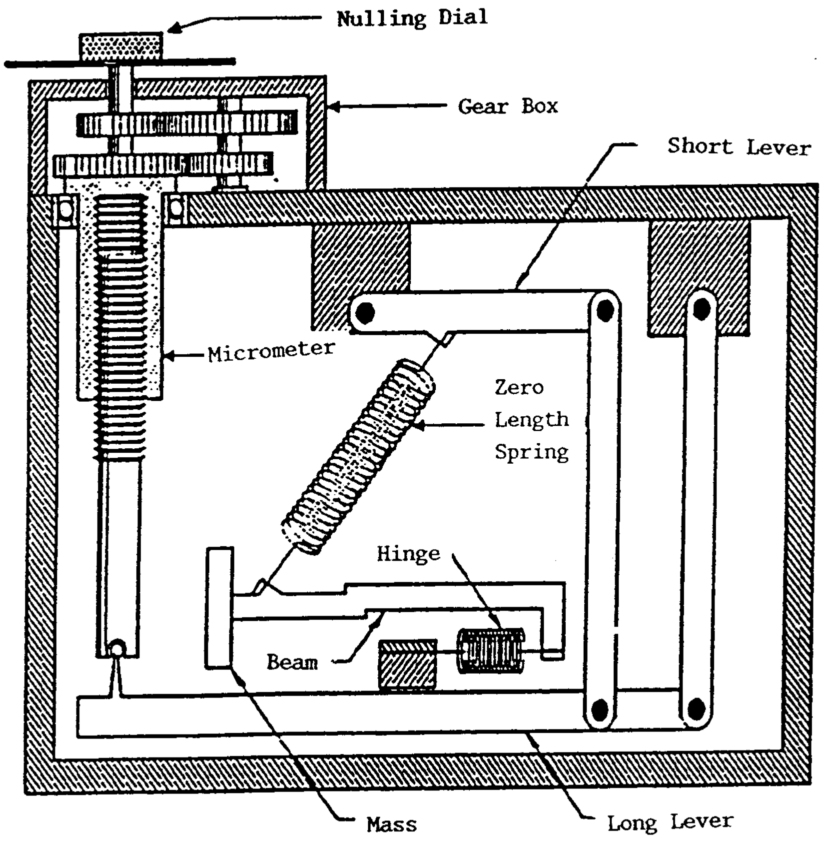


Figure 5.3 A Simplified sketch of the innards of a La Coste and Romberg Model D Gravity meter, in profile. From the Instruction Manual, 1990.

can provide valuable information on major structures at depth. The gravity meter used in this phase of the project was a LaCoste and Romberg model D. The internal components are almost entirely made of non ferrous metal parts, and the case is shielded to eliminate the influence of magnetic fields. A simple diagram of the meter (Figure 5.3) shows a mass at one end of a hinged horizontal beam. The meter is operated by adjusting the nulling dial, which adds or subtracts a small amount of force necessary to balance the mass through a network of levers and springs. Numerical readings are recorded after making a nulling adjustment at each surveyed station, and later converted and corrected to milligal units.

Gravity readings from field work will be influenced by a number of factors, and must be corrected for variations in elevation, latitude, terrain, and earth-tides. Gravity varies inversely with the square of distance (from the attracting body), so it is necesary to correct for elevation changes from station to station and reduce all readings to a common datum surface - usually mean sea level. This is called the free-air correction; it makes no allowance for the attraction of material between the station and the datum surface. The Bouguer correction does, by assuming the material is a slab of infinite horizontal extent, and of a given uniform density. When gravity is measured above the datum surface, the Bouguer correction must be subtracted from it, the free-air correction added. A terrain correction must be added to the station reading when steep topographic irregularities are encountered, because of the upward attraction of hills and/or the absence of a downward attraction due to valleys. This requires an accurate topographic map of the area. The centrifugal acceleration due to the Earth's rotation and the change in shape due to the equatorial bulge make it necessary to apply a latitude correction for any significant (as little as 40 feet; Telford, et al., 1976) north-south excursions. Therefore, the station positions must be very precisely located. A significant influence on instrumental measurements can occur due to the gravitational effect of the movements of the sun and moon, which produce Earth tides. This value can be calculated for any latitude and time. Other corrections account for instrumental drift. Drift is mainly the result of metal fatigue within the springs of the instrument (or sudden, jarring motions) and to evaluate this it is necessary to reoccupy some of the stations during a survey every couple of hours in order to produce a drift curve. A detailed mathematical description of the corrections described above is beyond the scope of this paper, but can be found in Dobrin (1976) or Telford, et al. (1976). Access to the appropriate computer software is necessary to make the computations. Programs for the analysis of observations made in a typical gravity survey require that the minimum input include for each station, location (latitude and longitude), elevation, time, and reading (Table 5.1). The elevation information was obtained by surveying the station locations to a known benchmark with a standard transit scope and stadia rod.

As discussed earlier, local density variation is the quantity relied upon to cause a gravity anomaly. A major problem with this parameter is that this variation is typically very small - less than 2 g/cc, and density can vary considerably within a given lithology (Table 5.2), making it difficult to distinguish between rocks of different types. Accordingly, small samples of the formations

Gravity Survey IN
Niagara Mohawk Canal Road-Sherman Island

4791.1334	604.1680	375.74	base station coordinates and elevation
6	gravity meter #		
92/1/18	date		
1429.47	10:15	base station reading; time	
1428.01	14:13	" " " "	

station #	lat	lon	elev	time	reading
10	4791.8200	603.7780	375.60	12:10	1447.77
12	4791.8317	603.7976	374.77	12:00	1449.58
14	4791.8317	603.8172	377.38	11:48	1448.40
15	4791.8550	603.8270	377.81	12:42	1446.48
16	4791.8667	603.8368	377.47	11:40	1445.50
17	4791.8784	603.8466	377.36	12:50	1442.89
18	4791.8900	603.8563	377.30	11:35	1442.11
19	4791.9017	603.8661	376.85	12:56	1439.48
20	4791.9134	603.8759	375.86	11:22	1439.35
21	4791.9317	603.9003	375.65	13:02	1437.37
22	4791.9500	603.9246	375.76	11:10	1437.19
23	4791.9684	603.9489	375.91	13:08	1434.84
24	4791.9867	603.9733	376.31	11:00	1434.61
26	4792.0234	604.0220	376.61	10:53	1432.69
27	4792.0417	604.0463	377.37	13:15	1430.37
28	4792.0601	604.0706	377.64	10:44	1430.62
30	4792.0968	604.1193	377.49	10:36	1429.31
32	**4792.1334**	**604.1680**	**375.74**	**10:15**	**1429.47**
33	4792.1518	604.1923	373.29	13:23	1429.04
34	4792.1885	604.2410	371.55	13:33	1429.72
35	4792.2251	604.2897	370.20	13:40	1430.40
36	4792.2618	604.3384	369.21	13:47	1430.69
37	4792.2985	604.3871	368.29	13:53	1430.86
39	4792.4085	604.3871	368.49	14:01	1430.45
148	4792.0928	603.7393	376.75	12:22	1448.13
32	**4792.1334**	**604.1680**	**375.74**	**14:13**	**1428.01**

Raw data obtained in a gravity survey of the road adjacent to the Niagara Mohawk canal at the Sherman Island Hydroelectric Generating Facility on the Hudson River, west of West Glens Falls, N.Y.

Table 5.2

AVERAGE DENSITIES OF METAMORPHIC ROCKS

Rock	Number of samples	Mean density, g/cm³	Range of density, g/cm³
Gneiss, Chester, Vermont	7	2.69	2.66 –2.73
Granite gneiss, Hohe Tauern, Austria	19	2.61	2.59 –2.63
Gneiss, Grenville, Adirondack Mts., New York	25	2.84	2.70 –3.06
Oligoclase gneiss, Middle Haddam area, Connecticut	28	2.67	
Quartz-mica schists, Littleton Formation, New Hampshire (high-grade metamorphism)	76	2.82	2.70 –2.96
Muscovite-biotite schist, Middle Haddam area, Connecticut	32	2.76	
Staurolite-garnet and biotite-muscovite schists, Middle Haddam area, Connecticut	22	2.76	
Chlorite-sericite schists, Vermont	50	2.82	2.73 –3.03
Slate, Taconic sequence, Vermont	17	2.81	2.72 –2.84
Amphibolite, New Hampshire and Vermont	13	2.99	2.79 –3.14
Granulite, Lapland:			
Hypersthene-bearing	7	2.93	2.67 –3.10
Hypersthene-free	5	2.73	2.63 –2.85
Eclogite	10	3.392	3.338–3.452

From Clark, 1966

Table 5.3

Densities of sediments and sedimentary rocks

Rock type	Range (g/cm³)	Average (wet) (g/cm³)	Range (g/cm³)	Average (dry) (g/cm³)
Alluvium	1·96–2·0	1·98	1·5–1·6	1·54
Clays	1·63–2·6	2·21	1·3–2·4	1·70
Glacial drift	—	1·80	—	—
Gravels	1·7–2·4	2·0	1·4–2·2	1·95
Loess	1·4–1·93	1·64	0·75–1·6	1·20
Sand	1·7–2·3	2·0	1·4–1·8	1·60
Sands and clays	1·7–2·5	2·1	—	—
Silt	1·8–2·2	1·93	1·2–1·8	1·43
Soils	1·2–2·4	1·92	1·0–2·0	1·46
Sandstones	1·61–2·76	2·35	1·6–2·68	2·24
Shales	1·77–3·2	2·40	1·56–3·2	2·10
Limestones	1·93–2·90	2·55	1·74–2·76	2·11
Dolomite	2·28–2·90	2·70	2·04–2·54	2·30

