Issues the Core Team Needs to Address

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These comments provide an outline of four issues that the Core Team needs to address, if they have not already done so, as a prerequisite for any defensible proposal of a preferred alternative for decommissioning of the West Valley Demonstration Project and Western New York Nuclear Service Center. These four issues are not intended to be a comprehensive list of items needed for a defensible proposal of a preferred alternative, but they are crucial points, especially given the West Valley site's well-known sensitivity to water-related impacts such as erosional impacts from geomorphic downcutting, mass wasting, etc.

These four issues, if not already addressed, should have been on the Core Team's "radar screen" from the outset, based on the professional expertise of its members. The issues are not obscure. The West Valley Citizen Task Force has been attempting for several months to bring these issues to the Core Team's attention through a more convenient and interactive process (presentation to the Core Team at one of its meetings) but has not been allowed to do so; hence this written format. My own schedule during the past several months has been too busy to allow me to prepare these written comments until now. I regret the delay but reiterate that these are issues of which the Core Team should already be cognizant based on its members' scientific expertise.

The four issues, discussed below in more detail, are:

1. *Faults and seismic risk do not appear to be sufficiently characterized at the West Valley site to provide a reasonable understanding of long-term slope stability*. The seismic shock of earthquakes is a well-known cause or catalyst for slope failures. Seismically induced slope failures are a particular concern for this site because they greatly accelerate the erosional processes that will eventually undercut, expose, and release low-level, transuranic, Greater-Than-Class-C, and high-level wastes buried at the site. Seismic events can trigger near-instantaneous downslope movement of earthen materials, and the associated backcutting of ravine edges, that would take centuries under more ordinary erosional conditions.

2. *Current proposals to calibrate erosion modeling against the past several millennia of postglacial erosion cannot succeed unless postglacial uplift and the sequence of postglacial base levels are taken into account.* Lake Erie and its drainage basin have undergone very substantial differential uplift (glacial rebound) during the past several millennia, as is well known. The importance of uplift is well known to erosion modelers. The postglacial drainage history of the Cattaraugus Creek watershed has been discussed in the scientific literature and is at least approximately understood, and the associated base-level history of the Buttermilk-Franks tributaries can thus be inferred reasonably well. These complex histories are crucial parts of, and must be incorporated into, any erosion-model calibration. The importance of base level is well known to erosion modelers.

3. *Current proposals to calibrate erosion modeling against the past several millennia of postglacial erosion, and to conduct erosion modeling for several millennia into the future, cannot produce reasonable results unless both paleoclimate and future climate are estimated reasonably accurately.* As is well known, climate has not been uniform over the past several millennia; model calibration needs to reflect what is known. The best available predictions of future climate show a higher frequency of extreme events (including more severe droughts, more severe storms, etc.), as is increasingly evident in present-day weather events. These trends also need to be taken into account in a reasonable manner. The importance of climate is well known to erosion modelers.

4. *Performance assessments for various critical systems at the West Valley site cannot be defensibly based on deterministic risk assessment; they need to be based on probabilistic risk assessment*. Many of the decommissioning decisions for restricted release at the West Valley site require complex chains of logic, involving poorly constrained input parameters, to demonstrate that site integrity can be maintained for millennia in the absence of institutional controls. There is no reasonable way to apply purely deterministic risk assessment to such decisionmaking.

Faults and seismic risk in relation to slope failures

A seismic survey by Bay Geophysical (2001), conducted on behalf of West Valley Nuclear Services Company in response to comments by Vaughan (1996) and others, has identified two deep-seated faults near the West Valley site. These previously unacknowledged faults, termed the "Sardinia feature" and the "Cattaraugus Creek feature," are interpreted as being present in both the Precambrian basement and overlying Paleozoic bedrock. Both "appear to exhibit displacement extending from the Precambrian basement into the middle and potentially the upper Paleozoic section of the stratigraphic section." However, due to the limited scope of the work conducted by Bay Geophysical (2001), many questions remain unanswered about these faults. Neither the strike nor the extent is known for either fault (in other words, there is currently no reliable information on the geographic direction or the geographic extent of either one), so there is currently no way to know whether either fault extends close to, or directly beneath, the West Valley site. Other unanswered questions are whether future seismic activity or "reactivation" may occur on either fault, and, if so, what is the likely relationship between earthquake magnitude and recurrence interval. All of these questions are susceptible to investigation, and reliable answers are needed to understand whether future seismic activity poses a serious risk to site integrity.

The fault known as the "Cattaraugus Creek feature" has been mapped (Bay Geophysical 2001, Fig. 4-1) immediately north of the US 219 bridge over Cattaraugus Creek. Coincidentally or otherwise, this is the same location where severe landsliding, said to be reactivation of an ancient landslide. has recently occurred (Bonfatti 2007). Three faults of small displacement, visible in the gorge of Cattaraugus Creek near this location (Vaughan et al. 1993), may potentially be surface expressions of the "Cattaraugus Creek feature." The mapped location of the fault is roughly 3 miles or 5 km from the West Valley site; however, as noted above, it is not yet known whether the fault extends closer to the site. See Vaughan (2005) for further discussion of possible fault relationships in the vicinity of this fault.

The fault known as the “Sardinia feature” may potentially be the southwestward extension of a known fault, the Attica Splay of the Clarendon-Linden Fault. If so, the potential for reactivation of the “Sardinia feature” would appear to be high, based on the well-known 1929 Attica earthquake and other, smaller-magnitude seismic activity on the Attica Splay. However, the current lack of information on the strike and geographic extent of the “Sardinia feature” does not allow reliable conclusions to be drawn about either its connection with the Attica Splay or its nearest approach to the West Valley site. Its mapped location near Sardinia, NY (Bay Geophysical 2001, Fig. 4-1) is roughly 10 miles or 16 km from the West Valley site; however, if this fault is indeed a southwest continuation of the Attica Splay, its southwestward projection from Sardinia implies an approach much closer than 10 miles to the West Valley site. See Vaughan (2005) for further discussion of possible fault relationships in the vicinity of this fault.

Effects on slope stability are a primary reason why seismic activity needs to be well understood at the West Valley site. The site is highly susceptible to both erosion and slope failures (for example, see DOE and NYSERDA 1996, including Figures 4-13 and L-1; Albanese et al. 1984; Boothroyd et al. 1979; Boothroyd et al. 1982); however, seismically-induced slope failures have not been adequately addressed (see Vaughan 1996) and cannot be reliably addressed without a better understanding of faults, seismicity, and recurrence intervals in the immediate vicinity of the West Valley site.

The connection between slope failures (i.e., landslides of various types) and earthquakes is well-known (Sidle et al. 1985; Keller 1985). The 1964 Alaska earthquake produced slope failures on a massive scale (see photos in Hansen 1971, attached hereto as Appendix C), but smaller seismic events may also cause failures on susceptible slopes. Sidle et al. (1985) note that, “For most landslides that were initiated by earthquakes, the direct physical and mechanical influence of the ground motions appeared sufficient to generate failures on slopes that were in a delicate state of balance.” The “delicate state of balance” criterion seems to be met at the West Valley site, given the many slope failures that are presently occurring at the site without additional seismic enhancement (DOE and NYSERDA 1996, including Figures 4-13 and L-1; Albanese et al. 1984; Boothroyd et al. 1979; Boothroyd et al. 1982). The main impact of seismically induced slope failures at the West Valley site under loss-of-institutional-control scenarios would be an intermittent acceleration of the backcutting of ravine edges. Such accelerated backcutting of ravine edges would expose buried wastes more rapidly than under normal erosional conditions.

Slope failures elsewhere in western New York State, including locations close to the West Valley site, may be instructive. See discussion in Vaughan (1994), including information on unstable lacustrine sediments (similar to quick clays) observed north of Springville, NY, by Owens et al. See also Gephart-Ripstein (1990) for a review, based on anecdotal and historical sources, of earthquakes and slope failures in Wyoming County, NY. Some of the earthquakes cited there are apparently not listed in standard modern earthquake catalogs. The slope failures reviewed by Gephart-Ripstein (1990), typically involving 6 to 20 acres, are not explicitly linked to earthquakes yet are of interest because some are located along the Tonawanda Creek valley between Attica and Varysburg, NY. This creek valley, apparently structurally controlled, follows the Attica Splay of the Clarendon-Linden Fault (see Fakundiny et al. 1978; Fakundiny and Pomeroy 2002). Despite the lack of explicit linkage to seismicity, the slope failures may be fault-related if the fault serves as a conduit for fluid flow.

Postglacial uplift and base levels

The Lake Erie drainage basin has experienced a very high rate of differential uplift (glacial rebound) during the postglacial period. As can be seen from Newman et al. (1981), the Great Lakes are within an area that experienced one of the highest uplift rates anywhere in the world during the past several thousand years. Holcombe et al. (2003) provide a more detailed view of the uplift, particularly the differential uplift, for Lake Erie. The net effect, recognized for more than a century (see Tarr 1897), is that the eastern end of Lake Erie has risen tens of meters relative to the western end during the past several thousand years. Holcombe et al. (2003), p. 693 and Fig. 8b, infer a differential uplift of 45 m over a horizontal distance of 100 km during the past 13,400 years. This uplift rate applies to the eastern basin of Lake Erie, which is the portion of the lake into which Cattaraugus Creek drains. Tarr (1897), p. 113, infers a somewhat lower uplift of about 1 foot per mile at the eastern end of the lake. These rates, while they cannot be applied verbatim to the generally parallel and immediately adjacent drainage basin of Cattaraugus Creek, serve as a reminder that the differential rates of uplift for Cattaraugus, Buttermilk, and Franks Creeks are likely to be high and need to be established with reasonable accuracy for any erosion modeling runs conducted for the postglacial period.

Not only the intra-watershed differential uplift but also the regional uplift and associated base levels need to be established for modeling runs. Tucker and Bras (1998), for example, use a parameter U to represent the uplift or base lowering rate; this parameter, which they equate with the steady-state erosion rate, is incorporated into key equations of their model.

Even without the added complication of differential uplift, the sequence of postglacial base levels for the drainage areas that we now identify as Cattaraugus, Buttermilk, and Franks Creeks is complex. During and after glacial retreat, drainage along today's west-flowing Cattaraugus Creek was blocked for some period of time. Flow was initially blocked by the ice dam of the glacier face and subsequently by a sequence of rock and/or ice dams that persisted until downcutting of today's Zoar Valley gorge (between Zoar Bridge and Gowanda, NY) was achieved. Impounded water, called "Lake Cattaraugus" for purposes of this discussion, persisted for some period of time and served as the sporadically decreasing base level for the drainage areas that we now identify as Buttermilk and Franks Creeks. This "Lake Cattaraugus" base level fell as new outlet channels for the lake became available, following the sequence described by Fairchild (1932). Briefly, this sequence starts with glacially impounded meltwater in the eastern part of today's Cattaraugus Creek basin; the surface of this initial glacial lake was about 1640' above modern sea level, corresponding to the elevation of the lowest terrain (near Machias, NY) over which water could flow out of the glacial impoundment. As the glacier retreated west and north, the level of "Lake Cattaraugus" fell progressively as the receding ice uncovered lower outlet channels at Ellicottville (1620'), Little Valley (1610'), New Albion (1440'), Persia (1320'), and finally Perrysburg (1300' and lower). However, when the level of "Lake Cattaraugus" dropped to roughly 1200', it was rock-dammed by the "Zoar Valley" bedrock into which the spectacular modern gorge had not yet been cut. Downcutting through the shale bedrock, no doubt aided by existing joints and small splays of Bass Island Trend faulting, was required before Cattaraugus Creek could flow westward through its modern channel. In the interim, "Lake Cattaraugus" remained the base level for the drainage areas that we now identify as Buttermilk and Franks Creeks.

Any landscape evolution modeling of the postglacial period needs to make reasonable assumptions about the duration of each "Lake Cattaraugus" base level *and also needs to make a differential-uplift adjustment or correction to each of the base-level elevations cited above*. Fairchild (1932) assigns the modern elevation above sea level to each of his outlet channels (e.g., 1300' at Persia, NY), but an early post-glacial "Lake Cattaraugus" surface level that matches a modern 1300' elevation at Persia is not likely to match a modern 1300' elevation in the drainage areas that we now identify as Buttermilk and Franks Creeks. Differential uplift requires some amount of correction, including both an initial adjustment and an ongoing, time-varying adjustment. Such adjustments are not necessarily large but may be important for a landscape evolution model whose initial condition "is a nearly flat surface seeded with a small random perturbation in the elevation of each cell" and where the boundary condition "is a single fixed outlet in one corner" whose base level is apparently lowered at a rate U (Tucker and Bras 1998). The "nearly flat surface" assumed as an initial condition is not likely to remain flat and horizontal as differential uplift proceeds. (Phenomena such as stream capture or flow reversal may occur as the surface tilts.) The rate U needs to be tied to both the changing elevation of the "Lake Cattaraugus" outlet and the elevation correction needed to compensate for differential uplift between the lake outlet and model outlet.

Paleoclimate and future climate

The relationship between precipitation and erosion is very non-linear. Much greater erosion occurs when a given amount of rain falls during a short time (e.g., in an intense storm) than when the same amount of rain falls gently over an extended period. Part of the reason is the difference in velocity, and especially the difference in kinetic energy, of the water flowing though stream channels in the two different cases.

Because erosion is so dependent on the rate at which precipitation is delivered, erosion modelers need to 1) model the precipitation-erosion relationship accurately, using appropriate algorithms in their computer code, and 2) use realistic precipitation data, or realistic sequences of assumed precipitation, for modeling runs that simulate either *past* or *future* erosion.

For example, any modeling of past erosion at the West Valley site (e.g., for model calibration purposes) needs to use realistic sequences of assumed precipitation that are based on, and consistent with, available paleoclimate information. Likewise, any modeling of future erosion at the West Valley site needs to use realistic sequences of assumed precipitation that are based on, and consistent with, a good understanding of climate change.

Two potentially useful sources of paleoclimate information, not intended to be exhaustive, are Noren et al. (2002) and Holcombe et al. (2003). Based on sediments deposited in lakes in Vermont and eastern New York, Noren et al. (2002) identified four periods of intense storminess that occurred about 11,900, 9,100, 5,800, and 2,600 years ago. Interspersed between the second and third of these storm periods was the middle Holocene climatic optimum (9,000 to 6,000 years ago), during which "warmer temperatures and greater aridity" characterized the climate of the Lake Erie region, according to Holcombe et al. (2003). Such paleoclimate information provides guidance needed for modeling of past erosion at the West Valley site.

Any modeling also needs to ensure, in accordance with principles of mass balance, that a reasonable sink exists for water discharged from the model outlet. While this is not likely to impose a substantial constraint on modeling, and certainly would not be a constraint under today's drainage conditions where the Atlantic Ocean is the sink for water discharged from Cattaraugus, Buttermilk, and Franks Creeks, modelers should be aware that Holcombe et al. (2003) consider Lake Erie to have been a closed basin during part of the postglacial period. At times when the lake is considered a closed basin, the discharge flow rate of water from the outlet of a landscape evolution model should not be an disproportionate share of the flow that could reasonably be accepted by the closed lake basin.

Future climate inputs to erosion models must reflect the increasing frequency of extreme weather events, especially intense storms, that are a predicted consequence of climate change. Recent intense storms in New York and surrounding areas, regardless of whether they are early signs of this trend, provide perspective when compared to the most severe storm to affect the West Valley site in the past decade. This storm, whose 3.25" overnight rainfall produced noticeable erosion on the West Valley site and washed out part of Schwartz Road near the site, delivered a mere fraction of the rain that recent intense storms have brought to other locations such as Peterborough, Ontario (8" in July 2004), parts of PA-MD-VA (over 12" in June 2006), the Painesville, OH area (10-12" in July 2006), and north-central OH (9" in August 2007). Short-duration storms of this type are not uncommon and are likely to increase as climate change becomes more severe. (Part of the reason is simple: Warmer air can carry more moisture.) Thus, any modeling of future erosion at the West Valley site needs to incorporate these storm trends.

Probabilistic Risk Assessment

Probabilistic risk assessment (PRA), used by various industries and regulators, "allows analysts to quantify risk and identify what could have the most impact on safety." It "systematically looks at how the pieces of a complex system work together to ensure safety." (NRC 2007.) A complex system might consist of a space shuttle, all of whose components must function properly to ensure a productive mission and safe return to Earth, or it might consist of a nuclear waste disposal system, all of whose components must likewise work properly to protect public health and the environment. Members of the Core Team should recognize the applicability of PRA to complex waste disposal proposals (such as in-ground closure of grouted high-level waste tanks at the West Valley site) and should insist that PRA be used in preference to a "seat of the pants" approach.

PRA is a good way to analyze complex results, especially where there is uncertainty in the results and in the values that must be assumed to calculate results. PRA results "do not take the form of a single number. Instead, PRA provides a spectrum of possible outcomes. The frequency with which each of these outcomes is expected is a *distribution* of values." PRA results can often be summarized by a single representative value or *point estimate*, but PRA's main advantage is that it helps decisionmakers understand "how much larger or smaller the actual risks might be." (NRC 2007.)

The Nuclear Regulatory Commission has used PRA for many years for nuclear power plant analyses. According to NRC (2007), "PRA use is expected to continue growing as part of a longstanding NRC policy for increased use in all regulatory matters. This should result in a more predictable and timely regulatory approach throughout the agency." Even though NRC has not yet made an effort to apply PRA to the West Valley site, it should do so. PRA methods are needed at the site regardless of whether NRC sees its role as "regulatory." Other Core Team members need to ensure that PRA methods are adopted at the site *or need to be willing to discuss, in an accessible forum, why they are reluctant to do so.* Complex outcomes such as in-place closure scenarios at the West Valley site are ideal candidates for PRA; they would benefit from its structure and logic.

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Appendix A

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NEW YORK PHYSIOGRAPHY AND GLACIOLOGY WEST OF THE GENESEE VALLEY

BY

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CONTENTS

General Statement

This district, the extreme western portion of the State, includes a variety of singular and highly interesting features, physiographic and geologic, many of which have their first notice in this writing. In the paper number 5 of the attached list of writings Dr. Frank Leverett, in 1892, described the morainal belts; and a quarter of a century ago the writer mapped the inscriptions of the widespread glacial waters in a set of colored charts in paper 10. The features to be specially noted in the present writing include the remarkable Cattaraugus Valley and the series of parallel valleys on the north, with their romantic glacial lakes.

The unusual assemblege of striking features in this district, with their origin and glacial history, deserved detailed description long ago by the local scientific interests. In the lack of such publication the Rochester Academy of Science presents this outline of the physical characters of the neighboring province.

When the paper number 10 was published the topographic mapping by the United States Geological Survey had not covered all of the area and a number of the topographic sheets were lacking. Plate 4 in that Bulletin 106, covering a stretch from Gowanda north to Hamburg and Orchard Park, was based on imperfect county maps, and the glacial lake shores and the stream channels could not be precisely shown in either location or form, and the topography not at all. That district is now shown, in reduced scale, in the accompanying plate 29. However that former paper, State Museum Bulletin 106, is yet requisite for the general glacial history as introduction to the present paper. Unfortunately it has long been "out of print," but copies should be found in the libraries of the larger High Schools and certainly in the city libraries.

The geographic divisions of the district as based on the river systems or hydrographic areas are shown in plate 27. The present disposition of the drainage dates from the close of the Glacial Period. Previous to the "Ice Age" and the invasion of the Quebec glacier all of the major streams flowed northward, as indicated in figure 2.

The continental ice sheet in its relentless overriding of the land piled heavy fillings of rock rubbish, the glacial "drift," in many valleys, thereby interfering with the ancient stream flow. The present divide between the Allegheny River and Lake Erie was created by the drift blockade, and also the divides north and south of the

Cattaraugus Valley. This anomalous valley, which is out of harmony with the topography and stream flow of the large district, was developed at the front of the waning ice sheet by glacial and drift damming and the forced glacial drainage. This remarkable valley is entirely due to glacial interference, and will be briefly described, with the aid of the map, plate 28.

Plate 29, emphasizes the series of eight parallel north-leading valleys, with the forced, or ice-border, drainage during the recession of the ice front. The map, with its great reduction in scale, is mainly suggestive. The glacial stream channels are much exaggerated in width. Students, in and out of the schools, may find satisfaction and instruction in correcting possible errors and in the discovery of unmapped features and interesting details.

This writing does not attempt a complete, detailed description of the district, with its wealth of very instructive physiographic and glacial characters. Such treatment would require, and deserves, a large monograph. Its aim is to direct attention to the salient characters of the region as records of dramatic geologic activities; and also to encourage people with interest in nature to explore their home surroundings. The topographic maps should be utilized by the schools for recognition and study of their local features. Exploration in the field with intelligent study of the geologic phenomena is the highest intellectual exercise, and may well replace much study of books.

Map Illustrations

Plate 27 distinguishes the several drainage districts or stream system. Three of these are entire, the Cattaraugus, Erie-Niagara and the Ontario west of the Genesee Valley. For New York the Genesee Valley is also complete. The Allegheny headwaters is widely shown.

This map should be compared with figure 2. It will be seen that the Genesee River is the only large stream which has retained its general northward course across the State, in spite of the interference and opposition of the continental glacier.

Plates 28 and 29 have utilized as base maps the topographic sheets by the United State Geological Survey. Great reduction in size has been necessary. The sheets are easily available for study.

Plate 28 shows the central portion of the singular Cattaraugus Valley, using the Cattaraugus, Ellicottville and Franklinville sheets.

The three rock ravines may be located. On the southern divide the low passes which were the outlets of the high-upheld glacial waters are indicated, with some exaggeration. This map adjoins, with the same scale, the bottom of the map in plate 29. The two maps cover all of the Cattaraugus drainage except the few miles west of Versailles Village.

Plate 29 emphasizes the series of north-leading parallel valleys. Comparison of drainage should be made with figure 2. The many passes which provided outlets for the ice-impounded waters are interesting features and conspicuous in the mapping, but are not complete for the northern part of the area.

The moraines and the outwash plains, with their kettles and kettle lakes, are not specially designated. They may be recognized by the topography, and are noted in later chapters.

The shore lines and deltas of the widespread glacial lakes on the lower ground are heavily indicated. The Whittlesey features are incomplete, being mapped only where seen in the field. Some enterprising student should map them with refinement.

For areas north and south of plate 29 the lake features are mapped in paper 10.

The figures, 1, 2, 3, are described in the text.

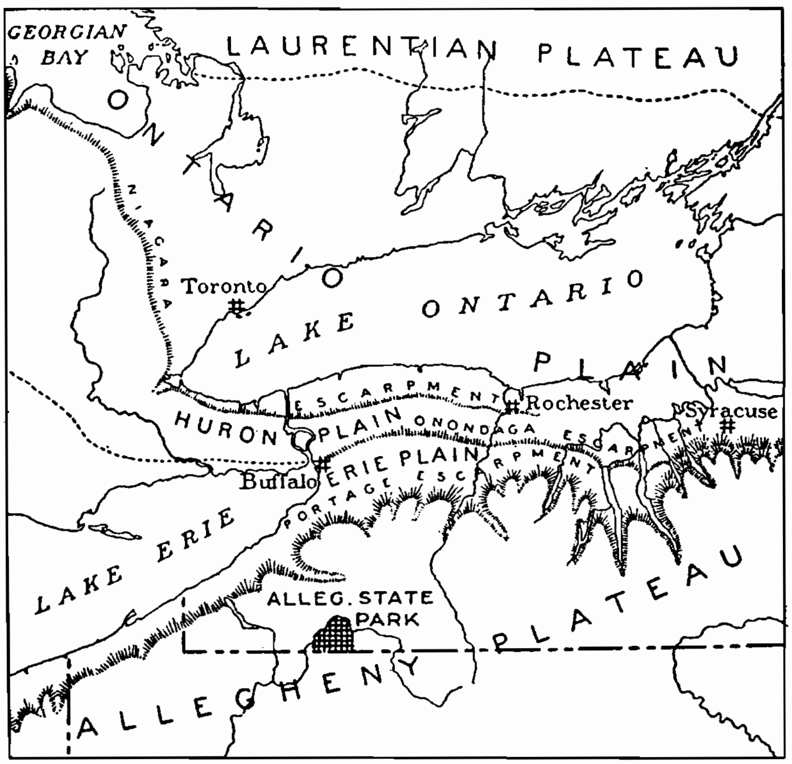
Physiography

One of the physical characters of which western New York may properly boast is its singular physiography. Its land surface is recognized as the type example of "Cuesta" topography, a titanic stairway with insloping treads.

The production of this singular structure required an unusual combination of geologic elements, and is the record of especially interesting history. When the area of western New York and Ontario was permanently raised out of oceanic waters, the interior or epicontinental sea, the newly created land, a vast coastal plain, had a general slope toward the south and southwest. But to-day the land surface rises southward, from the level of Lake Ontario, 246 feet above tide, and Lake Erie 573 feet, to much over 2,000 feet in the southern tier of counties.

During unnumbered millions of years exposure to the destructive agencies of the atmosphere, along with high uplifting and deep erosion of the rock strata, the land has been carved into a series of plains with sharp upslopes or escarpments. The three plains, Ontario Lowland, Huron and Erie, with three escarpments, Niagara, Onondaga and Portage, rise to the elevated Allegheny Plateau. These interesting physiographic elements are clearly shown in figure 1, reproduced from Professor A. K. Lobeck's figure 9 in his Handbook for the Allegany State Park, paper 21 of the appended list of writings.

The succession, rising southward, of plain and scarp, is an expression of three conditions; (1), the varying resistance of the different



Courtesy of the New York State Museum

Figure 1. WESTERN NEW YORK PHYSIOGRAPHY

rock strata to the atmospheric erosion; (2), the high-uplifted attitude permitting deep erosion; and (3), the northward upslant, or southward "dip" of the strata of perhaps fifty feet to the mile.

The extended succession of rocks, in all variety of the sedimentary class, may not be described here, being noted in all text-books and treatises on the New York stratigraphy, and given in detail in paper

16. The succession of less resistant rocks constitute the plains, the "treads" in the physiographic stairway; while the more resistant beds make the scarps or "risers." The wide outcrop of thick and weak Ordovician strata, the beds beneath the Medina sandstone, were eroded in Tertiary time to produce the deep and wide Ontario basin, with its eastern extension as the Mohawk Valley. This great depression reversed the direction of river flow, described below.

In the western part of the State the Ontario Lowland is faced by the Niagara escarpment, which is capped by the Niagara or Lockport limestone. The outcrop of the thick and weak Salina shales, containing the rock salt of New York and Michigan, has been excavated to produce the east and west depression through western and central New York, named in figure 1 the Huron plain. The Onondaga limestone forms the scarp at the south border. This Salina depression had gathered, by the end of Tertiary time, a large part of the regional drainage into east and west courses, as shown in figure 2; and that control of drainage is yet exercised in spite of the glacial interference.

Overlying the Onondaga limestone the thick shales and sandstones of the middle Devonian strata constitute the Erie plain of figure 1. The heavy strata of the Portage and the Chemung build the high scarp that forms the north-facing front of the Allegheny Plateau.

The minor physiography, especially the preglacial and the present river valleys, will be described below.

Tertiary, Preglacial, Stream Flow

The ancient river flow, as noted above and as mapped in figure 2, was northward or northwestward and tributary to the Mississippi through the master rivers in the Ontario and Erie basins. However, this northward drainage was not the earliest stream flow. Far back, in the Devonian Period of Paleozoic time, the earliest or primitive streams were created when the area was permanently raised out of the interior or continental sea, and their direction of flow was southward or southwestward. With further uplift of the continent the Ontario-Mohawk and Erie valleys were carved out of weak strata. And these east and west depressions captured the drainage and reversed the direction of flow. With the northeastern part of the continent standing perhaps several thousand feet higher than at present the north-flowing streams were vigorous and rapidly cut deep, steep-sided valleys. This drainage history is described in papers 14–16.

As noted in the preceding chapter the present land surface with its striking topography is the final effect of the removal of great thickness of rock strata, producing the deep valleys and the basins

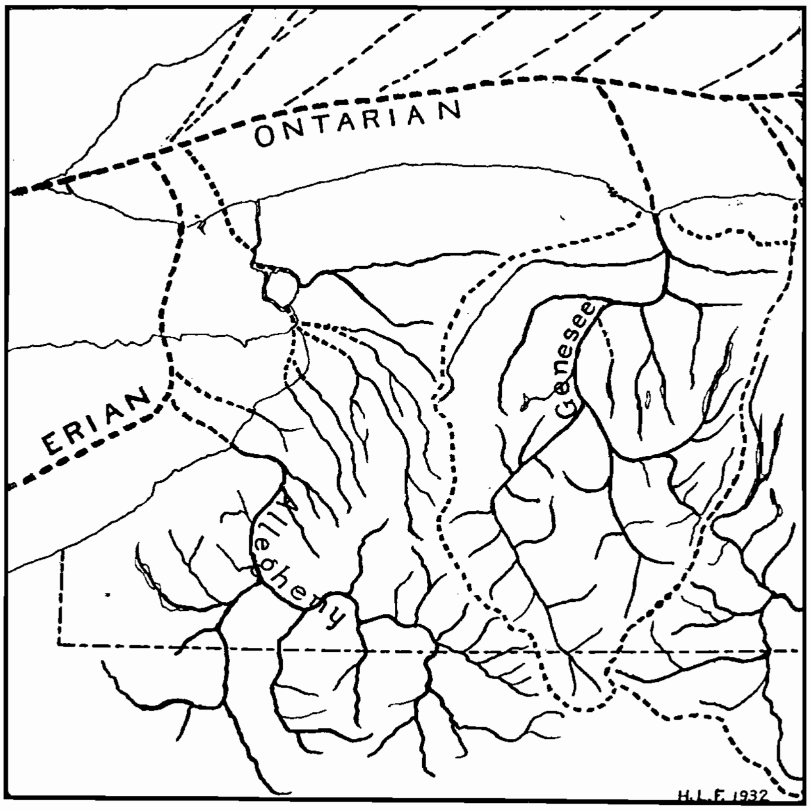


Figure 2. PRECLACIAL DRAINAGE IN WESTERN NEW YORK

now holding lakes Erie and Ontario. The strata which constitute the highland, the Allegheny Plateau, once extended far north into Canada. The detritus derived through millions of years of destructive processes has been largely contributed to the Mississippi River deposits in lower Louisiana.

In the lowering by erosion of the land surface, with the shifting of river flow, the paths or valleys of the primitive, south-flowing rivers are obliterated or quite obscured in all the area under description, and any doubtful relics are not here considered. North of the Cattaraugus Valley the many valleys declining northward (plate 29) are relics or survivors of the preglacial, Tertiary drainage. And an interesting fact which has not been recognized is that they are the western members of the remarkable series of parallel valleys lying eastward through central New York.

South of the Cattaraugus basin, where all of the present stream flow is southward, to the Allegheny River, the preglacial, north-leading valleys are deeply filled with detritus, from glacier, lake and river, and are now holding wide plains and swamps. These wide valleys with deep filling and slack drainage are abnormal features, being an indirect effect of the invading and overriding ice sheet. The deep, north-declining valleys were blocked in places by glacial drift fillings. South of the drift dams the valleys have been filled by detrital outwash from the ice sheet, and subsequently, in some cases, by glacial lake deposits and finally to the present level by the stream detritus of the reversed, or south-leading drainage. The conspicuous example is the wide, capacious valley now slowly drained by the Conewango Creek. Along with the lower stretch of the Cattaraugus the Conewango valley was the path of the preglacial Allegheny River.

Deep drillings in the line of the larger preglacial valleys go below sea level before reaching "bed rock," which is one proof of the high elevation of the region before glacial time.

The rock bottoms of these filled valleys were graded to the rock bottom of the Erie Valley. Drillings, for gas or water, down to bed rock will prove that the valley bottoms have a uniform slope northward. The amount of that slope has been slightly reduced, one or two feet per mile, depending on direction, by the northward uptilting of the land since the weight of the ice sheet was removed (papers 13, 14).

The diverting effect, direct and indirect, of the ice sheet invasion may be seen by comparison of the Tertiary drainage shown in figure 2 with the map of the present stream flow, plate 27.

The map of Tertiary drainage, figure 2, indicates by broken lines the stream flow which is chiefly hypothetic. Those north of the axis of Lake Ontario retain the primitive southerly direction of Paleozoic time. The drainage relation of the Erie basin to the

Ontario basin in preglacial time, with the ultimate outflow, is a difficult problem. In this map the Erian and Niagaran flow is tributary to the Ontarian. It must be understood that today the bottom of the shallow Lake Erie is over 600 feet above the bottom of Lake Ontario.

Glaciation

The high elevation of northeastern America in later Tertiary time was one cause, if not the principal cause, of the formation of an ice cap, or continental glacier, in Canada which expanded so as to cover most of New York. Students in the western states and in Europe find that the Glacial Period covered epochs of glaciation with intervening mild climate epochs of deglaciation. And the duration of the Glacial Period has been greatly extended, to cover 500,000 years, or more.

As applied to western New York our knowledge is limited to the closing stage of the Ice Age, the Wisconsin Epoch, and the operation and effects of the Quebec (Labradorian) ice sheet. The glacial phenomena are so well known that general description is here unnecessary, but some particular phases of the glacial behavior and interesting records will be considered in the following chapters.

The direction of movement of the overriding ice sheet, at least in the later phase, was southwestward from the Ontario basin, as shown by the orientation of the drumlins in Orleans and Genesee counties. The spreading flow from the Erie basin was to the southeast, as shown by the attitude of the drumlins in Chautauqua County and the trend of the terminal drift or moraines.

Ice-Front Recession

The geologic effects by the invasion (or invasions) of the ice sheet in western New York are a matter of conjecture or inference. The recessional records are conspicuous as the will and testament of the recently departed visitor.

The moraines in figure 3 mark the succession of ice-front positions as the glacier was reluctantly departing. South of Lake Ontario the trend of the ice margin was quite parallel to the lake shore. In the Erie basin the spreading movement of the ice body caused the ice front to lie slantingly or obliquely on the New York wall of the valley. And in consequence of this oblique relation the ice was not removed at once along the whole length of the wall or scarp, from State Line to Buffalo, but receded progressively from the southwest to northeast.

The glacial waters which produced the beaches mapped in plates 2–5, paper 10, and in plate 29, crept in from the west between the ice and the land slope, slowly extending northeastward. For example, the site of Dunkirk was yet beneath the ice sheet while Westfield was submerged in Lake Whittlesey.

The Lake Warren beaches were not formed until after the ice front had receded to the district of Marilla, fifteen miles east of Buffalo. At that time the far-west ice margin in the State of Michigan released an outlet lower than that of Whittlesey, and the imprisoned waters fell, some 45 to 50 feet, to the Warren level. The history is given in paper 10, pages 41–74, and in paper 17.

The moraines and outwash plains along the north side of the Cattaraugus Valley were under construction about the time, perhaps, when the Whittlesey waters were entering the State.

The glacier front in its recession over the areas mapped in plates 28, 29, probably had a general direction northeast by southwest. In its backward movement the drainage outlets for the glacial waters were probably opened sucessively westward.

Moraines and Valley Fillings

The slow melting of the ice sheet, with the backward recession (to northwestward) of the margin of the ice, left, normally, a thin mantle of ice-laid material, the "till," on the deserted land. But the waning of the glacier was far from steady or uniform. Climatic and other factors made the ice-front recession somewhat spasmodic. Along some stretches the ice margin hesitated and lingered, producing masses of the drift, as moraines. Occasionally the ice front readvanced after it had receded less or greater distance. Such pauses at readvanced positions produced the heavier moraines on the uplands and the massive blockades in the valleys.

These drift deposits are responsible for the changes in the drainage and for some of the anomalous features noted above with others to be described.

Figure 3 gives the location of the more extended and massive accumulations of frontal or morainal drift. The matter is discussed in paper 18.