From Telford, et. al., 1976

encountered in the field were obtained in order to make laboratory measurements of their density. Unfortunately, these results may not necessarily provide a true bulk density of a material as samples may be weathered, fragmented, dehydrated, or mineralogically non-representative. Nevertheless, it is important to constrain formation densities in a particular field area as closely as possible. Samples of the Adirondack basement from the field area included biotite, hornblende, hypersthene, augite, plagioclase granulites; and quartz, biotite sillimanite, plagioclase, garnet gneisses (the Dresden Granulite and the Hague Gneiss respectively; Fisher, et al.; 1985). They have respective average densities of 2.97 and 2.79 g/cc, within the range of values determined by Clark (1966; Table 5.2). Borehole data indicates that the overburden is largely composed of sand. Samples recovered at different depths (courtesy of Empire Soils Investigations and the Town of Queensbury Water Filtration Plant) ranging from 35 to 80 feet yielded densities between 1.49 and 1.62 dry, and 1.85 to 1.99 wet, (a value of 1.9 g/cc was used) within the range of values from Telford et al. (1976; Table 5.3). Samples were tightly packed to simulate compaction at depth. It is assumed that the deep sand covers Ordovician Utica Shales or Trenton group limestones because immediately east of the fault, slivers (horses) of shale (in outcrop) or black limestone (from bore hole data) exist; and outcrops of Glens Falls limestone occur about 8 km ENE of the field area. Also, deep boreholes have located shale or limestone bedrock 5 km east of the Sherman Island damsite (Fisher, 1985), and 'Carbonate' rock (probably limestone) about 4.5 km to the southeast (Heath, et al., 1963; U.S.G.S. files). Shale and limestone have average saturated densities of 2.40 and 2.55 respectively.

Given a general idea of the location of the fault using outcrops, topography, borehole data, aerial photographs, and bedrock map information, the gravity survey routes were planned. The traverses were intended to be oriented as perpendicular to the inferred fault trace as possible, to ensure a sharp relief signal. These routes are roughly east-west and afford the added benefit of keeping latitude corrections to a minimum. The traverses are located along the 1) Niagara Mohawk Sherman Island Powerplant Canal Road and Corinth Road (NM Grav); 2) Overhead powerline trending N 100°E which intersects with Potter and Butler Roads (Pott Grav); and 3) Spier Falls Road (Spiegrav; see map, Figure 5.1). The gravity data from these traverses was supplemented by several other data points, and plotted using U.T.M. coordinates from N.Y.S. D.O.T. 7.5 minute quadrange maps (Figure 5.4). A gravity anomaly was determined (in milligals) for each point, based on corrections (Bouguer; free-air; etc.) of the measured gravity values. The anomaly calculation also included a correction for those portions of bedrock covered by overburden assuming a constant density contrast of 0.9 (density of gneiss 2.8 minus density of overburden 1.9). This assumes that the gravity contribution of carbonate bedrock (and possibly shale also) which overlies the basement on the downthrown fault block does not significantly alter the overall gravity signal. These points were then sequentially connected by a straight line (Figure 5.5); the effect being that points of roughly equal value are contoured by a series of crudely parallel lines. Profiles oriented perpendicular to these "contour" lines will reveal the steepest gravity gradient. A single profile of all the data points showed a great deal of scatter which suggested that the fault is curved or bent. Separate profiles for the northern and southern sections of the field area

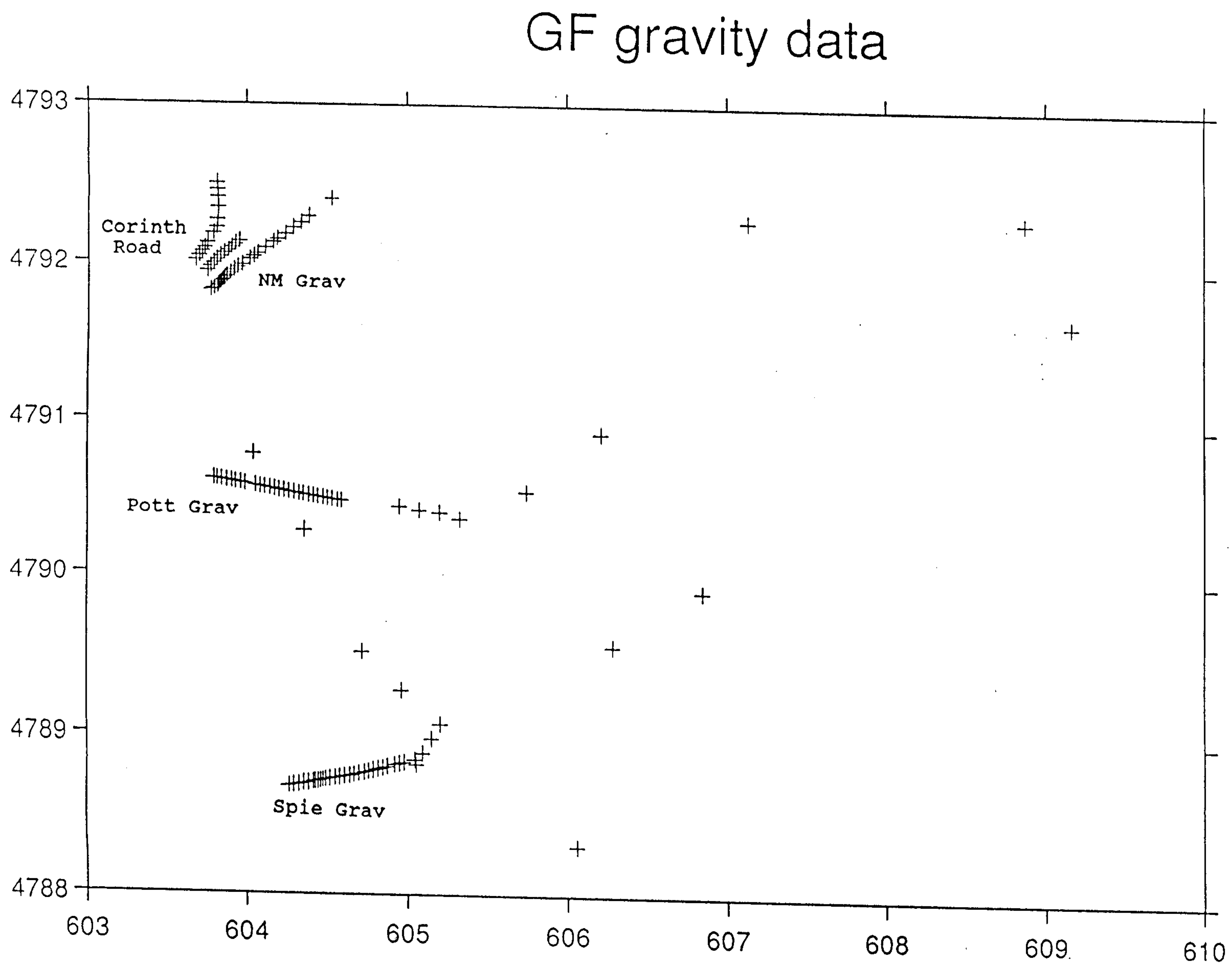


Figure 5.4 Map of Glens Falls Gravity data station locations in the field area, (refer to figure 5.1) including traverse names. X and Y axes are UTM coordinates.