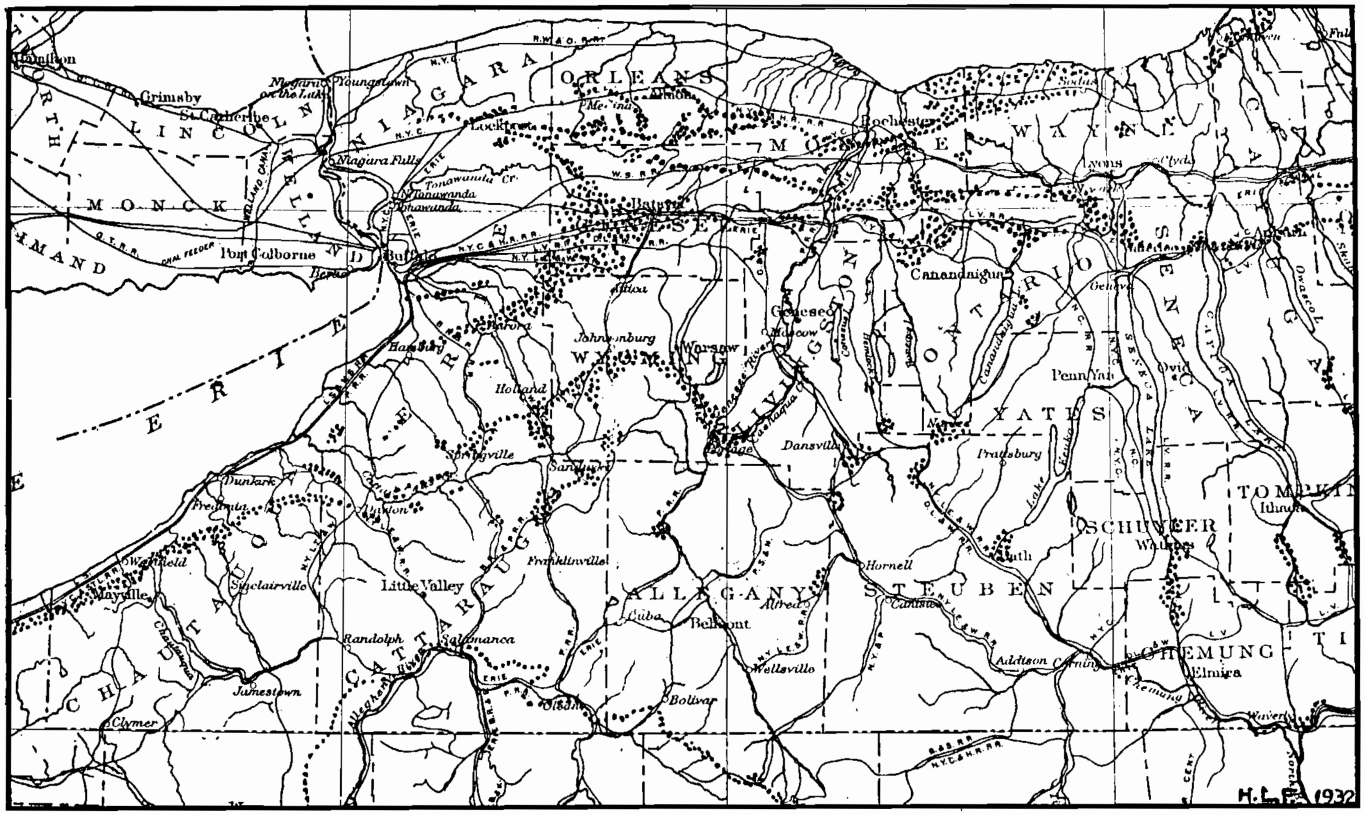


Figure 3. MORAINES IN WESTERN NEW YORK

The most continuous and extended moraine in the area is the one which forms the crest of the steep slope, or scarp, facing Lake Erie. This was described and mapped in paper 5, and named by Leverett the Lake Escarpment moraine. It appears on the Westfield and Dunkirk sheets of the topographic map, in plates 1–3 of paper 10, and in figure 1 of paper 18.

Distinct and definite moraine ridges, marking precise positions of an ice front, are rare and seldom will be found above the Terminal Moraine or on territory with high relief.

On plate 28 the irregular topography east of Dayton indicates a morainal tract. Also along the Cattaraugus Creek north and south of Zoar Bridge. A conspicuous moraine field lies south of Delevan and north of Machias; and another one at Sandusky, on Clear Creek in town of Freedom (Arcade and Franklinville sheets).

On plate 29 a series of heavy moraines and valley fillings are seen at the heads of all the north-leading valleys except that of Hunter Creek. They lie on the Eden, Springville and Arcade sheets. These fillings helped to produce the northern divide of the Cattaraugus basin, and they apparently correlate with the Escarpment moraine and its eastward continuation as the Valley Heads moraine in central New York. This moraine is mapped in papers 5, 10, 18.

The long standstill of the ice margin in the production of these fillings and the blockade of the ancient valleys forced the westward flow of the proglacial, or ice-border, drainage which developed the Cattaraugus Creek.

Outwash and Valley Plains

Interesting features connected with the valley moraines are the plains of "glacial outwash." These water-smoothed plains lie in front of the moraine fillings and were built of the detritus, gravel, sand and silt washed out of the glacier by the subglacial streams that drained the melting ice sheet. The detritus filled the incipient lakelets which were held in the heads of the valleys by the ice dam, the fillings being smoothed by the lake waters, and later channeled by the streams of the outflow.

These plains are recognized on the topographic sheets by their many basins, the "kettles," and the kettle lakes and lakelets, which features are characteristic of frontal moraine deposits.

The outwash plains in this area have no superiors, and lying on the north border of the high Allegheny Plateau they are remarkable for their altitude.

On plate 28 a good example is seen at Machias, which includes Lime Lake. Another is seen at and south of Freedom village, which includes Crystal Lake, with elevation 1,761 feet.

On plate 29 are the larger outwash plains. Extensive plains appear at East Concord and Springville, 1,400 feet elevation; at Chaffee-Yorkshire–Sardinia, 1,400 to 1,450 feet. The most elevated is at Eagle village (Arcade sheet), four miles southwest of Bliss, at the head of Wiscoy Creek in the Genesee drainage. Its altitude is 1,900 feet, which probably distinguishes it as the most elevated outwash plain in America if not in the entire world.

Valley Plains lead away, southward, from the heavy moraines on the divides, the detritus filling the valleys being derived at first from the glacier and supplied later by the lakes and the land drainage.

Good examples of these features are seen in the two valley plains which head in the moraine south of Dunkirk. One is the Bear Lake Valley, which holds the village of Stockton. The other is the upper part of the Cassadaga Valley, headed by the Cassadaga lakes. These two valleys unite to make the greater Cassadaga Valley, with a breadth at the junction of over three miles (Dunkirk sheet). The full display of the handsome valley is mapped on the Dunkirk, Chautauqua and Jamestown sheets.

Passing eastward we find definite outwash plains, connected with moraines, at Mud Lake, the head of the West Branch Conewango Creek; at East Mud Lake, the head of the North Branch Conewango; and at the village of Cottage on Slab City Creek. The plains of glacial outwash are shown on the Cherry Creek sheet.

Another excellent example, and with large kettle lakes similar to the Cassadaga lakes, is at Machias, at the head of the Ischua Valley, as shown on the Franklinvile sheet, and on plate 28.

The valley plains which are headed by the outwash deposits noted above are very conspicuous geographic features on the topographic sheets, and require particular notice.

The elevation of Bear Lake is a litle over 1,300 feet; and the upper Cassadaga Lake is 1,306 feet. In the distance of twenty miles to the junction of the Cassadaga Valley with the Conewango Valley the decline in elevation is only forty feet. This remarkable flatness is because the ancient valley declined northward and has been deeply filled by glacial, lake and stream detritus. Such peculiar valley forms are produced by morainal blockade and deep filling in valleys of reversed stream flow. These valley features are mapped in plate 2, paper 10, and also in paper 9, plate 38.

Twelve miles east of the Cassadaga Valley lies the very broad valley of the Conewango Creek. The creek heads in six branches, and their very interesting directions and relations are clearly shown on the Cherry Creek and Cattaraugus sheets. They all unite in the district of South Dayton and Markham villages. The recognized source of the creek is far east, by the hamlet of New Albion, three miles south of Cattaraugus village.

The Conewango Valley is remarkable in its size, form and history. It begins in a great basin at Markham, Dayton and Persia, four miles in width, and extends as a swamp, two or more miles in width, for twelve miles to the village of Conewango Valley. With similar width the ancient valley continues southeast fourteen miles, through Randolph and Steamburg, to the Allegheny River. It was the course of the preglacial Allegheny River.

In singular and unexpected manner the Conewango Creek deserts the broad ancient valley three miles northwest of Randolph and enters a narrow, steep-walled valley leading southwest, through Kennedy village, for six miles to Poland Center, where it joins the Cassadaga Creek in another very wide ancient valley, east of Jamestown. The geography is all mapped on the Cherry Creek, Cattaraugus, Jamestown and Randolph sheets.

These valley plains should be distinguished from the smooth areas filled and leveled by standing waters or lakes. Examples of the latter are seen on plate 28 in the towns of East Otto (Cattaraugus sheet); and Ashford (Ellicottville sheet). Also in the terraces along Ischua Creek (Franklinville sheet).

On plate 29 lake plains are seen in the town of Collins (Eden sheet), and north of Cattaraugus Creek in Sardinia (Springville sheet). As in the last example the lake plains may adjoin or blend into the outwash plains.

Recognizing water as the agent in eastern America producing fluid levels we must discriminate at least three classes of plains, according to their manner of origin or their genesis; the outwash plains, lake plains, and valley plains.

Existing Lakes

Lakes are ephemeral features and cannot occur in normal drainage. They are an effect of drainage blockade, and must be young, speaking geologically, as they face ultimate extinction. The myriads of lakes and lakelets in northern lands are an indirect effect of recent glaciation (paper 16, pages 185–194).

Lake Chautauqua is a "morainal" lake, due to drift damming, with elevation 1,308 feet. It occupies a shallow basin, a partially filled valley, which before glacial interference drained northward; as did all of the streams of the region. The glacial flow, forced southward from the escarpment moraine, produced the swampy tracts north of the present lake (Westfield and Dunkirk sheets).

Except Chautauqua the ponded waters are small in size and relatively few for the large area. Most, if not all, of them belong in the singular category of "kettle-basin" lakes. These basins were primarily occupied by blocks of ice, detached from the ragged front of the ice sheet, and were buried or enclosed in the gravel and sand swept out of the glacier. The eventual melting of the buried ice, and the consequent slumping of the detrital cover and enclosing walls, produced the "kettles." Multitudes of the kettles hold no water.

Kettles and kettle lakes are characteristic features of morainal tracts, occurring especially in outwash plains. They are well represented in the plains at Machias; between Yorkshire and Protection; and at Springville and East Concord.

The Cassadaga Lakes (Dunkirk sheet) are good examples of kettle lakes in wide outwash plains. Bear Lake has the same origin and character (paper 9, plate 38).

Eastward, on the Cherry Creek sheet, is Mud Lake, 1,380 feet elevation; and East Mud Lake, 1,330 feet, both in the headwaters of the western branches of Conewango Creek.

A group of lakes lie near Sandusky (Franklinville sheet), and Crystal Lake, three miles from the village, has elevation of 1,761 feet. Two lie in Elton Creek drainage. Lime Lake at Machias, probably the largest kettle lake in the area, has elevation 1,631 feet.

Java Lake, on the Arcade quadrangle, 1,651 feet altitude, is the head of Cattaraugus Creek. In the moraine, five miles southwest of Bliss, the Arcade sheet depicts a multitude of kettles, both dry and watered, having elevations up to 1,900 feet.

Lake Erie Plain

WIDESPREAD GLACIAL WATERS

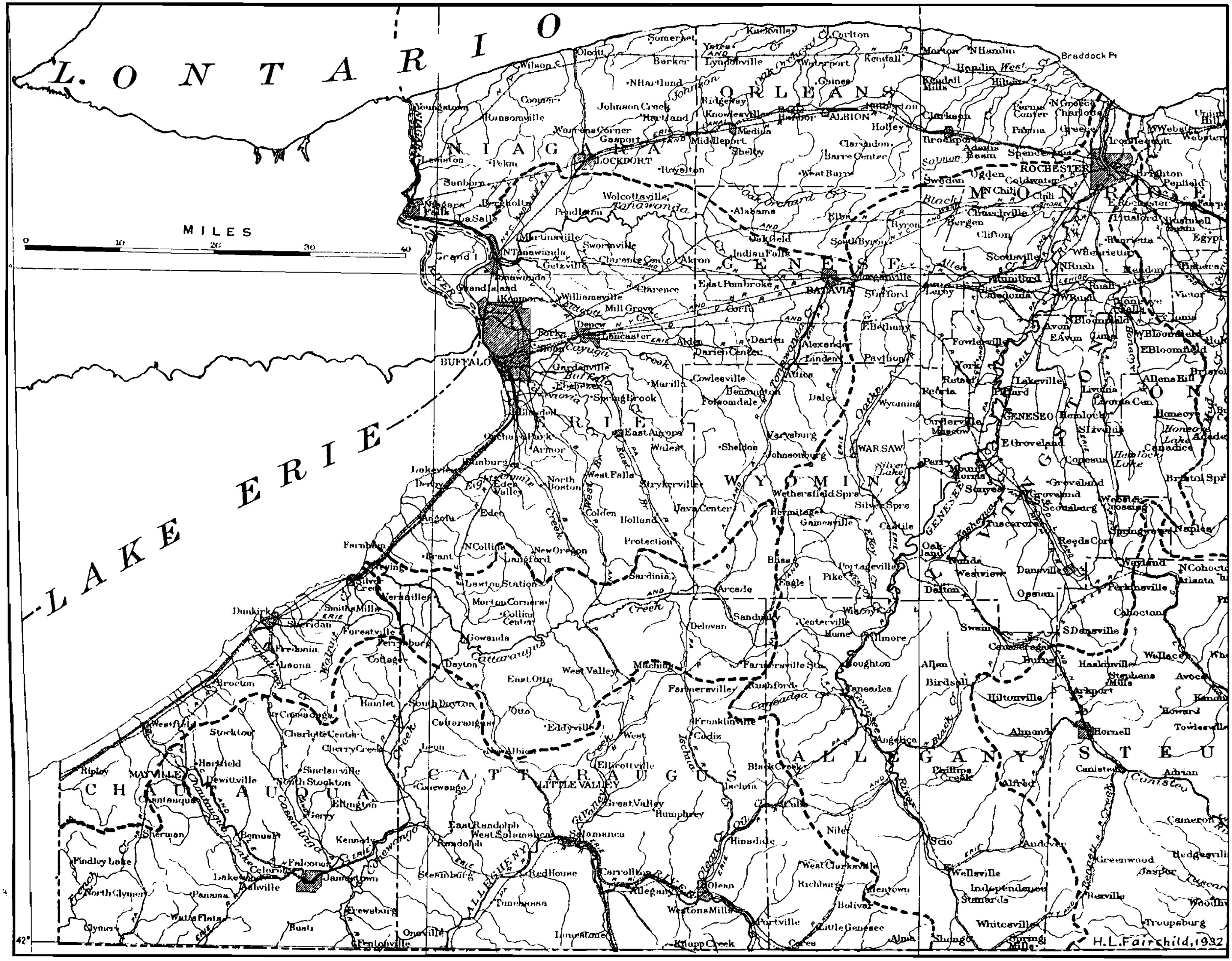
The New York records of the extended glacial lakes which occupied the Erie and Huron basins were described in paper 10 (year 1907) and the beaches mapped in six colored sheets. It is deemed undesirable to rehearse the facts and data given in that publication, but a change is made in the interpretation of the lake records as regards their succession and history. The more detailed description and discussion is published in paper 17.

The succession of glacial lakes in the former paper was given as lakes Whittlesey, Warren and Dana. Further study through the years of the glacial records, with some change in their interpretation, make demand for two lakes at the Warren level, and between them a long stage of deglaciation and free drainage eastward to the Mohawk-Hudson valleys. The glacial features which are concerned in this reinterpretation lie in the Ontario drainage area, and are discussed in paper 17.

Under the new view the lake succession in the Erie basin is as follows:

1. Lake Whittlesey; with westward outflow.
2. Lake Warren, the first; with westward outflow.
3. Downdraining of Lake Warren, eastward, with the outflow control on the salient, the Onondaga scarp, north of Batavia, followed by a long stage with no ice-impounded waters in the Great Lakes area.
4. Lake Warren, the second. Readvance of the ice barrier in central New York slowly lifted the imprisoned waters until they rediscovered the western outlet across Michigan, at the Warren level.
5. The second downdraining, eastward, but with the outflow controlled on the high ground in the Syracuse-Utica district.
6. Lake Dana; due to a long pause in the downdraining, with outlet at Marcellus.
7. Lake Erie. The incipient lake was very small, but has been lengthened and deepened by the deformative land uplift.

The shore inscriptions of Lake Warren, with spacing of the bars in their northern reach had been recognized, and was discussed in paper 10, pages 64–79. The present theory appears to explain



RIVER SYSTEMS AND DRAINAGE DISTRICTS IN WESTERN NEW YORK the discrepancy in the water planes by attributing the lower wave work, at least in part, to the second Lake Warren.

STREAM CHANNELS AND DELTAS

The multitudinous stream channels, remarkable in their relations and succession, which carried the outflow of the glacial lakes described above, are depicted on plates 2, 3, paper 10. They were initiated at the edge of the ice sheet, but many of them became land streams as the ice front withdrew. Some of the later channels in the district are shown in plate 29.

As far northeast as the village of Marilla, six miles northeast of East Aurora, the stream channels carried the flow and detritus into Lake Whittlesey. But at and beyond Marilla the streams contributed to Lake Warren. A few of the larger delta tracts are indicated on plate 29.

The drainage into the second Lake Warren, and the succeeding Lake Dana, was derived wholly from the land. This later stream flow was not so heavily loaded with detritus, and the delta deposits are not readily discriminated from the earlier ones.

Lake Ontario Plain

This stretch of plain, bordering Lake Ontario for seventy miles, between the Genesee and Niagara rivers, and with a width of twenty or more miles, is in striking contrast with the Erie lowland; with which it merges in the area east of Buffalo.

This plain carries one conspicuous glacial lake feature, the strong embankment beach of Lake Iroquois, from Lewiston to Rochester. The plain rises from 246 feet, the Lake Ontario level, to 385 at Lewiston, 420 at Ridgeway and 435 at Rochester.

The beach of Lake Warren (papers 8, 11), with elevation of 880 feet, may be regarded as the southern limit of the Ontario plain.

During the downdraining of the first Lake Warren (paper 17) the plain was mostly buried under the glacier. A portion on the south and west was covered by the second Lake Warren; and a larger area by lakes Dana and Dawson. A few drumlins in the district north of Batavia rise high enough to carry the wave work of the Dana waters, at about 700 feet, as partly mapped on plate 2, paper 11. The northern belt was covered by the shortlived Lake Dawson, but the shore features are weak and have not been traced.

The Ontario plain is not traversed by any preglacial river channels. As shown in figure 2 the Tertiary drainage in this province was held to east and west directions in the depression caused by the thick and nonresistent Salina shales; the same as east of the Genesee Valley. Today this depression is occupied by the Tonawanda and Black creeks. The only trenching of the Niagara formation (Lockport limestone and Medina sandstone) is by the Irondequoit and Sodus valleys (figure 2).

This Ontario plain was smoothed by the friction of the ice sheet, with flow southwestward. In its latest flow it rubbed the till mantle into great ribs, or flutings, producing a drumlinized surface. This is described in two papers on drumlins, *N. Y. State Museum Bulletin* No. 127, and *Proceedings of the Rochester Academy of Science,* volume 7, 1929, pages 1–37.

Allegheny Drainage Area

The portion of this area which is included in the Allegany State Park is described in paper 21.

The southward stream flow of this area is due to glacial interference, which has wholly reversed the preglacial direction, as shown in figure 2. The rivers in Tertiary time were tributary to the Erie Valley. The new divides are morainal valley fillings.

The boundary of the area is roughly semicircular; the divide on the west lies over against Lake Erie; on the north against the Cattaraugus Valley; and on the east against the Genesee Valley.

When the ice front receded from the terminal moraine, figure 3, it was very deliberately, by pauses and readvances. The more effective standstills produced the valley fillings, thereby establishing the present water partings.

Three classes of the glacial features are conspicuous and important. These are, the existing lakes; the outwash plains and valley plains; and the records of the vanished lakes.

GLACIAL LAKES

The term "glacial lakes" denotes the water bodies, small or large, that were held up or impounded by the glacier, which served as a movable dam. Except the Marjelen See, in Switzerland, none are recognized today. The innumerable existing lakes and tarns due indirectly to the ice sheet are properly classed as morainal or drift-barrier, kettle and plunge-basin.

While the glacier was lingering in the Allegheny area the front of the ice sheet during its slow rescession, and especially in its standstills, held lakes in the valleys which declined northward, or toward the ice. The more definite and extended glacial waters were in the Conewango Valley; the South Branch of the Cattaraugus; and the valley of Ischua Creek.

Probably the earliest of the glacial valley lakes was in the Ischua Valley. The Franklinville sheet shows level areas, mostly the deltas of side streams, at Fitch, 1,680 feet; at Cadiz, 1,600; north of Franklinville, 1,620; at mouth of Johnson Creek, 1,680; at mouth of Bear Creek and north to Gulf Creek good terraces at 1,680 to 1,720. Similar levels occur about Machias village, at East Machias School, and in the Elton Valley.

The standing water which produced these level stretches may be named the Franklinville glacial lake. The blockade, either drift or rock, or both, or perhaps drift-buried ice, must have been in the very narrow valley at Ischua Village, with present elevation of about 1,530 feet; and also on the west in the constricted valleys at Devereaux and Ashford, some six miles northeast of Ellicottville.

The valley of Elton Creek, holding the villages of Elton and Farmersville Station, appears to have been flooded by the Franklinville lake through the valleys southwest of Elton Station.

Eventually the river flow across the barrier at Ischua Village eroded and lowered that outlet sufficiently to drain the Franklinville Lake. Evidence of this event in the history is found in a definite stream channel cut by glacial flow along the east side of the valley at Machias Junction. This channel is described by Mr. H. W. Clough,[1] in a letter, as heading at Lime Lake and extending along the east wall of the valley over three miles, and blending into the open valley a mile southeast of School No. 1. The Franklinville sheet gives the elevation of the channel on the divide, in a "swamp col," as 1,640 feet. The lake terraces on the west side of the valley below School No. 1 are 1,720 feet, and over, in elevation.

This Machias outlet channel implies a lake on the north. The geographic relations suggest that the upper portion of the Catta-

[1] The writer is indebted for valuable data relating to the Cattaraugus region to Mr. Clough, a student of the writer in far-gone days, with a life-work as meteorologist in the United States Weather Bureau, now retired, with summer residence at Arcade.

raugus Valley was, for a time, drained through the Machias outlet. In this view we may name the lake the Arcade glacial lake. It belongs in the glacial lake succession of the Cattaraugus Valley, recounted below.

The two swamp divides, five miles north and northeast of Ashford, on the Ellicottville sheet, and called "Beaver Meadows," indicate ponding of water in the valleys on the north.

Going west to the Conewango drainage area we find evidence of ice-impounded waters. When in the waning of the glacier the ice front had receded north of Randolph a lake was held in the valley of Little Conewango Creek, which we will call the Randolph lake. The barrier and outlet was at Steamburg, with the present elevation 1,420 feet. With further recession of the ice front the Randolph waters found lower escape through the narrow side valley in which rest the villages of Waterboro and Kennedy (Jamestown sheet).

As the ice front backed away the Randolph lake became the extended Conewango glacial lake. Eventually this lake received the overflow of the Cattaraugus glacial lake, by the outflow channel at Persia station on the Erie Railroad, three miles southeast of Dayton. This channel was described in paper 10, page 17.

The heavy outwash of rock-rubbish from the ice sheet, and the copious inflow of the many creeks draining large territory and sweeping in detritus for many thousands of years, accounts for the valley filling and the swampy condition. The creek is now a sluggish, serpentine flow, hundreds of feet above the bed of its preglacial ancestor. State drainage ditches now help to reduce the watery condition of the great valley.

The later glacial waters are described in the chapter on the Cattaraugus Valley.

The above account of the Allegheny area is only an outline. The varied and complex features have much of geologic and geographic interest and will well repay careful, detailed study.

Cattaraugus Valley

CREATION OF THE RIVER AND VALLEY

This valley is remarkable in its complex origin, its singular physiographic relations, its rock ravines and its glacial lake history. The area is covered by eight topographic sheets; Silver Creek, Eden, Springville, Arcade, Cherry Creek, Cattaraugus, Ellicottville and Franklinville. The features and relations which can be shown in black and white are seen on plates 27–29.

Plate 29 joins directly on plate 28 and the two cover all of the Cattaraugus drainage area except the seven miles below Versailles, which is shown on the Silver Creek sheet. The central and more critical part of the valley and river is shown on plate 28. The head-water part of the valley, from east of Springville to Java Lake, a stretch of twenty miles, is shown on plate 29. The lower stretch of the valley, below Gowanda, which was in the course of the pre-glacial Allegheny River, is partly displayed on the two plates.

The topography of the upper valley suggests that in preglacial time a water parting existed at East Arcade. North of this divide the flow was, probably, to the Tonawanda valley. Below East Arcade to Arcade and Yorkshire the tributaries appear to lie in their preglacial courses. But below Yorkshire the river is postglacial.

In its westward course the new river lies across the deep pre-glacial valleys that led northward, as shown in figure 2. During the removal of the ice sheet those ancient north-leading valleys were blocked, at first by the receding front of the ice sheet and then permanently by the morainal fillings. The escaping waters were forced to westward flow along the ice margin and took possession of some east and west valleys which had been tributary to the north-flowing rivers. In three stretches the glacial flow was compelled to cut across intervalley highlands, thus producing the deep rock canyons (plate 28).

The wider, more open, portions of the valley represent preglacial stream work, while the narrow, steep-walled rock ravines register the new, postglacial erosion by the newly-created river. The canyon section of the South Branch also represents diversion of stream flow. Careful study of deep borings may determine the old lines of drainage.

One of the two intersections of the preglacial, north-leading valleys is seen in the wide plain of water-leveled moraine and glacial outwash at Chaffee, Sardinia and Yorkshire (plate 29). This area is part of the great valley which headed at Ischua (plate 28) and passed northward as the East Branch Cazenovia. At Springville and Cascade Park is another great drift filling and pitted-plain, evidently related to the valleys of Buttermilk and Eighteen mile creeks. These extensive water-smoothed plains, with their kettles and lakelets, are the most unusual and interesting of the glacial features.

The courses in direct line of the Connoisarauley and Derby creeks with the South Branch Eighteenmile Creek is suggestive of a preglacial north-leading valley deserving of study. In this relation the valley of Spooner Creek must be considered.

The rock ravines will be described in a later chapter.

The quite direct course of the Cattaraugus Creek from East Arcade to the junction of the South Branch, a right-line distance of thirty miles, is a striking feature, but probably fortuitous.

The windings of the river in the rock ravines are "intrenched meanders" that were acquired when the stream was lazily swinging on detrital plains overlying the rock strata.

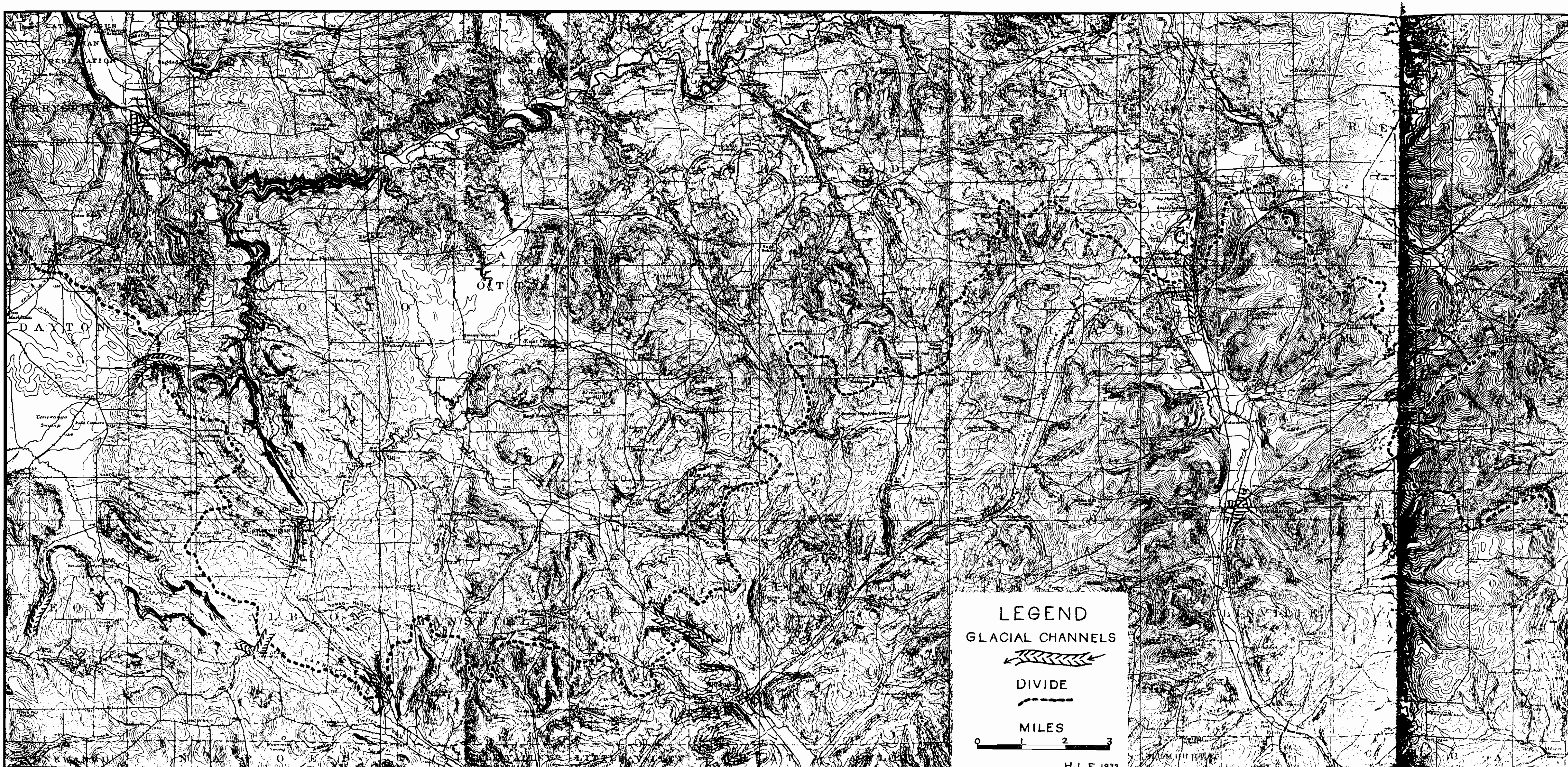
GLACIAL LAKES

The Cattaraugus Valley declines toward the southwest. The front of the waning ice sheet receded westward, and consequently the valley held ice-impounded waters, glacial lakes, in the main valley and its tributaries.

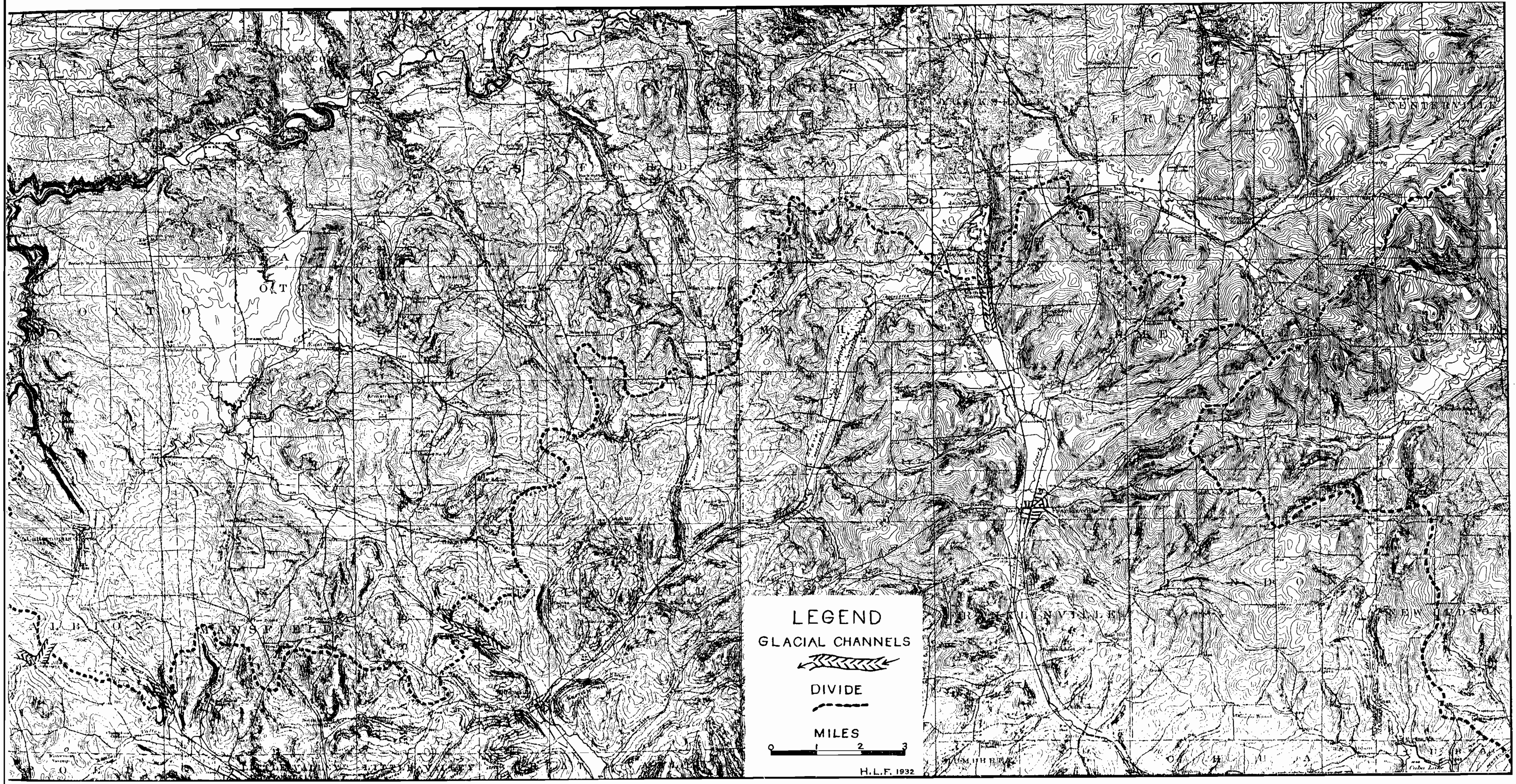
The lake history is complicated and not easily translated. In the production of the new valley the north boundary, created as described above by ice and stream deposition in the preglacial north-leading valleys, so blocked the valleys that even today there is no northward flow from the Cattaraugus basin. Except during the latest presence of the ice front all of the outflow was across the southern divide into Allegheny drainage (plate 28). That divide was uncovered by the ice removal from east to west, and the low passes permitting outflow of the ice dammed waters were opened successively westward. The height or elevation of the lakes and their extent were determined by the elevations of the points of outflow. The several outlets are indicated on plate 28.

Passing westward along the crooked line of water parting between north and south flow no pass with elevation lower than the Machias outlet, of 1,636 feet, is found until we travel fourteen miles in direct line and reach the head of Mansfield Creek, two miles northwest of Ellicottville. Here a pass leads southeast to Great Valley Creek, with elevation 1,620 feet and with form and relation indicating copious flow of water.

Further westward, seven miles in direct line, another pass ten feet lower is found two miles northwest of Little Valley (Cat-



CATTARAUGUS VALLEY GLACIAL WATERS AND THEIR OUTFLOW CHANNELS



CATTARAUGUS VALLEY GLACIAL WATERS AND THEIR OUTFLOW CHANNELS taraugus sheet), but with the form less favorable. The Erie railroad found this pass available in passing from the Allegheny basin over to the Cattaraugus.

Five miles northwest of Little Valley, and three miles south of Cattaraugus village, a much lower pass occurs at the village of New Albion. This outlet has elevation of 1,440 feet and opens to the head of Conewango Creek. But the outflow by this pass was forced through the deep canyon west of Kendall Corners to the Mud Creek, tributary to the Conewango, until the ice front had moved away from the Wells Hill, six miles south of Dayton. The features are shown on the Cattaraugus sheet.