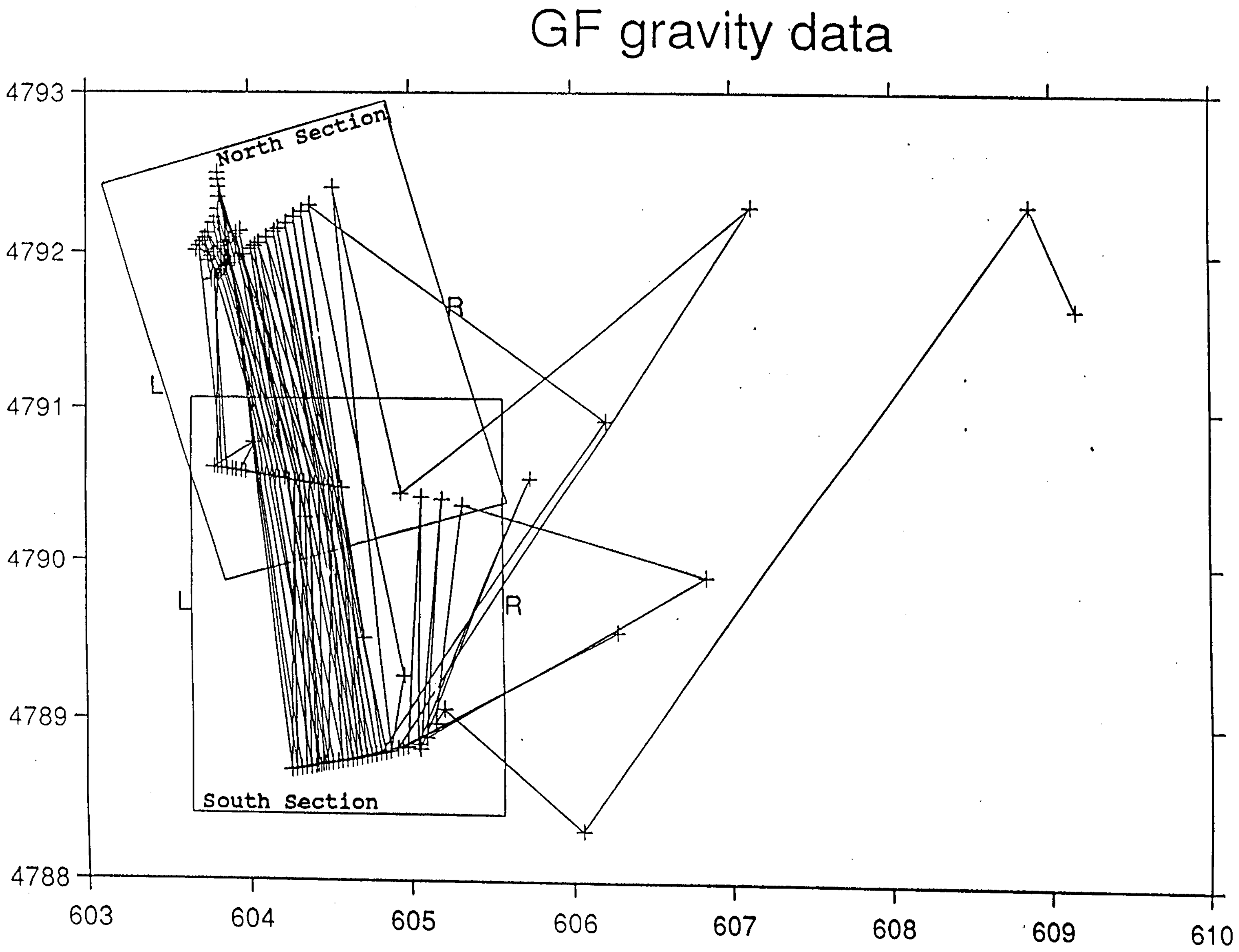


Figure 5.5
Gravity anomaly "contour" map. Points of roughly equal gravity values are connected by straight lines. It is interpreted that the contours roughly parallel the buried bedrock slope, and assuming the McGregor Fault is at the base of that slope (the slope is fault controlled), that the orientation of the McGregor Fault changes direction.

(Figures 5.6 and 5.7) each had very little scatter and displayed a very similar curve shape. This suggests that while the fault dip and bed thicknesses remain the same, the orientation of the fault strike has changed from approximately due north in the southern section to about 344° in the northern section.

In theoretical calculations the gravity anomaly curve across a fault always consists of a trough and a peak, however, in practice either the trough or peak is small and not easy to recognize (Geldart, 1966). In the general relationship between the anomaly and the fault geometry, the negative anomaly is always over the downthrown side (for faults with less dense material on the downthrown side). For a reverse fault the amplitude of the positive anomaly is larger than that of the negative, the opposite being the case for a normal fault (Figure 5.8). Curiously, the data from the gravity profiles in figures 5.6 and 5.7 are consistent with the curve shape and anomaly magnitudes that would be expected for a westward dipping high angle reverse fault (curve CD, for example in Figure 5.8). However, drawing that conclusion from such a small local data set would be premature, considering the large scale regional structure. Two data points 4 km east of the field area show the negative anomaly continues to decrease up to -2.3 milligals. The gravity signiture further west of the McGregor is unknown. I am inclined to be more comfortable with the interpretation in figure 5.9 which suggests that a sequence of normal faults, (and fault horses accompanied by drag folding) may explain the precipitous initial decrease in gravity (from W-E) followed by a very gradual decrease. The effect would be to break up the negative portion of the anomalous curve C'D' from figure 5.8 and stretch it out toward the east. While

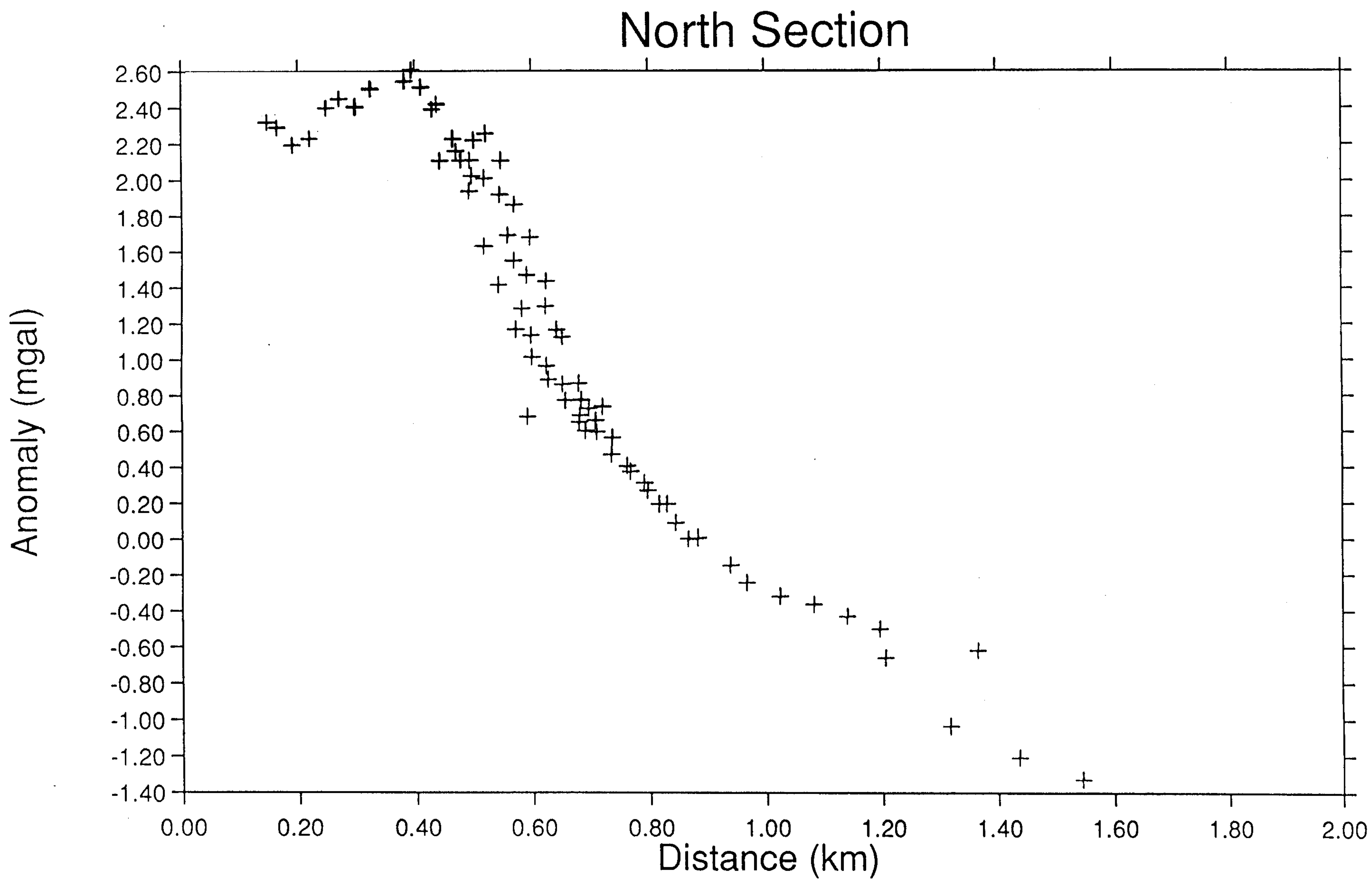


Figure 5.6 Gravity profile of the northern section of the field area, oriented perpendicularly to the long axis (general countour trend) of the northern section as shown in figure 5.5, looking north. This profile orientation shows minimal data scatter and is the steepest curve. It crudely represents the buried bedrock slope.