The westernmost and lowest pass is found at Persia flag station on the Erie railroad, with elevation 1,320 feet. This outlet, the latest across the southern divide, is three miles southeast of Dayton, leading to the wide Conewango Valley.

The next and the final escape for the Cattaraugus glacial waters was five miles west of Gowanda and northwest and north from Perrysburg, appearing on the Cherry Creek sheet and in plate 3 of paper 10. The outflow began at 1,300 feet, in the channel followed west of Perrysburg by the Erie railroad, while lower channels permitted outflow down to the level of Lake Whittlesey, 850 feet. Three interesting channels south of West Perysburg have elevation of 1,200 feet.

In review, these several outlets of the glacial waters are:

Machias,	elevation	1,640 feet.
Ellicottville,	"	1,620 "
Little Valley,	"	1,610 "
New Albion,	"	1,440 "
Persia,	"	1,320 "
Perrysburg,	"	1,300–850 feet.

The several outlets with the falling levels gives us the following theoretic lake succession, all of which, with exception of the last one, contributed their overflow to the Allegheny River.

1. Arcade-Yorkshire-Delevan lake. The outlet was at Machias village, with elevation about 1,640 feet, leading to Ischua Creek. The water in this stage filled the upper stretch of the valley nearly to the river's head in Java Lake, and the lower portions of the tributary creeks. The Arcade and Franklinville sheets depict the features.

2. Springville Lake. This occupied the middle stretch of the main valley, and the tributary Buttermilk and Connoisarauley creek valleys. Apparently the lake had outlets at Ellicottville, 1,620 feet, and at Little Valley, 1,610 feet.

3. South Branch Cattaraugus Lake. This three-branched water body had its central point at Cattaraugus Village. Perhaps its early outflow was at Little Valley but its chief outlet was at New Albion, elevation 1,440 feet, over to the Conewango Valley (Ellicottville and Cattaraugus sheets).

During this stage a pass at Brooklyn School, three miles north of East Otto Village and three miles southeast of Zoar Bridge, at 1,360 feet, appears to have given connection with the water in the main valley above Cascade Park. The intrenching of the rock canyon below Zoar Bridge had not begun in this stage of the history; but the lake deposits which completed the fillings in the intersections at Yorkshire and at Springville probably began during this stage.

4. Main Cattaraugus Lake. The outflow was by the Persia channel, elevation 1,320 feet, over to the wide Conewango Valley. This was probably the longest in life of the several lake stages, with larger area, and more copious outflow. The higher lake plains and the deltas by tributary streams were completed during this stage. Some of the features are described below.

5. Gowanda Lake. This was the closing stage of the glacial waters. It existed while the ice front weakened and slowly receded on the north-facing slope at Perrysburg and West Perrysburg, from elevation of 1,300 feet down to 850 feet, when the water blended with Lake Whittlesey.

This falling level did not occupy the Cattaraugus Valley above the district of Gowanda because the canyon above Gowanda had not been cut. But all the water of the basin passed through the Gowanda lake, reaching the latter by cascades or cataracts.

The first and highest of the Perrysburg scourways, produced by the escape of the Cattaraugus waters along the ice front, is followed by the railroad that used Persia channel, five miles on the southeast. The elevation should be less than the bottom of the Persia channel, and is shown by the Cherry Creek sheet as about 1,300 feet. Plate 3 in paper 10 shows the succession of downdraining channels from

1,300 feet down toward 900 feet, where the Cattaraugus waters blended with the great glacial Lake Whittlesey in the Erie basin. The wide plain of the Cattaraugus Indian Reservation and of the State Asylum are records of the lower lake waters, from 900 to 800 feet (plate 28).

6. Land-locked Cattaraugus Lake. When the glacial water in the Gowanda stage had lowered to about 1,200 feet, by the channels near Perrysburg, the water above Zoar Bridge was held in by the rock barrier. This lake was not, like its predecessors, a glacial or ice-barrier lake, but a rock-barrier lake.

This lake has long since been drained by the downcutting of the rock dam and the production of the great gorge or canyon above Gowanda. The lake was initiated while the ice margin lay on the ground northwest of Perrysburg, but its life, of tens of thousands of years, was mostly in postglacial time. Although the lake disappeared long ago the canyon erosion is yet in progress. The other rock gorges up stream, and the one in the South Branch, have similar and contemporaneous life history.

The curvatures or meanders in the rock sections were established before erosion of the rock began. They were formed on detrital plains when the new valley was filled with glacial and stream deposition to perhaps 1,300 feet elevation, at the close of the glacial history in the fourth stage, while the water was held up to the Persia outlet and the highest of the Perrysburg scourways.

ROCK RAVINES

The three ravines or canyons in the course of the Cattaraugus Creek (plate 28) mark the stretches where the stream was compelled to carve entirely new paths. In the intermediate open stretches of its course the stream, under control by the glacier front, discovered old valleys that had been tributary to the north-flowing rivers.

The surface of the ancient rock plain, of upper Devonian strata, appears to have been quite uniform in the district, as the ravines are similar in elevation and depth. The dimensions are approximately as follows:

Upper ravine, Cascade Park, two miles long.

Elevation of top of rock section ..... 1,280 feet.
Elevation of the stream .... 1,080 feet.
Depth of ravine .... 200 feet.

Middle ravine, below Frye Bridge, four miles long.	
Elevation of top of rock section	1,200 feet.
Elevation of stream	1,000 feet.
Depth of ravine	200 feet.
Lower ravine, below Zoar Bridge, five miles long.	
Elevation of top of rock section	1,200 feet.
Elevation of stream	900 feet.
Depth of ravine	300 feet.

The carving of the ravines by stream erosion is in postglacial time, for the district, and is a measure of the uncounted years since the ice sheet melted from that locality. The cutting of the rock did not begin until the ice-impounded water in the Cattaraugus basin and the newly-created river valley had fallen to the level of the shale and sandstone. As that level is today about 1,200 feet while the Persia outlet (plate 28) is 1,320 feet it is apparent that the ravine cutting is subsequent to the lake outflow at Persia flag station. The lake water was lowered, and the river encountered the rock obstructions when the outflow was held against the north-facing slope west of Perrysburg. At that time, as noted above, the water held in the Cattaraugus Valley had become a land-locked lake.

A striking feature is the winding course of each ravine. They are excellent illustration of "intrenched meanders." The meanders antedate the initiation of the rock cutting, having been established on the detrital plains which overlay the rock surfaces. The drop in the lake surface with the desertion of the Persia outlet caused the river to rapidly deepen its winding channel. The curvatures in the rock ravines are an inheritance from the meanders in the superimposed detritus.

Of course the deepening of the middle and upper ravines was dependent on the downcutting in the lower ravine, above Gowanda. And the erosion is yet in slow progress.

LAKE PLAINS

Without much careful exploration and study, with precise measurements, the history and records of the earlier lake stages in the basin of the Cattaraugus cannot be described in detail. An interesting task for some enterprising geologist is the examination of the outlet passes on the southern divide, with determination of the correlating features and of the history in detail.

Beginning with stage 4, control by the Persia outlet, there is more of certainty and much of interest. The present elevation of the Persia outlet after its downcutting is 1,320 feet. The surface of the river in this channel determined the elevation of the impounded waters in the basin. The Cattaraugus topographic sheet shows extended smooth plains at 1,320 to 1,360 feet in the towns of Otto and East Otto. The valley plain at the village of Cattaraugus is 1,320 to 1,340 feet.

Examination of the topographic sheets will show that the plains, indicated by the white spaces, rise in elevation to the northeast, attaining 1,400 feet at Sardinia and Yorkshire. This increase in altitude is due in part to the postglacial uplift of the land, upslanting toward two feet per mile in the northeast direction. In the twenty-five miles of rightline distance from Persia to Sardinia the deformation may be toward fifty feet. If the outlet river at Persia had in its summer floods a depth of ten feet that alone carries the lake plane to 1,380 feet.

The wide plains at Sardinia-Yorkshire-Chaffee were built by the glacial outwash from the Cazenovia and Buffalo valleys into the Cattaraugus waters, and the abundant detritus was probably spread out, at Chaffee, above the lake level. At Arcade the delta filling, to 1,500 feet, was by the detritus of the upper stretch of the Cattaraugus Creek, and also by Clear Creek.

Another great outwash plain extends from Cascade Park through Springville to East Concord, first by glacial outwash and later by the outflow from the valleys of Cazenovia and the two Eighteenmile creeks. The elevations are 1,300 feet at Cascade Park, rising to 1,400 feet at East Concord (Ellicottville and Springville sheets).

The tributary valleys were also filled with detritus at the Persia level. A very handsome display is seen in Ashford Town (Ellicottville sheet) extending up the wide valley of Buttermilk Creek, and west to the Connoisarauley Creek, with elevations 1,300 to 1,360 feet.

During the thousands of years that the Cattaraugus river has been cutting the rock canyons detrital plains and deltas were forming in the land-locked lake by inwash by the tributaries, with slowly falling levels, from 1,200 feet down to the present creek.

The South Branch Valley has a similar history, produced by the rock barrier at Forty Bridge, with corresponding altitudes.

In the slow deepening of the ravines the open-valley stretches above the ravines were always filled with stream detritus at the level of the rock channel downstream. With the deepening of each rock channel the stream was correspondingly lowered in the up-stream open-valley stretch, producing terraces and benches at various levels in the abandoned floodplains. These are abundant and conspicuous above Cascade Park in the fourteen miles to Arcade, with elevations from about 1,460 feet down to 1,100 feet, the present elevation of the river at the Park. An excellent example is found on the north side of the Creek, northwest of The Forks and south of Sardinia (southeast corner of the Springville sheet).

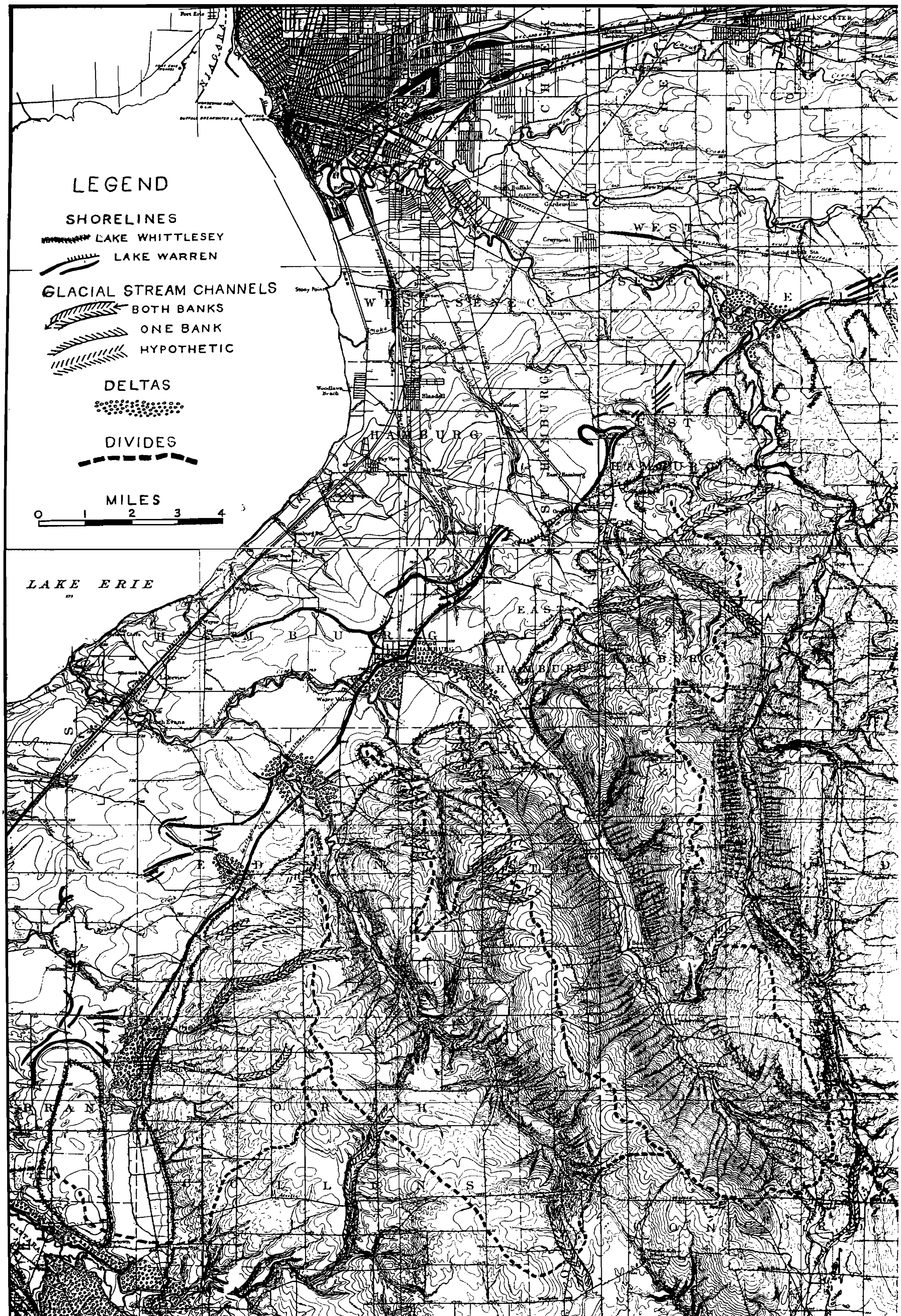
Below Gowanda, and beyond the control by the rock dam up-stream, the wide old valley and adjacent land were yet under the glacial waters, controlled by the Perrysburg scourways. An excellent evidence of the slowly falling waters is seen in the drainage area of both branches of Clear Creek, in the town of Collins and North Collins (Cattaraugus and Eden sheets). Marshfield is on a plain at 1,320 feet. Westward, down stream, the beautiful plains and terraces decline to 850 feet, the level of Lake Whittlesey; while lower plains represent glacial Lake Warren, 780, and down to Lake Erie, 573 feet. The interesting succession of plains and terraces in the Gowanda region are described in paper 9, pages 137–139 and in paper 10, page 38.

North-Leading Parallel Valleys

ORIGIN AND CHARACTER

The remarkable series of parallel valleys in central New York, tributary to the Ontario Valley, have long been famous. The westward representatives of the great sisterhood of preglacial valleys, the ones leading to Lake Erie, have been given small attention, although of much physiographic and geologic interest. The neglect has been due partly to the absence of lakes, partly because the valleys are deep and narrow with less advantage for occupation and agriculture. Also the area has high relief and few important villages.

The Erie basin series include eight valleys with general decline northwestward. From west to east they are: the lower portion of the Cattaraugus, which was the course of the Tertiary Allegheny; South Branch Eighteenmile; main Eighteenmile; West Branch



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PARALLEL VALLEYS AND GLACIAL WATER

LEYS AND GLACIAL WATERS NORTH OF THE CATTARAUGUS VALLEY

Cazenovia; East Branch Cazenovia; Hunter; Buffalo; Cayuga. East of these the Tonawanda is a connecting link between the Erian and Ontarian groups, the creek having a course east of north and then west into Niagara River.

These open valleys are remnants and relics of much more extended valleys of Tertiary drainage. As with the valleys of central New York the former southern or headward portions and the northern, terminal portions of most of them have been obscured or even buried by the work of the ice sheet and the glacial lakes. West of the Cattaraugus the heavy moraine deposits on the scarp of the plateau have buried the channels of northward flow.

This group of valleys is an inheritance from Tertiary time, when all of the drainage of central and western New York, including the Allegheny River, passed northward (figure 2). The streams which carved these valleys performed the same function, and at the same time, as the Ontarian tributaries, namely, the removal of the precipitation from the northern scarp of the Allegheny Plateau. And, like the Ontarian group, they have been dissected or beheaded by moraine filling.

As these deep parallel valleys decline to the northwest, and as the blockading ice front lay across them transversely, the outflow from each valley was into the adjacent valley on the west. In the case of two valleys, Hunter and Cayuga, the water was held up to confluence with its neighboring lake, or lakes.

In the precise study of the elevations of the outflow channels and the planes of the lake surfaces it is necessary to take into account the postglacial land uplift, with the northward uptilt of about two feet per mile; and the lowering of outlets by the stream erosion.

MORAINES AND VALLEY BLOCKADE

The moraines in this area were mapped by Frank Leverett in paper 5, plates 19, 25, and described in pages 651–684. In a general way they are shown in figure 3, but require further study.

The ice front acting as a barrier, along with the moraine fillings which it piled in the valleys, produced the Cattaraugus Creek, which cuts across and has beheaded the longer preglacial valleys.

The outwash plains facing the valley fillings are extensive, and pitted with kettles and ornamented with lakelets. They are shown on the Eden, Springville and Arcade sheets. They are unequalled in New York; but kettled plains in comparison are the delta plains along the east side of the Black River Valley, described in the State Museum Bulletin 160.

The valley fillings along the north side of the Cattaraugus Valley are part of Leverett's "Escarpment Moraine System," described in pages 651–672 in paper 5, and partly mapped in the plate 19. It will be seen in the new moraine map, figure 3, that this moraine correlates with the "Valley Heads" moraine in the central part of the State, and has similar relations. In both areas the drift belt is poorly developed between the valleys, across the intervalley ridges.

The moraine deposits, the wide and smooth outwash plains blending with the Cattaraugus Lake plains, and the valley lakes and their outlets, described below, make a very interesting geographic and geologic complex.

BURIED VALLEYS

An interesting element in the study relates to the buried or obscured lines of the preglacial drainage, especially the southern or headward portions. With even the present limited data and information it is possible to locate and trace some of the drift-filled valleys, having in mind that the northern divide of the Cattaraugus basin is wholly of post-glacial origin, but that the southern divide mostly dates from preglacial time.

Excepting the valley of Hunter Creek it appears probable that all of the deep north-leading valleys north of the Cattaraugus had their original headings south of the present divide; and some of the old drainage lines may be confidently inferred.

It appears quite certain that the stretches of the Cattaraugus with wide fillings of lake deposits, as at Zoar, at Cascade Park and Springville and at Yorkshire-Sardinia are filled sections of ancient north-leading valleys. The directions and relations of the unfilled portions of valleys north and south of those filled areas clearly indicate the paths of some of the Tertiary rivers. Detaching, temporarily, from its binding the map, plate 28, and placing it in proper juxtaposition with that of plate 29 the above features appear in their true relationship.