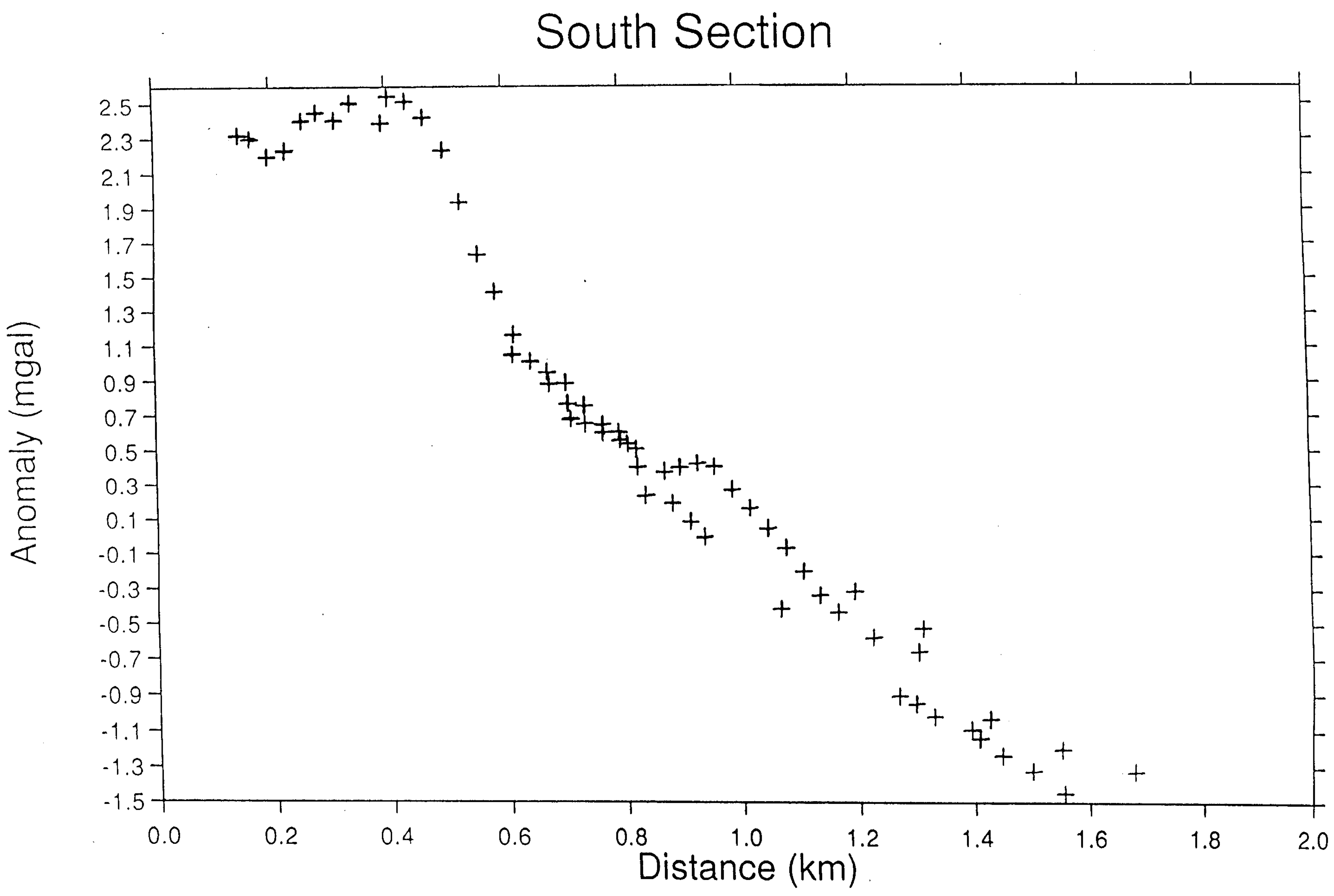


Figure 5.7 Gravity profile of the southern portion of the field area (shown in map view in figure 5.5) looking north. This profile orientation shows minimal data scatter and the steepest curve. If all data points from both the southern and northern section (figure 5.6) are plotted together, the data is difficult to resolve.

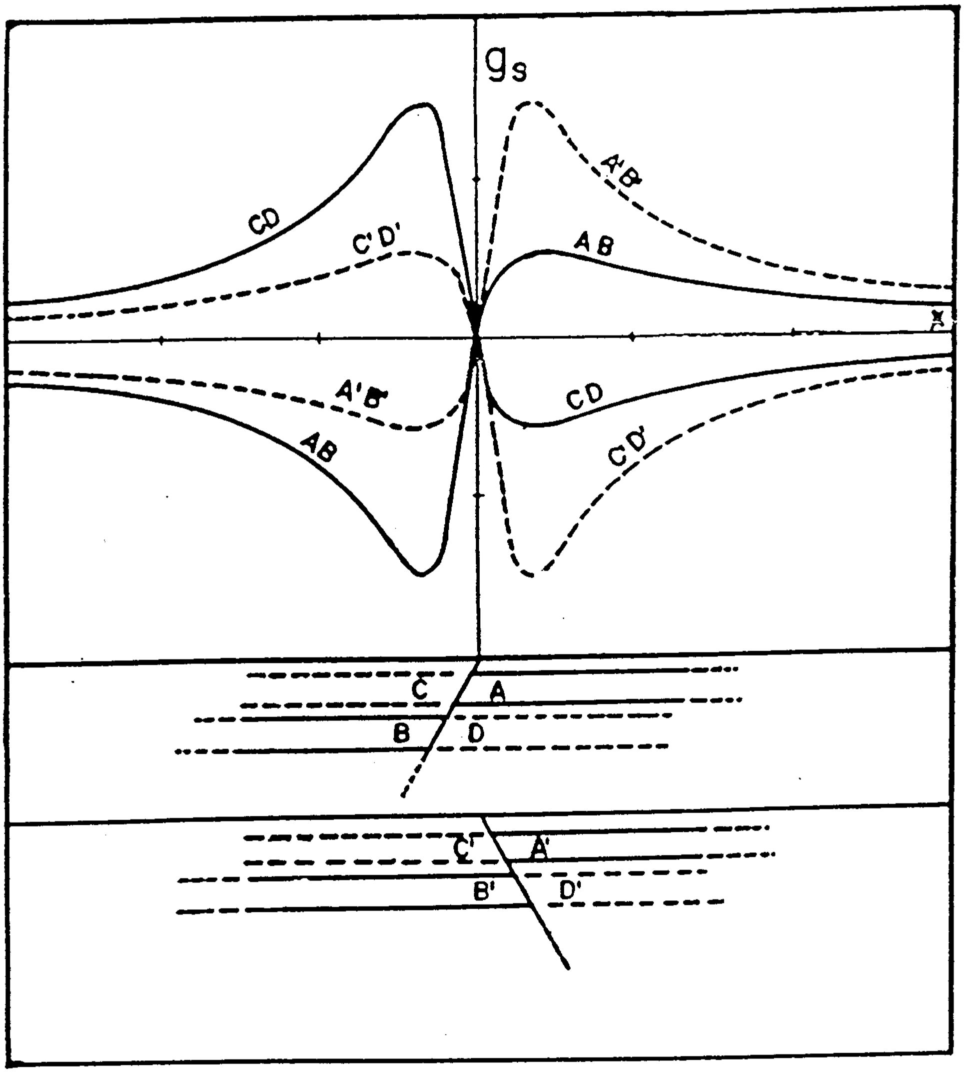


Figure 5.8 The theoretical relationship between the gravity anomaly curve and the type of faulting (the anomaly due to a fault-cutting basement is equivalent if one assumes that the fault continues downward without change for great depths). The vertical axis is the anomaly in milligals, and the horizontal axis is a distance X. Fault dip is 60 degrees. A vertical fault would result in a symmetrical curve. (Modified from Geldart, et al., 1966).

Figure 5.9A and B. A profile of gravity survey data from the southern section of the field area (A) and an interpretation of possible bedrock/fault/overburden relations (B). Note that the McGregor Fault is assumed to be normal, dipping 76°E. The profile is oriented approximately 74° (roughly WSW-ENE; perpendicular to the gravity "contours" shown in figure 5.5) so that the projection of the depicted fault up to the ground surface (at ~0.55 km distant) is roughly coincident with the McGregor Fault near the 117′ borehole as drawn on figure 5.1. The basement is Adirondack Gneiss, which is exposed in the Palmertown Range to the west, followed upsection by Potsdam Sandstone, Beekmantown Group Dolostones, and finally Black River and Trenton Group Limestones (from column 4, figure 3.2, and borehole data). It is possible that basal Utica Shale may be present, but it is not represented in the diagram (B).

The depth to bedrock values for the gravity curve are approximate, calculated with a formula from Dobrin et al. (1976; note that they are vertically exaggerated ~6X with respect to the horizontal distance). The values crudely agree with borehole data from figure 5.1 and the Glens Falls area which suggest that the deepest bedrock occurs approximately at sea level (Fisher, 1984; for example, the deep borehole north of Butler Road; adjusted to ~1.8 km distant on the profile was drilled to an elevation of 56′ never striking bedrock). The gravity data suggests that the deepest bedrock occurs at a depth of ~30′ below sea level.

The trough in the gravity signal at 0.2 km occurs over a kettle, and may be the result of the influence of higher density glacial drift observed there, laden with gneissic cobbles and boulders, and possibly a buried bedrock stream channel which may have accomodated syn-glacial meltwaters (Figure 3.7). Two boreholes drilled deep into gneissic bedrock along Potter Road support the latter speculation.

The main fault has been inferred to exist at the break in slope of the buried bedrock escarpment, denoted where the pronounced break in slope occurs in the gravity curve. East of the fault the curve appears to require an inordinately long distance to level off. This effect is probably a result of the vertical exaggeration of the depth to bedrock values (the calculated bedrock slope between 1.3 and 1.6 km is actually <4°), and the persistant gravitational influence of the massive root of the Adirondack Palmertown Range (as opposed to the effect of a "two dimensional" faulted slab model; figure 5.8).

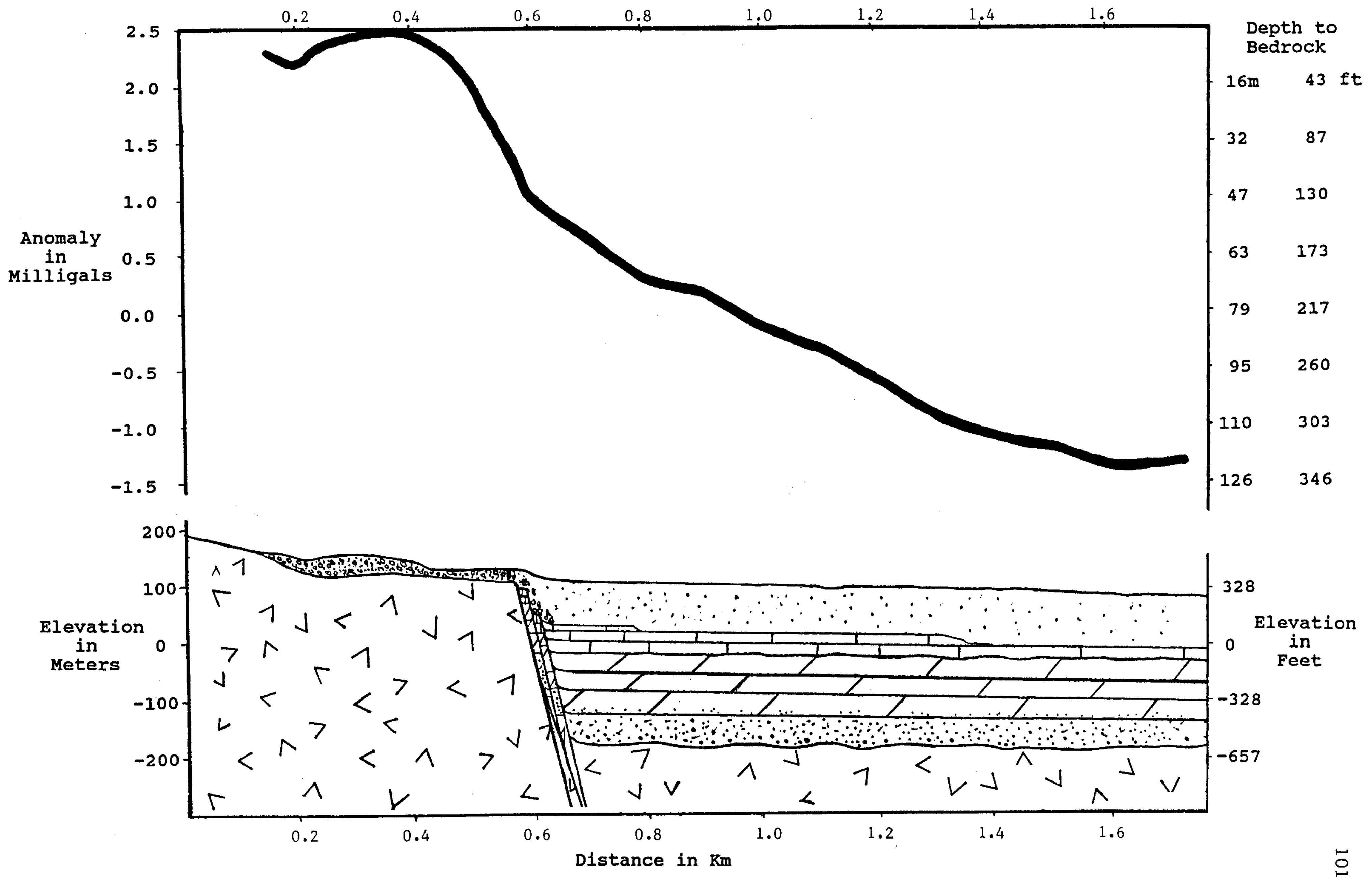


Figure 5.9A and B

this is speculative it better explains the absence of any recovery of the negative anomaly (consistent change in the positive direction) over large distances east of the fault. Also, figure 5.9 represents a more realistic interpretation of the gravity curve observed considering that this is a faulted basement relationship as opposed to the faulted slab model calculated by Geldart (et al., 1966).

The most striking feature of the profiles is the precipitous change in slope as the inferred fault is approached from the west, preceeded by an increase in the positive gravity anomaly of up to 0.43 mg (to a maximum of 2.54 mg) higher than values of 2.11 mg that are typically encountered on surface outcrops of gneissic bedrock west of the fault. The profiles of the survey traverses "Potter Grav", "Spiegrav", and "NM Grav" can be used to approximate the main fault location (Figures 5.10, 5.11 and 5.12). The first inflexion point at the base of the steep portion of the curves was interpreted to represent the westernmost edge of the fault zone (marked by X's on the profiles and figure 5.1). It is acknowledged that this determination may involve a large error. Nevertheless, these points correlate reasonably well with the location of the fault as determined by borehole and other data. I believe that the steep portions of the profiles represent the eastern-most limit of the buried gneissic bedrock escarpment of the Palmertown Range. I assume that the McGregor fault exists at the base of, or very near the eastern edge of this buried escarpment, as is observed to be the case with outcrop-constrained exposures of the fault zone south of this area. Note that while none of the profiles show a particularly smooth transition from steeply to gently dipping slope, they often suggest that an erratic,

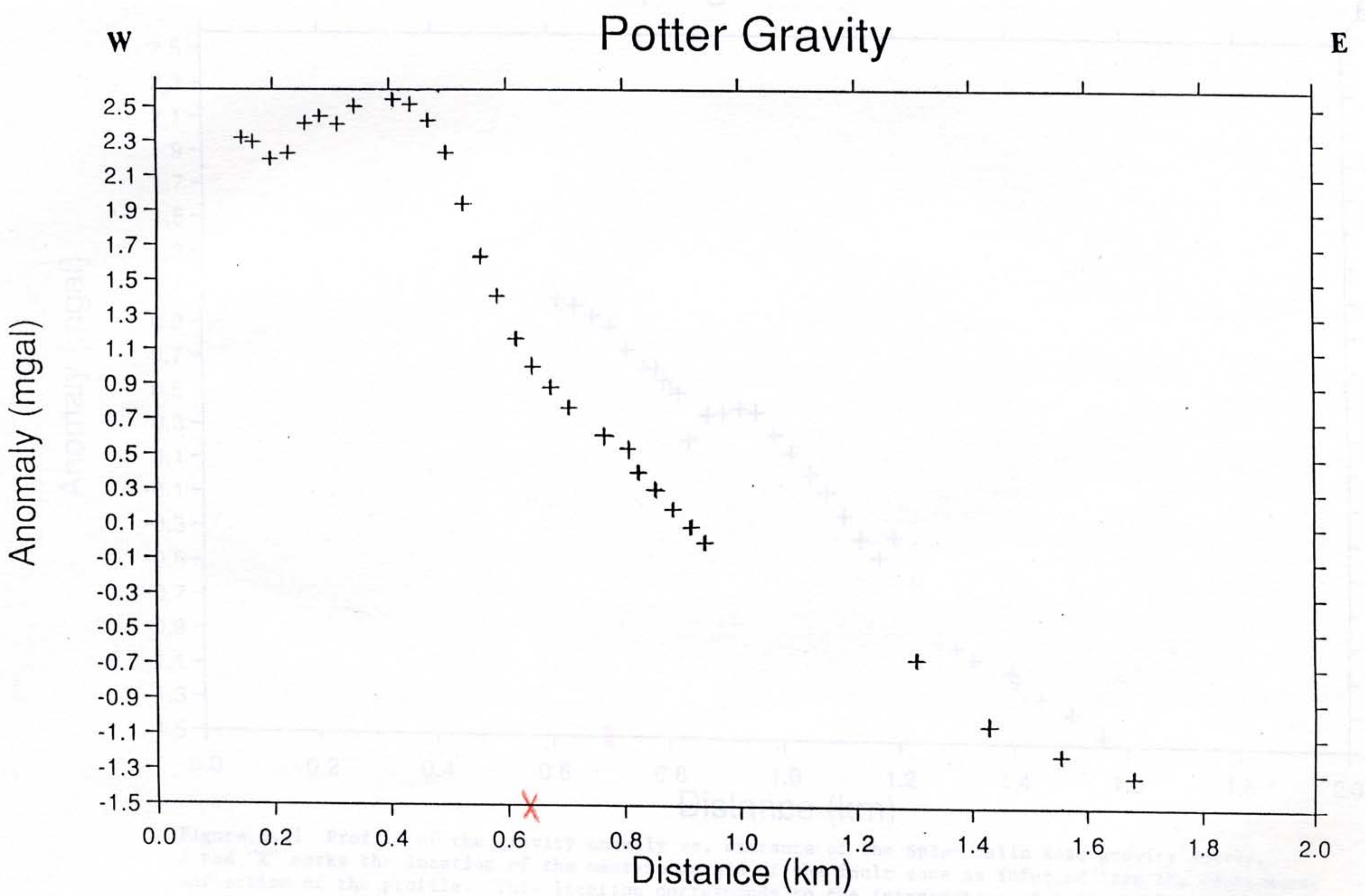


Figure 5.10 Profile of the corrected gravity anomaly vs. distance for the survey data obtained on the traverse along the power line intersecting with Potter and Butler Roads (see map Figure 5.1). A red "X" marks the location of the western margin of the fault zone, inferred from the change in slope (inflection pc int) of the profile. On the map this location corresponds to the intersection of Potter Road with the power line.