Perhaps the clearest of the old valleys is that which had its head in the constricted notch or col at Ischua Village (Franklinville and Olean sheets). Its northward course is well marked past Franklinville, Machias, Delevan, Yorkshire, Chaffee, and the now-open stretch from Protection to East Aurora; a distance of forty miles. North of East Aurora the old valley is obliterated, as are all of the old drainage lines on the Erie lowland. As a name for the Tertiary valley we may favor that of the deep, open stretch and call the stream the Preglacial Cazenovia River.

Passing westward we find that Buttermilk Valley, heading near Ashford, is in line, through the valley filling at Springville, with, probably, the Eighteenmile Valley; a length to Hamburg of about thirty miles. The close relation of the West Branch Cazenovia requires study of the rock exposures and of any deep borings.

Apparently the valley of the east Beaver Meadows also headed near Ashford, and was tributary, at Machias, to the ancient Cazenovia.

The Connoisarauley Valley, heading near the hamlet of Plato, appears to connect with the South Branch Eighteenmile Creek. The length to Eden Valley is about twenty-five miles.

The Conewango Valley was long ago (paper 1) shown to connect with the lower Cattaraugus as the Tertiary path of the Allegheny.

The lines of the ancient, preglacial stream flow can be positively and fully determined only by systematic deep drilling, and study of the well records along with the observable rock outcrops. Perhaps there may now be considerable available data for the use of some enthusiastic student of the local geology.

On plate 28 the headings of some minor lines of the old drainage may be noted. The site of the outlet channel northwest of Ellicottville was originally a divide between a tributary of the Allegheny and the present valley of Mansfield Creek. This suggests continuation through the towns of Otto, crossing of the Cattaraugus at Zoar, with probable continuation northwestward to the old Allegheny.

The channel three miles northwest of Little Valley was the location of a col which headed the valley holding Cattaraugus Village. Apparently this drainage line passed northwest toward Gowanda as an Allegheny tributary.

On plate 29, an ancient col at East Arcade was doubtless the head of a north-leading valley which was tributary to the Tonawanda Creek, and was the actual head of that drainage system. Only the stretch at and south of Java Lake remains unfilled.

GLACIAL LAKES IN VALLEYS NORTH OF THE CATTARAUGUS

The territory including these valleys is mapped on the Buffalo, Depew, Attica, Eden, Springville and Arcade sheets. The first three of the sheets, in part, and the last three are combined to form plate 29, with reduction in size.

All of these valleys decline northwestward, and consequently were effectively dammed by the receding ice front. The earliest impounded waters, the primitive lakes, had outflow southward, across the moraine fillings, into the newly-created Cattaraugus Lake. Later outflow was westward, across the intervalley ridges, with the ultimate escape into lakes Whittlesey and First Warren. The abandoned, cross-ridge channels are evident on the topographic sheets and conspicuous and fascinating in the field. They are fairly indicated on plate 29. The latest and complex outflow channels were long ago mapped in plates 3, 5 of paper 10.

These glacial lakes will be described in order passing from west to east.

New Oregon-Clarksburg Lake

In the South Branch Eighteenmile Creek

The area is wholly shown on the Eden sheet, and in plate 29. The primitive outflow at the valley head was apparently across the outwash, at 1,420 feet, into the Cattaraugus waters by Derby Brook. The outwash plain lies between Concord and Morton Corners.

The earliest well-defined lateral outflow was a mile south of Langford corners, at 1,240 feet, over into the North Branch of Clear Creek. This flow contributed to the plains and terraces along Clear Creek.

Another main outlet of the lake was five miles farther north and only a mile from the present stream, at the head of Franklin Gulf, 1,120 feet. The canyon form of the channel attests a copious and erosional outflow, which contributed to the Whittlesey beach and delta at North Collins.

Close study on the ground in this district will doubtless locate several scourways across the divide at intermediate elevations between the three outlets here described. Some of the unmapped outlets may correlate with the fragmentary stream channels shown in plate 4, paper 10.

The final downdraining was east of Eden Village; first by a channel at 1,000 feet, and later on a north-facing slope into Lake Whittlesey. See plate 4, paper 10.

Boston Lake

In the West Branch Cazenovia Valley

The earliest waters were probably confluent with the highest Cattaraugus waters, producing the extensive kettle-pitted plain north of Springville and west of East Concord. This level, 1,400 feet, must have existed until the ice front receded on the western divide some nine miles. Southwest and west of Boston Center escape was found at 1,400 down to 1,300 feet. Below this the control was west of the Hampton Valley, at East Eden. Here a set of channels begin one half mile south of the corners, at 1,140 feet and continue north and northwest of the village down to 1,000 feet. At this lowest level control was also held a mile west of North Boston. These channels were mapped on plate 4, paper 10.

Two miles north of North Boston the lake waters blended with those of Lake Whittlesey at or just under 900 feet.

TRANSPOSE

Glenwood-Colden Lake

In the Main Valley of Eighteenmile Creek

This is shown on the Springville sheet. The valley heads, like the former one, in the outwash plain at East Concord; and the primitive outflow appears to have been along the path followed by the Buffalo, Rochester & Pittsburg Railroad, at 1,420 feet.

The water in this valley was confluent with that of the Boston Lake, on the west, through the cross-valley pass two miles southwest of Colden, now occupied in part by Landon Brook. But when the Boston Lake was lowered about 200 feet this pass became the outflow channel. The swamp col has elevation of 1,170 feet.

This valley, like the others of the series, has walls too steep to hold large and conspicuous delta plains on its slopes. But some of the inflowing brooks should have built deltas that can be recognized in the field, and the altitudes definitely measured. Such small deltas are the very best criteria for determining the water planes and elevations.

The next lowest escape for the Colden waters was far northward, no pass below 1,080 feet being found until we reach Loveland, northwest of Jewettville and Griffins Mills. Here is a complex of channels, and one followed by the B. R. & P. RR. leading northwest curves around to southwest as a scourway opening to Lake Whittlesey.

The intake of the Loveland channel is 1,020 feet. Three miles north, and three miles west of East Aurora, a channel at 900 feet elevation leads west and joins the Loveland channel.

Holland Lake

In the East Branch Cazenovia Valley

The creek heads in several twigs in the moraine and outwash plain at Protection Station and Chaffee Village. The channel of primitive outflow is a capacious scourway across the divide, one half mile east of the Briggs School and two miles northwest of Chaffee, leading over to Hosmer Brook. The elevation is 1,440 feet. The features appear on the Springville and Arcade sheets.

The lake existed at the level of its early outflow until the ice dam had backed away on the high western ridge, a distance of eleven miles, to near the latitude of South Wales (Springville sheet). Two miles west by south from South Wales is a channel entered by the Darling Road and curving around to Pipe Creek. The elevation of the channel bottom at the intake on the divide is about 1,310 feet. The form of the pass and channel and the Pipe Creek canyon indicate a heavy stream flow. Indeed it appears that the channel carried not only the outflow of the Cazenovia Valley but also the later outflow of the Hunter and Buffalo valleys on the east.

On the dividing ridge, declining northward, scourways occur at successively lower levels, four of which are indicated on the map, plate 29, before this East Branch Valley joins the West Branch at East Aurora, at the level of Lake Whittlesey, about 900 feet.

Colegrove Lake

Hunter Creek Valley

This deep valley lies between the East Branch Cazenovia and Buffalo valleys, on the Springville and Arcade sheets. It does not head in a moraine and outwash plain like the other valleys. When the head of the valley was released from the ice sheet it was flooded by water from the Buffalo valley through a pass two miles north of Dutchtown, with elevation of 1,370 feet. During that phase the water in the Hunter Valley was a branch of the Wales Hollow Lake, described below.

The first outflow from the valley was southwest of Colegrove and two miles northeast of South Wales, by a channel at 1,300 feet.

This channel and the succeeding ones also served for the final escape of the Buffalo Valley waters.

West and northwest of Colegrove are three more passes which lowered the waters from 1,300 down to 1,200 feet. East and northeast of East Aurora the lower channels belong to the next lake, and are on the Depew quadrangle.

Wales Hollow-Java Village Lake

Buffalo Creek Valley

As shown on the Arcade sheet and on plate 29 the Buffalo Creek heads in many branches in the towns of Holland, Arcade and Java in an extensive moraine. The longest branch, through Dutchtown, heads close to the longest branch of the East Branch Cazenovia Creek.

Mr. H. W. Clough has drawn attention to an interesting path of glacial drainage which appears on the Arcade sheet but is not indicated on plate 29. This is a smooth and level stretch extending northeast from Chaffee past Punkshers Corners and Java to near Java Center, a distance of seven miles, which carried the earliest drainage of the Buffalo Creek basin.

This belt is quite level, with elevation 1,500–1,520 feet, and apparently was smoothed as a scourway of glacial flow along the margin of the waning ice sheet. It carried the overflow of the glacial waters on the northeast, especially those of the Tonawanda Valley. This ice-border drainage immediately preceded the flow through the Gallagher Swamp channel, three miles northeast of Java Village, and it carried the earlier contribution of detritus to the great outwash plain at Chaffee.

The early flow in the Java Center—Chaffee scourway was pressed against the northwest faces of two hills, one southeast of Curriers and the other northeast of Punkshers Corners, and produced undercutting and oversteepening of the eroded slopes. This is shown by the close-set contours, 1,540 up to 1,700 feet.

A later escape of the glacial waters appears to have been through the moraine by a pass across the divide a mile north of Chaffee, at 1,480 feet elevation, leading to the outwash plain at Chaffee and Sardinia. This pass exhibits no channel features. It is the north portion of the outwash plain and carries many and deep kettles (Arcade sheet). The kettles were produced by the melting of blocks of ice which had been buried in the drift, and the melting did not occur until the ground was exposed to the atmosphere and leaching rains.

The history, the succession of events and lake conditions, is not evident and positive. An important, and uncertain, factor is the position of the ice margin in this area in its time relation to the ice-front position on the southern divide of the Cattaraugus Valley having control of the glacial lake waters. The Chaffee-Sardinia outwash plain was built in the higher glacial Cattaraugus waters; and the question now arises, how long in time and how far in northward distance did the Cattaraugus waters occupy the Buffalo Creek basin.

Evidently the Buffalo Valley waters flooded the Tonawanda Valley, on the east, by the pass at Gallagher swamp, 1,420 feet elevation, some four miles east of Java Village. And the Tonawanda waters flooded the valley of Cayuga Creek through a pass one mile northeast of North Java Station. Thus it appears that one water body occupied portions of four distinct stream systems, the Buffalo, Hunter, Tonawanda and Cayuga; to which perhaps we may add the Cattaraugus as a fifth drainage area. The relations geographic are shown on plate 29.

Apparently the only ultimate escape for the Buffalo Valley and its confluent waters was by the Chaffee pass until the ice front had receded eleven miles so as to uncover the Hunter Valley outlet, one mile southwest of Colegrove, the outlet of the Colegrove Lake, at 1,300 feet. With the fall of the Colegrove Lake (Hunter Valley) the Buffalo Valley waters became tributary to the Hunter Valley, Colegrove Lake, by northward outflow through the pass two miles north of Dutchtown, described above as the strait which had been the connection of the earlier waters of the Hunter and Buffalo valleys.

The northward outflow of the Wales Hollow Lake persisted until the two lakes blended together south of Sales Center (plate 29). The final flow was by a network of channels northeast and north of East Aurora, from 1,000 down to 900 feet. This latest flow appears to have been forced by the ice front around to the Holland Lake, just before their lowering into Lake Whittlesey. The features lie on the Depew sheet, and are mapped on plate 5, paper 10, and plate 29 of this writing.

In Valley of Cayuga Creek

This is the most easterly of the larger valleys tributary to Lake Erie. The slender creek heads in a swamp with elevation 1,376 feet, one mile north of North Java Station on the Arcade & Attica Railroad; and very close to the west fork of the Tonawanda Creek. The glacial history is tied in with that of the Tonawanda, the waters of which flooded the valley through the swamp noted above. The early waters, also the earliest glacial water of the Tonawanda, had elevation of 1,420 feet.

The first lateral outflow of the combined waters was one mile southeast of Bennington Corners (Attica sheet), at what is now 1,400 feet. A remarkable network of channels carried the later outflow, covering seven miles northwest. These are mapped on plate 5, paper 10, and partially on plate 29.

Johnsonburg-Varysburg-Attica Lake

Tonawanda Valley

Here is the most easterly of the parallel valleys with the streams having escape to the Erie basin. Today it contributes to the Niagara River. The next important valley on the east, the Oatka, contributes to the Genesee River and Lake Ontario, and belongs with the central New York series, described in papers 11 and 17.

The Tonawanda Creek has two sprawling forks in the towns of Java and Weathersfield, shown on the Arcade sheet. The west fork lies close to the Java Lake branch of Cattaraugus Creek. It runs only one half mile south of the head of Cayuga Creek, as noted above. The single deep valley begins two miles south of Johnsonburg.

The primitive glacial lake appears to have had outlet through Gallagher swamp and a pass one mile southwest, with elevation 1,420 feet. This led to the Beaver Meadow Creek, a fork of the Buffalo Creek. This level carried 44 feet over the head of Cayuga Creek, described above, and consequently flooded the Cayuga Valley.

The first westward outflow of the Tonawanda glacial waters was on the northwest-facing hill slope, less than two miles south of Bennington Village. Allowing for the tilting land uplift the first flow between the ice front and the sloping ground was at elevation over 1,400 feet. The series of successive channels below about 1,380 feet are mapped on plate 5, paper 10.

That flow past the ice front, south of Bennington, acquired drift from the ice and constructed a very interesting set of deltas, along the east slope of French Brook. Their elevation of 1,240 feet determines the elevation of the water surface, at that time, in the Cayuga Valley; under control of the outlets near Bennington Corners.

The next lower escape of the Tonawanda Valley waters was by the Konawaugus Valley, Gillett Creek and French Brook, passing close to the corners at Bennington.

Eventually a lower outlet was found four miles north, at East Bennington, at elevation 1,300 and down to 1,240. And two miles yet further north a low pass northwest of Attica, at 1,080 feet, which is utilized by the Erie Railroad. Yet later escape was by channels two miles north of Alexander, at 1,040 and down to 940 feet.

All of this northern territory is covered with moraine, and the Attica glacial waters found a plexus of low channels which carried the flow westward into Lake Warren at Alden and Crittenden.

The mapping of the glacial stream channels on plate 29 is not intended to be complete, only some of the principal channels being indicated. A keen observer, with a trained eye, may find many other minor scourways; and perhaps some errors in this mapping. The topographic sheets are the "guide, counselor and friend" in the field study.

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Appendix B

Gephart-Ripstein, A. (1990). "The Earth Moves," *Historical Wyoming*, January 1990.

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THE EARTH MOVES

by Anita Gephart-Ripstein

During our school years, we studied in some form or another, earth science. As I recall it was very boring studying the action of the earth's continental plates, magnetic fields, inner core temperatures, and all the allied information classified as earth science. But as the years progressed and earthquakes and landslides were witnessed, this boring subject took on a new interest.

Since early times, men have recorded the occurrences of earthquakes and landslides. The records of some of these are appalling in the number of casualties and loss of property. We have read about the great earthquake and the eruption of Mount Vesuvius in the year 79 A.D. which totally destroyed the cities of Pompeii and Herculaneum. And we are constantly reminded of the Great San Francisco Earthquake of 1906.

Though most major earthquakes and landslides have occurred far from the boundaries of Wyoming County, the county has experienced these phenomenons of earth science. The history of seismic activity in Wyoming County has, generally speaking, never been recorded. Little can be learned as to whether the area experienced such activity in pioneer times. No sources exist that refer to such occurrences during this time frame. It is probably safe to say that the large earthquakes in southeastern Missouri in 1811 and 1812 were felt in Wyoming County, as the shocks were felt as far away as Boston. Also it is probably safe to say the Attica area did experience seismic activity due to the frequency of known quakes in this area.

There are three major fault areas in Western New York. The most active and the most dangerous is the Attica Fault which stretches from south of Batavia to Attica. Modern scientists feel there is a strong possibility of a major quake occurring along this fault. Though the probability is low due to the fact that the area is geologically stable.

The second is the Clarendon-Linden Fault, slightly east of Batavia and very close to Wyoming County. It travels over the greatest distance in Western New York and it is one that the scientists know more about. Also it is not as active as the Attica Fault.

The third fault is believed to run through the Niagara Falls area. This fault is also not very active.

A fact that points to seismic activity in pioneer times is that seismologists know today approximately 150 to 200 earthquakes per year are recorded in New York State. The majority of these go unnoticed except to the scientists who study such occurrences.

October 23, 1857

On this date, an earthquake shook Warsaw, the Tonawanda Valley and Arcade areas. The **Western New Yorker** carried the following news item: "**WAS IT AN EARTHQUAKE?** - Something shook us up in this region on Friday afternoon, between three and four o'clock, which has been very generally attributed to an earthquake. The movement was more perceptible in brick and stone buildings than in wood structures, which gives color to the supposition that the disturbance was caused by movement beneath the earth's surfaces. There wasn't much of a 'panic' in consequence of the trouble, and no 'run' was made upon the bank."

Searching for futher evidence of this earthquake in the area we discovered the report of Dr. Charles E. West, entitled "On An Earthquake in Western New York October 23, 1857". Dr. West presented the report at a meeting of the American Association for the Advancement of Science, held in Baltimore during May 1858. It says in part:

"It occurred in Buffalo, at a quarter past three o'clock, p.m., and was violent compared to other earthquakes in the

Northern States. I was seated in a chair with my head leaning against the mantel of the fireplace when the shock occurred, and so great was its violence as to throw me forward to my feet." "A farmer living in Aurora, a town sisteen miles southeast of Buffalo, was digging potatoes in his field, at the time of the earthquake, and so powerful was the chock that he instinctively leaned upon his hoe-handle, and while in this posture he observed the dirt shake back and forth over in hoe, which was partially buried in the soil."

The **Western New Yorker** on November 3, 1857 published a news article from a Buffalo newspaper.

"Then again people differ widely as to the time which elapsed during the convulsion. It is no uncommon thing to hear five minutes insisted upon as the duration of the shock. The better informed, however, or those most competent to judge, are of the opinion that the vibration did not continue to exceed one minute. The first experience of the shock was by far the hardest portion of it, after which it gradually subsided like a series of waves. A little reflection will convince anyone that had the vibration lasted five minutes as it began, but few of our strongest walls would have been able to sustain themselves."