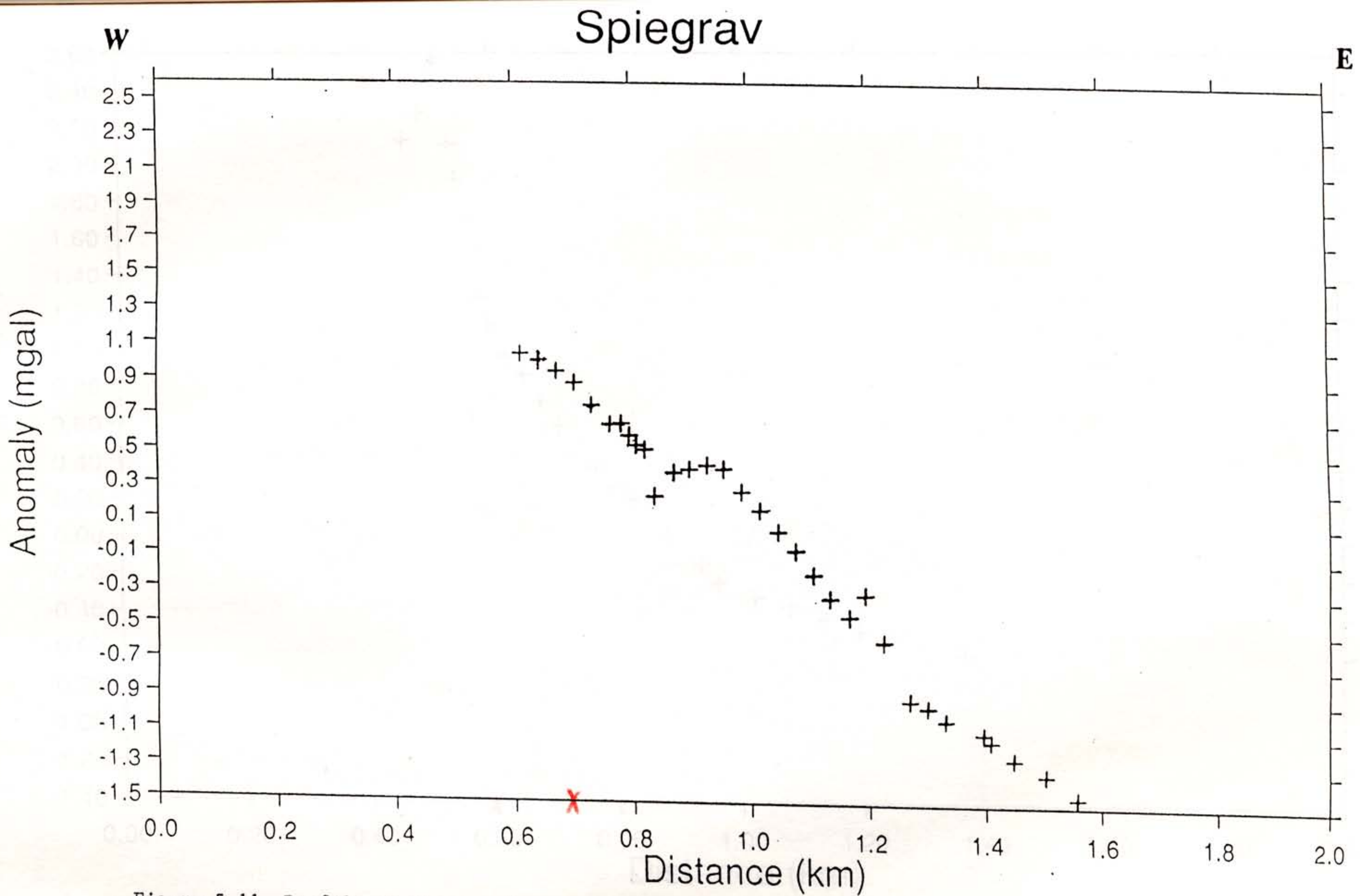


Figure 5.11 Profile of the gravity anomaly vs. distance of the Spier falls Road gravity survey. A red "X" marks the location of the western margin of the fault zone as inferred from the westernmost inflection of the profile. This location corresponds to the intersection of Spier Falls Road with the line delineating the McGregor Fault on the map, figure 5.1.

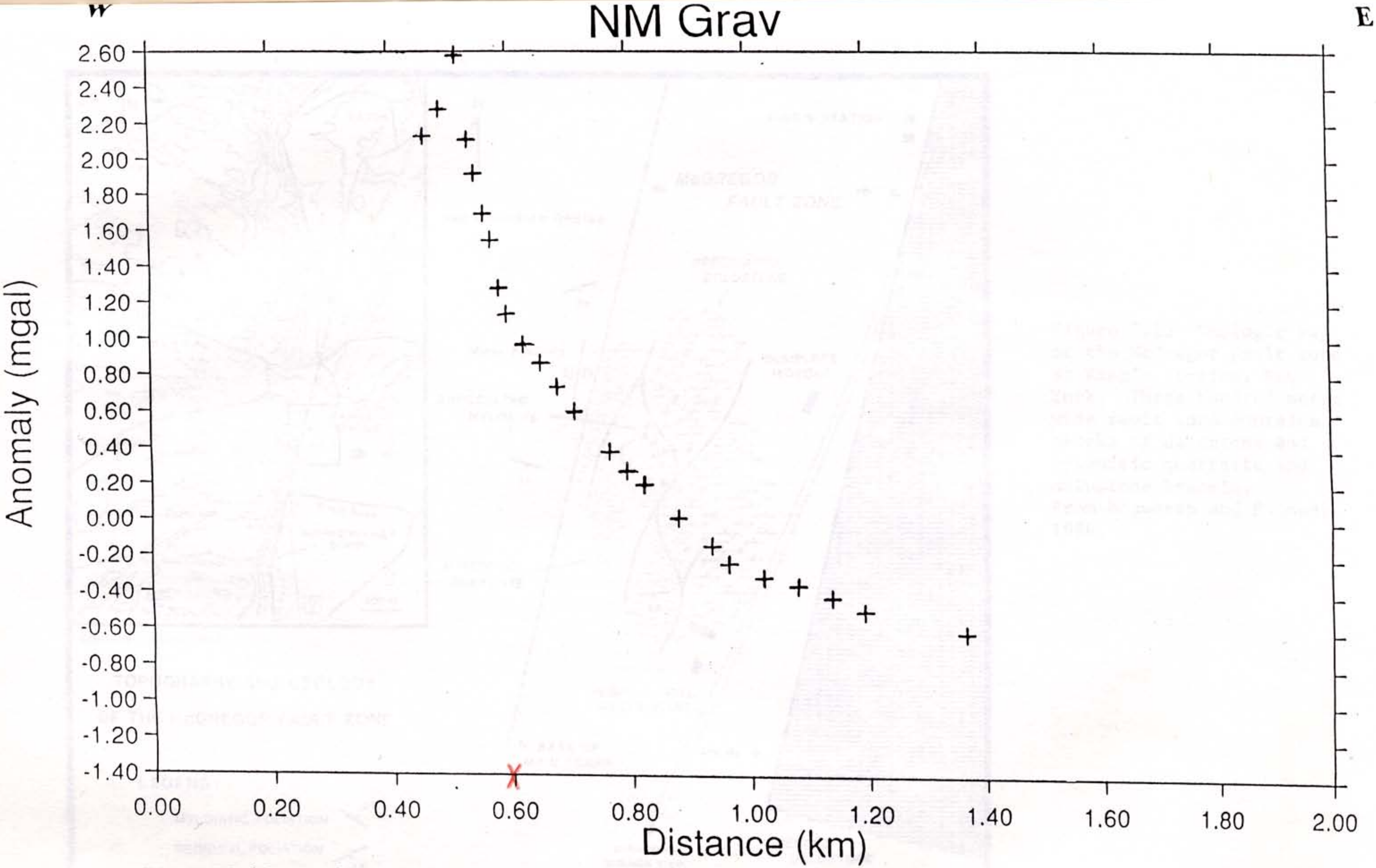


Figure 5.12 Profile of the gravity anomaly vs. distance of the Niagara Mohawk Canal Road gravity survey at the Sherman Island hydroelectric power generating facility on the Hudson River. A red "X" marks the location of the western margin of the McGregor Fault zone, inferred from the westernmost change in slope of the profile. This location on the surface is found at the intersection of the Canal Road with the McGregor Fault as drawn on the map, figure 5.1.