"We have already stated that the direction of the vibration was from the southward and westward - proceeding towards the northeast, in a series of three waves. We think the wave-motion abundantly established by the sinking, dizzy sensation related by everyone in his or her experience. The effect was precisely similiar to that experienced on shipboard, when the vessel descends into, and rises from the trough of the sea."

"We have been frequently asked since the affair to give some rational explanation of the cause of this phenomenal, but of course we can only approximate to such a result. The real cause of earthquakes is still unknown or unsatisfactorily demonstrated by men of science. Many content that they proceed from electrical disturbances; and there is one savant in Philadelphia we believe, the editor of some scientific journal, who proceeds to predict six months ahead their occurrence and thus far we must acknowledge with considerable success. Their appearance is always accompanied by some marked peculiarity of the atmosphere, either of excessive dryness, or, as was the case in this instance during the prevalence of a lowering mist which seems by its overshadowings to betoken some unusual occurrence. In all cases the atmosphere is exceedingly calm - we never heard of any occurring in the midst of a tempest. The prevailing opinion among scientific men who have made these occurrences the subject of their investigation, is that they are caused by the accumulation of air in the cavities of the earth, or by the generation of gases in connection with certain mineral substances sufficient in their escape to more large portions of the earth's surface. It is not an unusual thing for them to be accompanied by the discharge of water and smoke or flames, thus affording strong probabilities of the correctness of the latter theory."

According to current articles on local earthquakes, the quake of 1857 would have measured 4.9 on the Richter scale and affected an area of 18,000 square miles. It is the only earthquake during the 19th Century that documentation can be found in Wyoming County. These sources also give insight into the minds of science during these years and their reasoning as to the cause of earthquakes.

Numerous other quakes occurred during the 1800's, but local newspapers did not record these events. The dates of these shocks were: 15 Janaury 1858 with the epicenter near Niagara Falls, Ontario, Canada; 6 July 1873 measuring 4.9 which affected 30,000 square miles; 8 January 1876 near Lockport, New York; and 21 August 1879 northwest of Buffalo which measured 4.3. It is also very likely local quakes did occur in the Attica area but Attica newspapers of this period do not exist.

During mid September, this author was interviewed on the subject of Attica earthquakes by a news reporter for a local Rochester television station. The first question that was presented by the reporter was - "Do the people of Attica fear a great quake such as the 1906 San Francisco quake or similiar quakes of high magnitude?" Generally speaking I said no. The fact is that Attica lies a great distance from San Francisco and that Attica has really never experienced such a quake. With so little known about the Attica Fault, people do not fear a destructive quake but take very little notice of the quakes that do occur in this area.

At the time that the San Francisco quake of 1906 did occur, people were very much concerned but as time passed the concern was pushed back into their memories. If such a quake in San Francisco does occur again I am very sure people will think twice the next time they feel a quake in the Attica area.

Even though the San Francisco earthquake of 1906 did not directly affect the homes of Wyoming County people, it did indirectly affect their lives. Many Wyoming County natives and relatives lived in the San Francisco area and it is from their letters we gain insight into this great phemomenon.

Letters Home

A letter written April 25, 1906 by Miss Affia Martin was received by a friend in Warsaw. She writes: "I am safe from the earthquake and fire, but have had many strange experiences since Wednesday the 18th of April, when I was awakened by that terrible shaking. I shall never forget the sensation and the bricks and glass all tumbling down. We remained in our house until the next afternoon, when the fire was so near we had to leave for the hills and were out two nights and saw most of the city burn down. It was an awful sight. And - Oh dear! just now another shake came, the hardest we have had since last week and it makes me feel sick and discouraged, for every one has thought we would not have any more hard ones."

"This was not as bad as the first, but rocked the house a good deal, and I hardly know what I wanted to say."

"We are in rooms now, but not allowed to have any fire in the house; everyone has to cook in the street, and we have had no lights at night until Monday. We are allowed a candle until 10, but they are scarce. The uptown stores that were saved are nearly empty."

"I am trying to be very thankful that I have been spared the dreadful suffering and trouble so many have had. I don't thinl the papers can ever picture it as bad as it is; so you may believe what you read. The first dreadful sight, was a four story hotel near us, that dropped three stories into the ground. It was built on "made ground" and an old creek was under it. I think most of the people in it were either killed or maimed; but that is just one instance. I am sorry that last shake came just now, for it sent everything from me and I feel weak, so I cannot tell you much. Tell all my friends that I am alive and give them my love."

The following is part of a letter dated April 23, 906 and was received by Mrs. Henry Austin of Perry Center from her daughter in San Jose, California.

"The first anvone knew of it, we were all swinging side to side. Some were thrown out of bed. I jumped out, but it threw me flat on the floor, then I crawled under the bed so if the plaster fell it would not strike me. One chimney is near the bed, but the bricks fell out and onto the ground instead of through the roof, as many did. It lasted 27 seconds and seemed an age to me. After swaying back and forth it wound up like taking them by the shoulders and giving them a vigorous, fast shaking. Light shocks last 24 hours, coming about every two to three hours, kept us wrought up to a pitch that is hard to shake off. Tongue and pen can never describe the feeling."

"After it ceased I crawled out and tried to open the door, but could not. Finally it yielded and I flew downstairs and out. All the neighbors were fleeing. When I came to my senses I found I was carrying a hand mirror, was barefooted and in my night clothes."

On April 24, 1906, Mrs. Julia L. Matteson of Varysburg, who was visiting a son in San Francisco, wrote a letter to her son George M. Davis of Rock Glen.

"I sent you a telegram soon after the earthquake here but as they would not take pay here for same am not sure you received it. It is wonderful how we escaped such a shaking up and down. I hope never to experience again. Our house moved several feet, fell about six feet and was crushed. Somehow we all escaped with no serious injury. Fred built a tent on a vacant lot across from his wrecked home, covering it with carpets, and we all lived in that tent until last night. Yesterday it commenced raining and is raining still. Fred got me a room nearby and I undressed and went to bed for the first time since the earthquake, six days ago. Soon after retiring we had another shock, quite severe one too. We have several light ones since the first. Since the rain set in, Fred and his family have found sleeping room in a barn nearby. He moved a davenport into the barn and he and Mable and daughter, Katherine, slept on that. In another stall covered with nice straw and a mattress on that, slept Theron and Charles. Mabel has a brother and his sons here, they slept in the tent. He acts as our cook, has a small tent with a stove in it. We have plenty to eat. The Relief Corp supply all we need. The rich and poor share alike in rations."

"I am coming home as soon as Fred can get me a lower berth. So many are going away, the cars are all crowded. The railroads will carry you out of the state free, and for one cent a mile to any Eastern point."

Mrs. Matteson did return shortly after to Varysburg and until her death in 1926 she always feared the shaking of the earth.

1914

The area around Three Rivers in Quebec, Canada has been the epicenter for many severe earthquakes which have affected New York State and Wyoming County. One such quake occurred on February 10, 1914. The disturbance was felt throughout state and Warsaw reported it was quite pronounced. Shortly after 1:30 P.M. that Tuesday afternoon, the buildings in Attica and Warsaw areas began to quiver. Lasting 15 to 30 seconds, this earthquake did no damage in Wyoming County.

For a few years the earth settled back down and earthquakes were the farthest thing from anyone's mind. But on the night of February 28, 1925, this was to change.

EARTHQUAKE SWEEPS THE COUNTRY

On the Saturday evening of February 28, 1925 at 9:23 P.M., the whole eastern part of the continent shook. Scientists claimed it was the most severe earthquake this part of the country had experienced in a hundred years. From Detroit to the Atlantic, from Virginia to Canada, business blocks and houses rocked violently. The epicenter was near the village of Tadousac, Quebec, Canada, some seventy miles north of mouth of the St. Lawrence River and about 650 miles from Western New York. This same area had produced the quake of 1914 and earlier ones recorded in colonial times before Wyoming County was settled.

The first seismograph report of the quake came from Canisus College which houses the second oldest seismograph in the country. The instrument recorded seismic activity for a period of fifteen minutes. The tremors began at 9:21 P.M. and continued until 10:11 P.M. They reached the maximum at 9:24. At 10:50, the seismograph was still vibrating although it was only noticeable through a microscope.

Though the epicenter was far away and not much damage was reported in Wyoming County, many people were considerably frightened by the tremors. The newspapers that week were filled with accounts of the earthquake.

The **Sheldon Democrat**, published in Varysburg, gave the local account.

The earthquake of Saturday evening was plainly felt throughout the village. In all the homes the earth's tremor was evinced by chairs rocking and sliding about, heavy furniture moving, chandeliers swaying, clocks stopping and crockery sliding from shelves. The disturbed fowls in neighboring heneries squacked as they flew from their roosts, raising a general alarm as the loose panes of glass in the windows of their shelters were either thrown to the ground or splintered. Panes of glass more strongly set in barn windows were cracked as Old Mother earth heaved, grumbled and apparently turned over in her sleep. The earth movements were from east to west and the reverse. The vibration seem to have followed the Tonawanda Valley north to south while hill dwellers east and west felt little of the shock. Two miles north down the valley, at the home of Miel P. Caklins, the family had retired. The wire bed springs begin to vibrate and a sound as of the gentle opening of the front door, which was securely locked, was heard. At the home of John Ward on the Creek Road, the clock stopped as did those in the homes of Erwin Welker and Reinhardt Roth. At the Watson farm, the home of Mr. and Mrs. U.G. Calkins, the head of the house had retired. Feeling the bed rise beneath him, he clambered out while Mrs. Calkins sitting by the sitting room table, reading a magazine, believed herself suffering from a dizzy attack to which she is subject as the easy chair in which she was sitting moved sidewise. Her husband's empty chair rocked and the cloth table cover swayed. The bristling backs of her pet Persian kittens and the howling of the farm dog decided her otherwise. She glanced at the clock on the wall. It was still and registered 9:20...."

Warsaw's **Western New Yorker** commented: "It is very interesting to hear what people were doing at that hour and the peculiar and often funny things which happened. At Millman's store a shelf of canned corn fell into the middle of the floor and an empty orange crate was jarred loose and went tumbling down the cellar stairs. At the Congregational Church, the quartette was practicing. The organ began to weave back and forth and the floor to move in an opposite direction. It seemed as if the hallows of the organ would burst...Some people thought perhaps the Long Island people who recently expected the world to come to an end were only a little early in their predictions. At one house a door bell which had been disconnected for some time started to ring."

"The impulse of most everyone was to get outdoors. In no time the middle of Main Street was filled with people from stores, houses, hose and pool rooms, and offices. Most of them were looking up expecting to see the Main Block, the Crossett Block or some other of the higher buildings tumble down. The tremors continued for two minutes."

Warsaw, Varysburg, Attica, South Warsaw and Silver Springs felt the earthquake of 1925. But it was just the beginning of things to come.

The Greatest Earthquake

The most destructive earthquake ever recorded locally occurred on the morning of August 12, 1929. The first warning came at approximately 6:25 A.M. with a sound similiar to heavy distant thunder, immediately followed by a severe shaking which lasted for several seconds. This first shaking was followed by milder quakes lasting about one minute. A second jarring was felt by some people and was later confirmed by scientists.

Though this quake was felt throughout the state of New York and in areas in Ohio, Pennsylvania, Massachusetts and Ontario, Canada, the most damage occurred in the village of Attica. The intensity of the quake has been estimated to be 5.5 on the Richter scale. The seismograph at Canisus College indicated that the initial shock lasted for about 12 seconds with minor tremors being recorded thereafter for six minutes.

The most damage recorded in the Village of Attica took place east of the Tonawanda Creek. At least 200 chimneys in the village were demolished in this area alone. Buildings particularly along Main Street, Genesee Street, East Avenue and North View Park suffered the most destruction.

Searching local newspaper for damage accounts the following was learned. The Charles Benedict house on Main Street suffered extensive damage to its walls. This house today is 110 Main Street and was originally built in 1835 as Ingham University. The brick structure suffered a wide crack in the brick facing Main Street and numerous smaller cracks throughout the residence. The exterior crack it was said was wide enough that the interior of the house could be viewed. The four foot high balustrade that had been added to the roof in 1900 was also toppled or cracked. It had to be removed.

Another brick residence on Main Street that suffered extensive damage was the Eggleston home. This house was constructed with a brick cupola which was badly shaken and had to be remove. The large chimney on the Smith residence next to the Presbyterian Church broke through the newly shingled roof and landed in the bedrooms of this stately home.

The Charles Newell residence, one mile from the village on East Main Street, was shifted several inches off its foundation. The walls of the brick home of O.F. Bowman, one-half mile from the village on the Creek Road collapsed as did the walls of August Merle's stone house on Exchange Street. Both families escaped injury in these two homes. Roofs throughout the area were badly damaged as heavy chimneys fell in.

Other damage was reported in the papes. In the village stores, goods were thrown from shelves into piles which took several hours to pick up. The jar exploded a boiler at the milk plant on Exchange Street; several electric lines were broken and plaster was stripped from walls. The exterior of the Methodist Church on Main Street at the foot of Exchange Street suffered many cracks with numerous bricks toppled from their places. The church pipe organ was also damaged. The altar of the brick Presbyterian Church shaken from its foundation.

Perhaps the damage affecting the largest number of people was that doen to the new brick school on Prospect Street. The third floor especially was damaged by falling plaster by the bulging brick walls. The large chimney and coping were broken loose. Upon inspection by the firm of Morris & Allen, the builders of the school, the foundation was found to be stable. The chimney was rebuilt and the cracks filled in.

Newspapers also contained reports that the earth opened up near the homes of John Miller and Fred Baker, east of the village, giving forth sulphur springs. The rock walls of several wells loosened soiling the water and in cases cutting off entirely the supply of water. The small brook near Georges Drive on East Main Street went almost dry that morning but before nightfall was flowing again.

Westinghouse Electric & Manufacturing Plant was forced after the earthquake to shut down their stoker for several days. Upon inspection it was found that some of the machinery was wrenched out of place and the cupolas used for melting iron were damaged. It wasa found that they were so badly twisted out of place it took a number of days to make repairs. The concrete roof of the plant was badly cracked and three sections had to be totally replaced.

The quake not only damaged buildings but damage was noted in the area cemeteries. At Forest Hill Cemetery which is located just a few feet west of the creek, many of the tall marble monuments were moved from their foundations anywhere from one-half inch to three inches. Small ornamental objects on the monuments such as balls, vases, etc. were thrown to the ground. At Brainard Cemtery, east of the village, almost every monument was tipped over or moved from its foundation.

Many things happened which as time went on after the quake were very humorous and show how little people care for dignity in time of danger. All kinds of people in nightclothes and hair curlers were seen on lawns, porches, and hanging from windows wanting to know what was happening. Probably never before or since has there been such a general uprising in the village at 6:25 A.M. Dogs barked, cats yowled, and cows kicked over milk pails.

Granted the quake of 1929 was nothing compared to the San Francisco quake of 1906 but to the people of Attica the fear on experiencing that first shaking sensation was very much the same.

Soon after the earth had settled back down, the area was being visited by geologists. The source they believed for the quake of 1929 was activity along the Clarendon-Linden Fault. The Attica Fault was not known at this time and in later years the source would be placed along this Attica Fault.

On December 2nd and 3rd of that same year, Attica residents were again shaken by earth tremors but no damage was done.

Mystified Scientists

At 7:30PM Thursday February 23, 1939, cell doors in the Attica State Prison began to rattle throughout the whole institution. People attending a play in the Attica High School auditorium heard a sound like heavy thunder. At Crystal's Pharmacy, Dr. John Sturrock thought an explosion had occurred to the east. Many thought at first it was county employees blasting ice in a stream near Attica but no men were working. It was soon realized that the Village of Attica once again was experiencing an earthquake.

The big difference with this tremor was it failed to register on the seismograph and it was only felt within a ten mile radius of the village. Noted scientists could not explain this earthquake as it seemed to be centered in the Attica area. No known geological formation in Western New York could cause such a localized tremor. But they all agreed it was an earthquake. Professor Reginald H. Pegrum, University of Buffalo, theorized that it might had had its origin in the collapse of an abandoned mine or some underground cavern in the limestone formation. But as far as he knew no such mine or cavern existed. The scientists were mystified as to the origin of this 1939 earthquake.

Through the years Attica has experienced many of these "small" quivers and local residents seldom think twice about them. During the late 1930's and early 1940's they were very common. Some made the papers while others were only talked about in passing conversation. But the one thing that all of the tremors had in common were they were felt more strongly east of the Tonawanda Creek. Also there was always the sound of an explosion.

During 1942, there were two mild quakes - 2nd of February and 28th of May. No damage occurred only the rattling of dishes and the "earthquake sounds". But in the year 1944, the earth seems like it was constantly rumbling. Scientist still could not explain why the Attica area was quivering.

Shortly before 1 o'clock on Tuesday morning, September 5, 1944, area residents were experiencing a severe thunderstorm. Following a heavy clap of thunder the earth began to shake. The citizens of the Attica area immediately recognized the sensation as an earthquake and not just the vibration of heavy thunder. On this morning in the city of Cornwall, Canada, north of the St. Lawrence River a major earthquake had occurred. This was the tremor the people felt.

The next decades were not any different in the Attica area. The earth was still shaking and for generations had become a way of life. But with each new tremor, pieces of a puzzle were being fitted together to answer the question - Why does Attica have so many tremors?

It would be discovered over the years that the village of Attica is located directly over a major fault. A very active fault. The location of the fault lies one mile east and parallel to the Tonawanda Creek Valley and ten miles below ground.

The Attica Fault is still not completely understood. But with each new generation of scientist, new studies are being made. Maybe someday we will know the answer to the question - Why? Meanwhile the people of Attica and Wyoming County go about their lives like they have for generations while the earth beneath them moves.

Dates of Earthquakes

The following is a list of earthquake dates that have been reported in local newspapers as being felt in the county.

25 November 1988

3 March 1986

31 January 1986

7 October 1983

8 February 1973

12 August 1969

1 January 1966

26 August 1965

17 July 1965

16 August 1955

20 September 1946

5 September 1944

2 March 1944

26 February 1944

28 May 1942

12 February 1942

31 October 1939

23 February 1939

17 Janaury 1930

3 December 1929

2 December 1929

12 August 1929

28 February 1925

10 February 1914

(Some dates may be missing because some milder quakes may have not made local newspapers. Also 1800's dates not included due to lack of source material.)