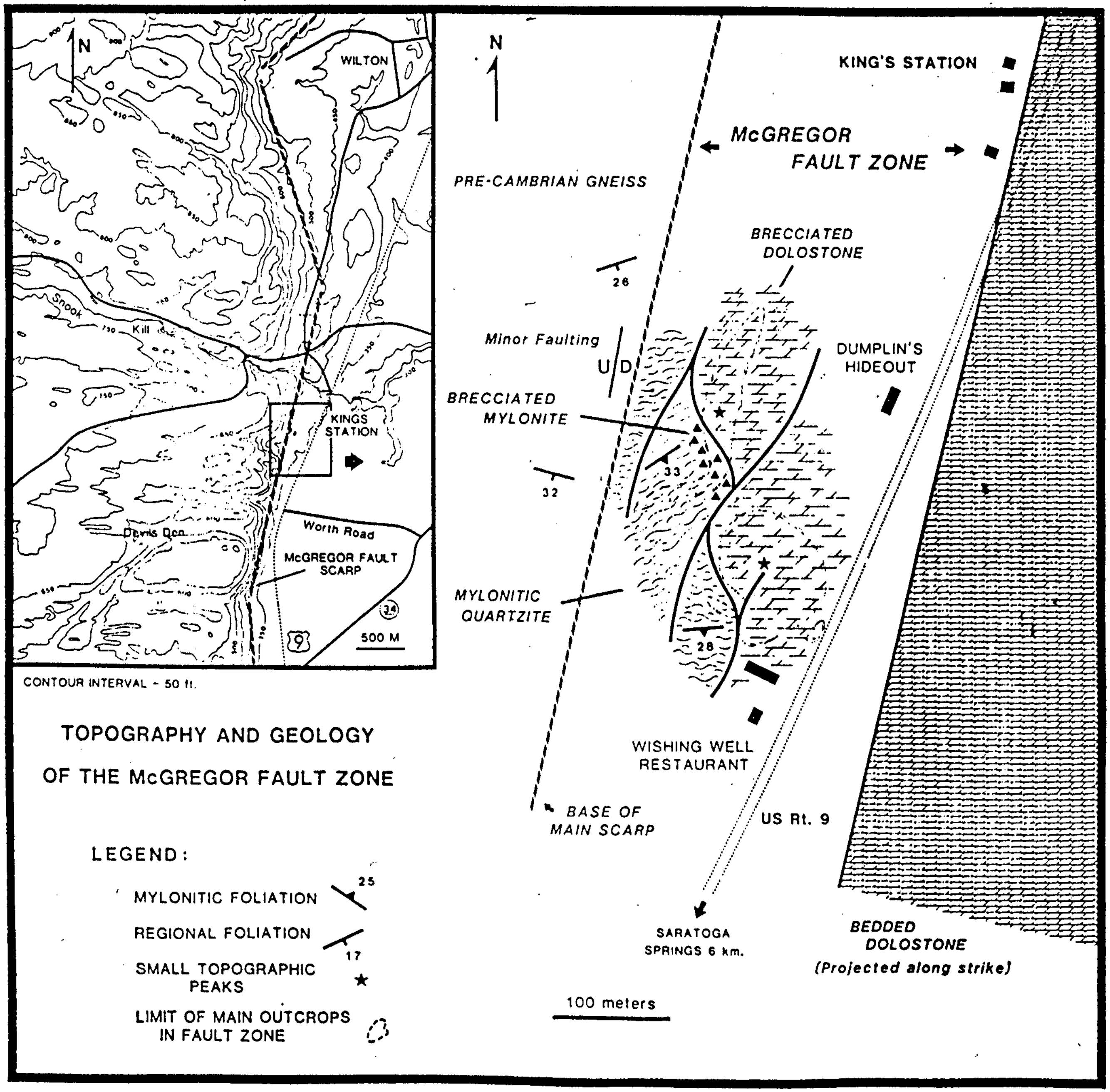


Figure 5.13 Geologic map of the McGregor Fault zone at King's Station, New York. Three hundred meter wide fault zone contains blocks of dolostone and mylonitic quartzite and dolostone breccia. From Bosworth and Putman, 1986.

step-like transition exists; especially the traverses of the southern section (Figures 5.10 and 5.11) which show an erratic pattern east of the western margin of the fault zone (fault margin inferred from the borehole and gravity data). The gravity profiles of the southern section also tend not to "flatten out" toward the east as much as the northern "NM Grav" traverse does (Figure 5.12). This suggests that the fault zone in the southern section of the field area may be wider than usual (10-20m; Geraghty and Isachsen, 1980). This may be an effect of the presence of horses or fault slivers within (or perhaps a fault splay intersecting with) the fault zone. A complex, wide fault zone (about 300 m) was mapped by Bosworth and Putman (1986; Figure 5.13) at Kings Station, near the intersection of the McGregor Fault and the Gurnspring Fault, which forms the southern margin of the Wilton Horst (Figure 1.3).

My original intention during this phase of the investigation was to constrain the precise location of the (buried) McGregor Fault to within +/- 15m (the anticipated length of the trench being 30m). However, the fault exists as, and is part of, a fault zone which may be as much as 300 m wide. I can confidently conclude that the fault contact between the Adirondack gneiss and carbonate rock represents the <u>western</u> margin of the fault zone. However, assumptions regarding where (along which margin or splay) the most recent movement has occurred within a wide fault zone would be no more than speculation without more direct supportive evidence. Since the western margin of the fault zone has been located with confidence, it represents the best initial opportunity for finding such evidence using paleoseismic trenching techniques.

6.0 DESCRIPTION OF THE TRENCH SITE AND OTHER EXCAVATIONS IN THE FIELD AREA

The excavated trench was located at the intersection of the lineament seen from aerial photographs and the fault trace inferred from borehole and gravity survey information (see Figure 5.1). The local topography is very level, and vegetation consists of second growth hard and softwood trees.

The trench was oriented at about 070°, or as nearly perpendicular to the strike of the inferred fault zone as possible. The dimensions of the trench were approximately 35 m long by 3.2 m deep (Figure 6.1 and 6.2). Bedrock is estimated to be at a depth of more than 22 meters, based on the borehole and gravity surveys. The overburden record, from the surface downward, began with a thin .15 m black soil horizon, rich in organics and roots, with some insect eggs. From .15 to .37 m was a fine to medium brown/red sand, well sorted. From .37 to 1.0 - 1.2 m, a massive bedded medium to coarse brown/red sand and gravel was found (thickness varied from about .63 to .83 m). An erosional unconformity exists between this and the unit below, a well sorted medium to coarse light gray sand with trace pebbles (Figures 6.2; 6.3; 6.4). The light gray sand unit ccntained well developed crossbedding (emphasized by prominent thin black magnetite rich sand layers) and extended to just over 3.2 m depth, the effective limit of the reach of the back-hoe for the trench width. Attempts to dig deeper encountered great resistance but occasionally the machine was able to remove well-rounded cobbles and one small boulder from the trench bottom, which was probed but not

Figure 6.1 Trench was 3.2 m deep by about 35 m long. Cobble in foreground is 24 cm long.

Figure 6.2 Prominent erosional unconformity @ 1.2 m depth between medium brown and light gray units.

Figure 6.3 Close up of erosional unconformity. Gray unit is very well sorted, cross-bedded sands, above is massive bedded layer of poorly sorted coarse sand and pebbles. Quarter for scale (bottom, center).

Figure 6.4 Interbedding pattern is part of a cut-and-fill structure. Probably the result of reworking of the lower unit. Quarter for scale (center, right).

further sampled or uncovered because of the collapse of sand from the trench walls (Figure 6.1). The consistent difficulty in digging suggests that the layer just at 3.2 m depth was composed largely of cobbles and small boulders.

The environment of deposition represented by this sequence suggests that a moderate flowing, (meandering) stream deposited the fairly well sorted light gray sands above poorly sorted cobbles and boulders. The brown/red unit was then deposited in a moderate to high energy environment, and most likely came from a different source. These fluvially deposited units may be glacial outwash or a braided stream complex laid down in a continuously changing pattern of flow. The well sorted fine sands above are probably aeolian and identical deposits of this "dune" sand (remanents of beaches on the shores of Glacial Lake Albany) occur about 3 km to the south in thick (up to 10 meters) commercial sand pits.

No evidence of syn- or post-depositional stratigraphic offset due to faulting effects was found. Occasionally, a discreet layer within the cross-bedded light gray sand unit displayed offset due to dewatering mechanisms, but this was on a very small scale (less than 3 cm) and not continuous downsection. Other complications exist at the erosional boundary between the light gray and coarser brown/red unit (Figure 6.3), which can locally show a complex interbedding pattern, where the upper bed appears to have been interjected into the lower unit (Figure 6.4). However this feature can be explained as a fluvial reworking of the lower unit, redeposited as a scour and fill structure during a temporary interruption of the deposition of the brown/red unit above. Evidence of liquefaction effects (due to earthquake shaking) were also not present. There are two basic explanations for the absence of such effects; 1) either there has been no significant earthquake activity at or near the site (within the time period constrained by the age of the upper 3.2 m of excavated overburden), or 2) the overburden present is poorly suited to record such activity, if it did occur.

In order to test the first explanation it is requisite to determine whether the second is satisfied. Stratification is present, therefore, offset of layers due to earthquake movement can potentially be recorded. However, the major limitation of this possibility is that the excavation must be located directly over that portion of the fault zone where bedrock offset has occurred. Another problem is that displacement effects propagating upward through the overburden will change direction and dissipate with increasing distance upsection, making it unlikely that small events will be evident at the top of a thick section (this situation is typical of this study - stratified sediment near the fault only occurs over deeply buried (>22m) bedrock). As discussed earlier, near surface intraplate earthquakes that involve surface offset (rupture) are historically rare. Also, only the western margin of the McGregor Fault zone was located with confidence, and it is possible that the fault zone could be significantly wider than the trench length (up to 300m, see chapter 5). This places a severe limitation on any success of locating evidence of paleo-offset (rupture) of stratified sediments by trenching (if such offset has occurred). Evidence of liquefaction effects also require that some initial stratification be present. But to find such evidence it is only necessary to excavate near an ancient (moderate to large magnitude) earthquake epicenter in sediments which have a good liquefaction potential. Factors that enhance liquefiability in a layer are a large compressibility (highly contractile sands, for example) and thickness (which increases the total volume of water available for upward flow) of a highly permeable saturated sand or fine sand (Dobry, 1989). While it has been noted that liquefaction is well developed only when this layer is confined (capped) by a layer of low permeability such as silt or clay, liquefaction can and does occur without such a cap (Dobry, pers. communication). The sediment encountered was not capped by a low permeability layer, and likely to have been unsaturated during much of its post depositional history. Unfortunately, these are not ideal conditions for recording disturbances due to liquefaction.

Liquefaction effects are not necessarily a reflection of a localized distribution of shaking near a shallow epicentral source, and often occur over large areas (20-50 km, Tuttle and Seeber, 1989; Amick and Gelinas, 1991; Obermeier, et al., 1991). Therefore, several sand pits and other exposures up to 7 km from the McGregor fault (a few are located in Figure 3.9), were examined. None revealed definitive liquefaction features. The sediment encountered included very fine sands to pebbly sands, with occasional thin gravel layers - but no significant finer fraction. These were typically bottomset lake sediments (after Connally, 1973), extensive deposits of glacial lakes Albany, Quaker Springs or Coveville. These deposits are likely to have existed under saturated conditions for a minimum of 2,000 years (La Fleur, pers. comm.), to perhaps twice that span at some locations. Of the sand pits visited only one was commercially active and allowed an excellent view of sediments in profile. This was the Hudson Valley Sand and Stone Company on Route 9 just east of Moreau Lake State Park (about 4.5 km south of the trench site and 1.5 km east of the McGregor Fault). Over 24 m of well sorted sands are exposed. At an elevation of about 350 feet, the nearly horizontal layers appear to represent the bottomset beds deposited in advance of a delta which was prograding into glacial Lake Albany (or possibly a kame delta related to and/or reworked from the Moreau Pond Kame Terrace; Figure 3.7). The uppermost 18 m are crossbedded medium sands, and a few 3 cm thick gravel cross-beds appear within the top 2 m, each grading upward to medium sand layers 12-18 cm thick which are unconformably overlain by another sequence. The graded sequences are slightly more inclined than the medium sands below, tentatively suggesting the arrival of foreset beds. The lower sediments (from 0-6 m) typically contain 8-20 cm thick cross-bedded sequences of fine and medium sand layers. Cut and fill structures are common (Figure 6.5). Near the bottom of the pit a 1.1 m thick sequence of these layers contains diapiritic de-watering structures that are quite extensive laterally (about 20 meters) and occur wherever the units are exposed (Figures 6.5; 6.6; and 6.7).

There are aspects of these "flame" structures that are not so typical of de-watering structures including 1) the diapirs are large (up to 15 cm) and widespread laterally throughout a thick (1.1 m) sequence of layered sediments and 2) the diapirs have occasionally detached from their source layer and risen above it, 3) in one exposure small scale (about 3 cm) high angle fault offset was

Figure 6.5 Cut-and-fill structures exposed by the Hudson Valley Sand and Stone Company excavations, about 4.5 km south of the field area.

Figure 6.6 De-watering ("flame") structures from the same locale as above.

Figure 6.7 Close-up of de-watering ("flame") structures.

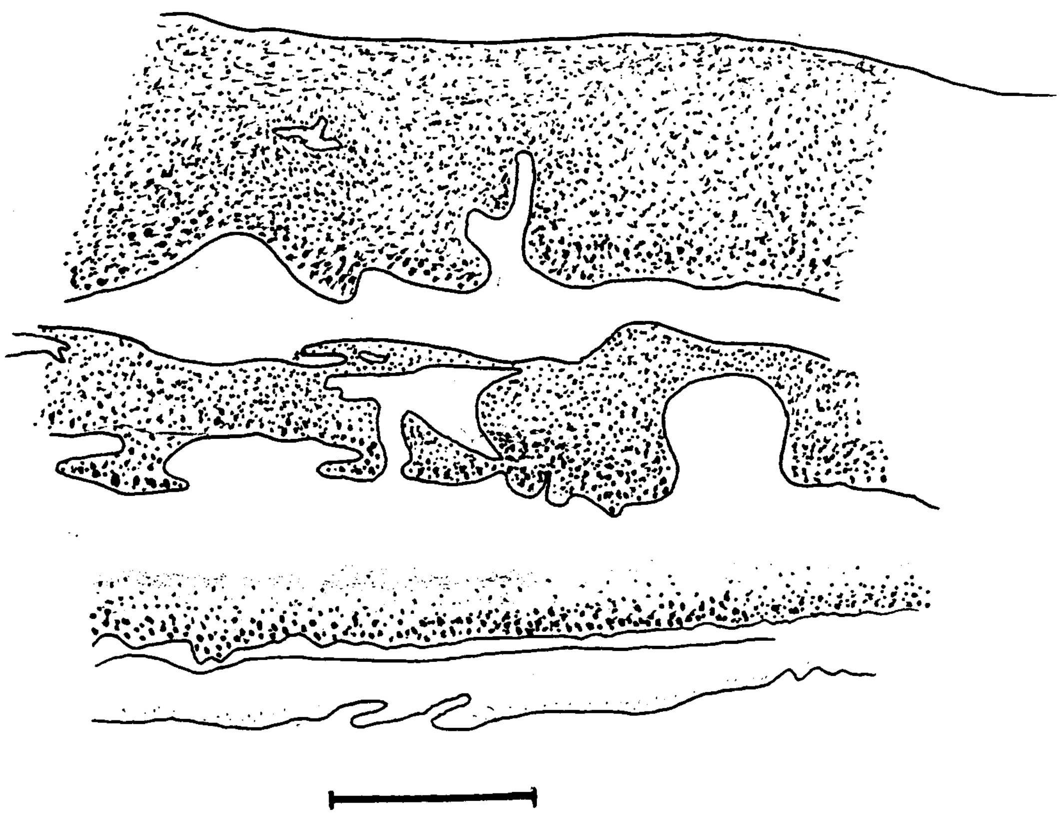


Figure 6.8 Sketch of sand diapirs in the Hudson Valley Sand and Stone Company excavation pit. Typically it is the contact boundary between the finest and most coarse layers that is "flamed". The coarse layers have not been mobilized as much and contacts between medium and fine sands show much less disruption. Scale bar = 15 cm.

observed. However, the conclusion that some vigorous convolution due to liquefaction has occurred to prompt this de-watering is not justified. Note that typically it is the contact between the coarse and very fine sediments that is flamed - only the finest fractions of sand have been "mobilized" upward. Contacts between medium and fine sands for example, can still be seen to be rather horizontal (Figure 6.8) and are not so commonly disrupted. This does not suggest that an earthquake induced "thixotropic" mixing event has occurred involving all the sediments. Also the beds below the deformed layers which in some places are composed of medium sand (gravel and sand in other places) over 1 m thick, show no sign of deformation. The small scale faulting which can be observed in "delta-toe" environments is not continuous down section.

While earthquake induced de-watering is a speculative mechanism, an alternative may be sediment loading above the "delta-toe" sequence due to pro-grading foreset beds; de-watering structures can be commonly observed in such environments. The local topography suggests that an upper limit to the thickness of this deposit was to an elevation between 360-400 feet, so 40 to perhaps as much as 80 feet of sediment has been removed. However, unconformities (including cut and fill structures) above the de-watered sequence have erased the continuous depositional record up-section (and in some cases have truncated diapirs) so no evidence exists that significant loading actually occurred at any time (directly) subsequent to the deposition of the de-watered sequence.

De-watering structures were also observed in fine sandy, silty bottomset beds exposed at an elevation of 315-325 feet west of Sherman Island near Big Bay Road and in an abandoned gravel pit near Fernwood (Figure 3.9). If the de-watering structures are laterally extensive and it could be established that they are within horizons which are stratigraphically equivalent (isochronous), a seismically induced de-watering event may be a more viable mechanism than a simple loading mechanism.

A conclusion of this investigation is that I would not recommend future paleoseismic trenching excavations on the McGregor Fault (system) unless a location can be found where 1) the width of the fault zone is observed to be narrower than usual - preferably within limits of the length of the trench itself, and 2) where the overburden is stratified and less than ten meters thick above bedrock. It is possible that these conditions could be satisfied along southern portions of the fault system, including the Saratoga and Ballston Lake Faults (Figure 1.3). Thin overburden would allow the fault to be located using seismic refraction techniques with a simple hammer and steel plate sound source. I would recommend that the geomorphology of these areas could be examined for evidence of rejuvenated scarp development and/or erosion features, such as fault stranded stream deposits (Figure 4.4). Some stream banks may have exposed enough overburden to allow a profile view.

It is unlikely that further examination of the delta and lake sediments of the western Glens Falls area will produce evidence of earthquake induced liquefaction effects because appropriate silt horizons or interbeds are rarely present.

An indirectly related course of future inquiry I suggest is the use of the Global Positioning Satellite system for determining more accurate rates of relative movement between different locations in and near the Adirondacks. Several temporary stations should be monitored monthly for at least a year, and may provide more specific information with regard to locations of maximum uplift, implicating areas most likely to accomodate faulting.

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