The Great Slide

The first recording of a major landslide in the county of Wyoming is mentioned in Dr. James E. Seaver's book on Mary Jemison, the White Woman of the Genesee. Mary speaks of the Great Slide but does not tell details. It is from the account told by Charles Strong to William Pryor Letchworth in 1871 we learn of this great phenomenon.

It was on the night of June 18, 1817. Some men came down the river with a raft of logs and, leaving the raft in the eddy, where they thought it to be safe, went to Mary Jemison's to spend the night. About ten o'clock they heard a terrible rumble, which they were unable to account for. They found later that seven or eight acres of the high bank, two hundred feet high with trees growing on it, had broken off and slid into the Genesee River. The broken surface of the hill and the debris of the slide were spread out in irregular hillocks in the river, covering fifteen or twenty acres. This turned the stream across the flats above the slide and around to the east bank. The raft of logs had disappeared forever. There was a large tree, Mr. Strong said, twelve feet in circumference, which was not uprooted but took enough earth with it so that it kept right on growing in its new location. In later years, when the land was plowed over, he found stems of trees forty feet long embedded in the soil. Strong said the slide was sixty rods from the cabin of Polly Jemison.

The next morning, says another source, Greenleaf Clark and his son, Edward, and

Samuel True, with his son Ira, came from Perry to hoe corn on Mary Jemison's flats, which they had rented. In after years, Ira stated that he had run ahead of the others and jumped on the great mass of bluish clay exposed by the slide. He said the air smelled so strong of sulphur, they could hardly breathe.

Since the Great Slide, there has been many landslides in what is today Letchworth State Park. Yet the Great Slide has always fascinated people. The late Woody Kelly, Town of Perry Supervisor and Chairman of the Wyoming County Board of Supervisors once told the following story: "I recall my Dad's uncle, Frank Phillips, of this community telling an inquisitive person about that happening. It seems being born in 1850, he had aged considerably by the 1930's and this person was trying to determine his age. In his query, he asked if Mr. Phillips remembered the great landslide on the Genesee River. He assured him that he did and added that he remembered that as a result of the slide, the woodchuck holes stuck out from the bank after the dirt slid away and that people came from miles around this locality with horse and wagon rigs to go to the spot and saw the protruding holes off and take them home to be used as stovepipes. This of course put an end to the question and answer session."

Though Mr. Phillips was much too young to remember the Great Slide in the days of Mary Jemison, he probably heard of the slide that occurred one year prior to his birth. On Sunday, October 13, 1849, an extensive landslide occurred on the line of the Genesee Valley Canal, near Portageville, by which ten thousand yards of canal embankment was carried into the Genesee River. It was on the section above what was termed the "cliff line". The canal at that point ran just above the brow of the river's bank, which was about one hundred feet in height, and though not perpendicular, was very steep. At this point a heavy embankment had been constructed on the river side of the canal. Deep underneath this embankment was a large bed of quicksand. The combination of the weight of the embankment and the presence of quicksand contributed to this landslide of 1849. In the rush of the landslide large trees were carried downward, some of which remained erect and continued to grow in the years that followed. The damage to the canal was very considerable as it took
hun
hundreds of hours of manual labor to replace the embankment along with twenty to thirty thousand yards of soil to repair the canal.

In the Tonawanda Valley, on the opposite side of the county, numerous unexplained landslides have occurred. One entered into a legal action in the court of claims when the New York State Highway Department was defendant in a suit for damages brought by the Arcade & Attica Railroad Corporation, seventy-seven years after the date of the landslide. The chief witness in the case was only seven years old at the time of the phenomenon.

A Day in May

On a bright spring morning, young Miel Pierce Calkins grabbed his fishing pole and headed for his favorite fishing pole. Sitting beneath the old wooden bridge that once spanned the Tonawanda Creek on the present day Eck Road, the seven year old boy passed the morning away only as a small boy knows how to. With blue skies overhead, the morning was filled with subdued sounds of life along a rural creek. When without warning there fell upon this quiet scene a sudden, startling, loud hissing sound. Young Calkins turned quickly only to witness the creek bed shoot up into the air twelve feet. Clamoring up the slippery creek bank, he ran towards home and safety.

Miel lived but a short distance from the creek as his father, Belus Calkins, worked the farm on the south side of the dirt road where it met the main highway between Attica and Varysburg. On the south side was the farm of Valentine Welker.

It was on this Thursday the 9th of May 1861, that twenty acres of farm land west of the Tonawanda Creek slid forward one hundred feet and in doing so lifted the creek bed. Turned from its natural course, the water spread over the farm land in every direction. As the alarm was sounded, neighboring farmers soon arrived at the scene with their farm teams and oxen and quickly a channel was plowed to divert the turbulent water back to a single waterway.

The following Sunday was turned into a Roman Holiday by the curious public. The dirt roads around Calkins Corners were lined for miles by the wagons and buggies of the sightseers.

Belus Calkins, who was also known as Captain, was a portly gentleman.And on this Sunday afternoon while pointing out the site of the upheaval in the creek bed, he stepped on what he thought was a substantial block of earth. Feeling the ground begin to revolve beneath his feet, he turned to run, as he did so his tall beaver hat flew from his head. It fell into a fissure and was never seen again.

As to the law suit in which Miel P. Calkins appeared as the only living witness to this happening, the out come is unknown.

During the years gone by, the Tonawanda Creek Valley produced many strange happening. Many were never explained by scientists. Years previous to the 1861 event, a landslide involving several acres of cleared land occurred about a quarter mile further north on the east bank of the Tonawanda. Just as a four-horse stage coach passed carrying the U.S. Mail and filled inside and out with passengers, the ground slid westward across the highway carrying all before it.

On a summer's day, Waldo Munger was cultivating a field of sweet corn a short distance from the previous mentioned slide. Next morning he arose early with the intent to finish his task but when he arrived at the open field, he found in the center a large round pool of dark bubbling water. Upon investigation, its bottom was never sounded. Throughout the years, frequently large logs and forest refuse were thrown into it and it never was filled. The pool at first was nineteen feet across. It increased yearly in size.

The most remarkable landslide along the Tonawanda Creek occurred on March 4, 1920, according to one source, but documentated to have occurred on February 27, 1920. The location was approximately in the same area as those recalled by the old-timers and was on the farm of Mrs. Edward Welker on the southern part of lot 11, Section 1 in the Town of Bennington. The farm house was occupied by tenants, Grover and Kate Wolf, who did not hear any unusual noise during the early morning hours when the slide occurred.

George S. Libby, the nineteen year old son of William and Adell Libby, of Varysburg was the first to discover the slide. Before sun rise, he had left his home in Varysburg by horse and cutter to travel to the Village of Attica on business. As he approached farm of Mrs. Welker, his horse came to an abrupt halt and refused to go on. Disembarking from the cutter, he went to investigate why the horse acted the way he did. He discovered in the half light and shifting snow that there was no way to go on as there was a "wall" on the highway. Turning his horse around, he headed back to Varysburg to tell of the happening. Stopping at the corner store of Frederick W. Embt, he learned that no one had come up the valley since around two o'clock that morning and everything was fine.

The following graphic account was published in the **Rochester Herald** after Professor George H. Chadwick, district representative of the New York Geological Department,visited the site:

"Even more spectacular and remarkable then was first pictured is the big landslide six miles south of Attica which at four o'clock last Thursday morning or there-abouts caused a thirty-five foot hill to disappear completely, leaving a formidable cliff where once the Attica-Varysburg highway, better known to Wyoming County folks as the Creek Road.

Instead of only a little more than an acre of ground, as was stated in early dispatches, measurements and computations made on the scene this morning show that more than six acres of land were dislodged and carried downward thirty feet or more. The area affected is 1000 feet long and approximately 500 feet in width and is located on a side hill at its lower extremity forms the east bank of the Tonawanda Creek.

The sunken road parallels the creek traversing the hill some 300 feet distant from the water's edge and at its highest point was sixty or more feet above the normal creek level. There is a twenty degree curve in the road at the point where the landslide took place and this is on a gradual down grade to the south.

Whereas, it was first stated that the sunken area included a section of the road about 500 feet long and the land 100 feet back from the highway on either side, the dimensions are much greater than this as the figures will show. Everything from a point about 150 feet east of the road to the creek bank which is fully 300 feet from the road on the west side, has slumped downward. That portion which lies nearest the stream, on the west side of the road, being at the foot of the side hill, has fallen a proportionate distance as compared with that on the east side of the road, which is on the hill proper. Here the cliff which has been formed by the depression, like that at the north end of the sunken road itself, is approximately thirty feet high its entire length.

This great mass of earth, in falling so great a distance exerted a tremendous pressure on the clay formation underneath. On three sides of the sunken area is hilly land, rising quite abruptly and it will readily be seen that there was but one, avenue of escape for the soft, water soaked clay when the surface of the earth gave way. This was towards the creek. And here a most unusual phenomenon.

The clay was squashed out, so to speak, underneath the creek bank, raising the bed of the stream between ten and fifteen feet above its former level for a distance of more than 500 feet forming a new plateau or new bank midway of the stream. Huge cakes of ice and frozen creek bed were forced upward in great upheavals which resembled the effects produced by an earthquake. These same jagged masses of snow, ice and earth maybe noted throughout the entire lower level of the sunken area, running criss cross and in every direction like great cracks or fissures in the frozen earth. In some places the earth seems to have buckled as if there had been an equal degree of resistance from opposite directions and in such instances jagged ridges and tent like formations have resulted."

It was estimated at the time of this landslide that over 1000 tons of solidly packed ice and snow covered the land which fell and this, together with the evidence of underground streams was believed to account for this strange occurrence. Over the seventy year period since this great slide on the Tonawanda Creek, the area has been relatively quiet but one never knows when the earth will move again. Nor does one know when the earth will vibrate along this creek of "rapid moving waters".

* *

At The Office

The Christmas Gift promotion for **Historical Wyoming** was very well accepted. Over 200 new gift subscriptions were received. We would like to thank all our readers in their help to promote our publication.

As noted in letters sent out during December of 1989, the price of **Historical Wyoming** will be increased in 1990 due to the increases in printing and mailing costs. The new subscription rate of $7.50 will go into effect when the new volume is published in July 1990. Also at this time a new look to the publication will also be introduced. So if you have not already done so - renew early!!

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Also many of our readers have asked whether back issues of the publication is available. We have copies of almost all issues available from the start of the publication in 1947. Some earlier issues are very rare but in some cases, they are still available or a photo-copy can be made for you. The price of back issues is $1.50 at the office or $1.75 if mailed.

Cowlesville Village Cemetery (continued)

YEOMANS

Richard 1819-1907
Caroline 1826-1908
Elizabeth, dau. d. June 25, 1865 15y10m 5d
Letitia, wife Charles, d. June 19, 1850 68y
C.Y. (no other data)
Mary Ann d. Nov. 4, 1898 76y8m6d (dau. Charles & Letitia)

ZIMMERMAN

Archie U. 1894-1945 Husband
Paula Krauss 1904-1953 Wife

The End

Appendix C

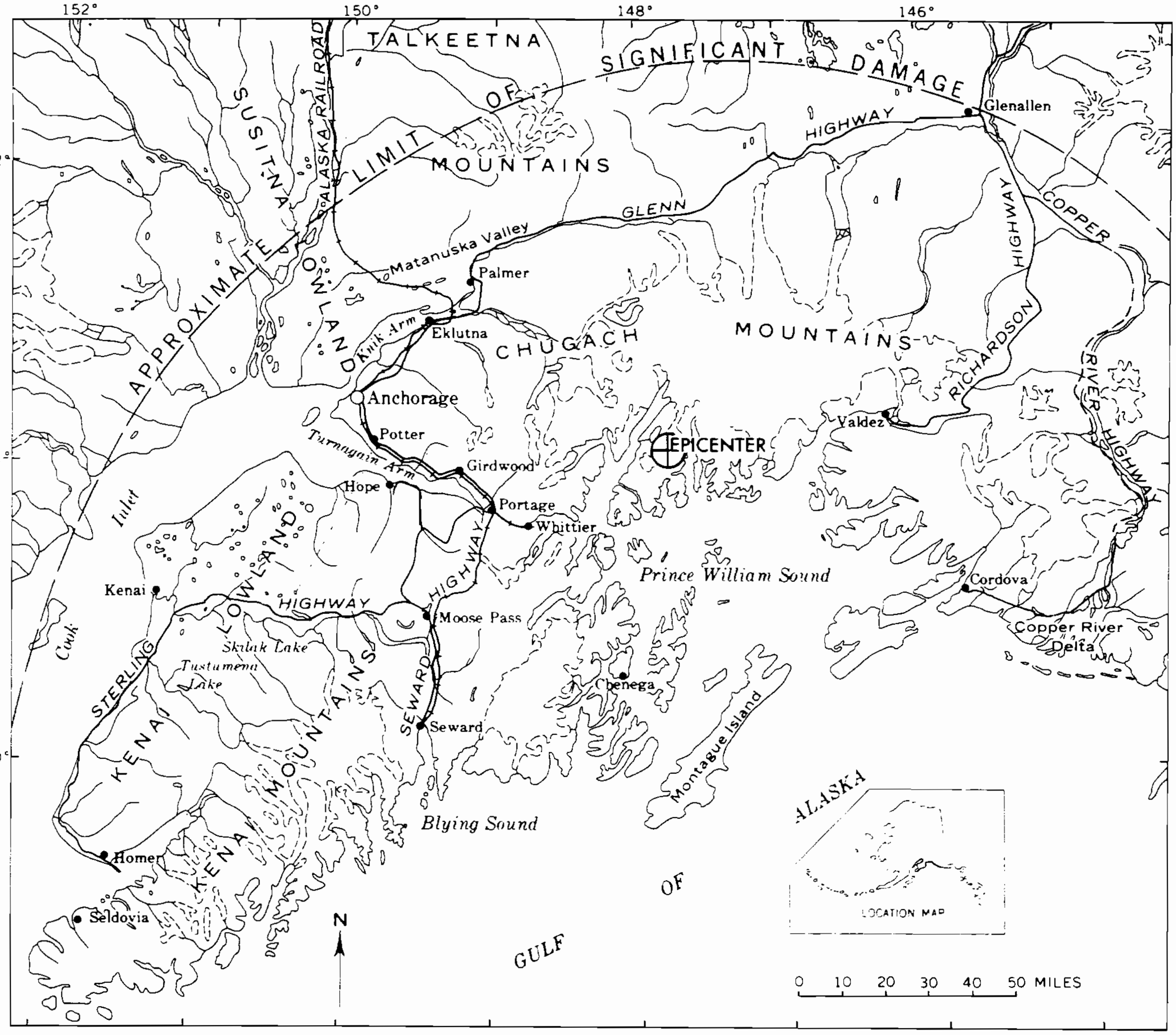
Hansen, W.R. (1971). "Effects at Anchorage," in *The Great Alaska Earthquake of 1964*, Washington, DC: National Academy of Sciences.

The Great Alaska Earthquake of 1964

COMMITTEE ON THE ALASKA EARTHQUAKE
OF THE
DIVISION OF EARTH SCIENCES
NATIONAL RESEARCH COUNCIL

GEOLOGY

NATIONAL ACADEMY OF SCIENCES
WASHINGTON, D.C.
1971



U.S. Geological Survey

FIGURE 1 Map of south central Alaska.

covered sessile organisms and salmon-spawning beds with silt, disturbed and killed salmon fry, leveled forests, and caused saltwater invasion of many coastal freshwater lakes.

The tectonic elevation and depression caused extensive damage to the biota of coastal forests, migratory-bird nesting grounds, salmon-spawning waters and gravels, as well as shellfish habitats, and initiated long-term changes in littoral and stream morphology. Clams, barnacles, algae, and many other marine and littoral organisms perished in areas of uplift. Spawning beds, trees, and other vegetation were destroyed in areas of depression.

Except for the major tsunami, which caused extensive damage in British Columbia and took 16 lives in Oregon and California, violence to man and his structures was restricted to the area of tectonic land-level change. Tsunamis, major and local, took the most lives. Landslides caused the most damage.

The number of lives lost in Alaska, 115, was very small for an earthquake of this magnitude. Factors that contributed to the light loss of life were the sparse population, the fortuitous timing of the earthquake, a low tide, the absence of fire in residential and business areas, the generally clement weather, and the fact that the earthquake occurred during the off-season for fishing. The earthquake came on the evening of a holiday, when the schools were empty and most offices deserted, but when most people were still

Contents

WALLACE R. HANSEN
U.S. GEOLOGICAL SURVEY

Reprinted with minor changes from
U.S. Geological Survey Professional Paper 542-A,
"Effects of the Earthquake of March 27, 1964, at Anchorage, Alaska"

Effects at Anchorage

ABSTRACT

Anchorage, Alaska's largest city, is about 80 miles west-northwest of the epicenter of the March 27 earthquake. Because of its size, Anchorage bore the brunt of property damage from the quake; it sustained greater losses than all the rest of Alaska combined. Damage was caused by direct seismic vibration, by ground cracks, and by landslides. Direct seismic vibration affected chiefly multistory buildings and buildings having large floor areas, probably because of the long period and large amplitude of the seismic waves reaching Anchorage. Most small buildings were spared. Ground cracks caused capricious damage throughout the Anchorage Lowland. Cracking was most prevalent near the heads or within landslides but was also widespread elsewhere. Landslides themselves caused the most devastating damage.

Triggering of landslides by the earthquake was related to the physical-engineering properties of the Bootlegger Cove Clay, a glacial estuarine-marine deposit that underlies much of the Anchorage area. The Bootlegger Cove Clay contains zones of low shear strength, high water content, and high sensitivity that failed under the vibratory stress of the earthquake. Shear strength in sensitive zones ranged from less than 0.2 tsf to about 0.5 tsf; sensitivity ranged from about 10 to more than 40. Sensitive zones generally are centered about 10 to 20 feet above sea level, between zones of stiff insensitive clay. Many physical tests by the U.S. Army Corps of Engineers were directed toward analyzing the causes of failure in the Bootlegger Cove Clay and finding possible remedies. Strengths and sensitivities were measured directly in the field by means of vane shear apparatus. Atterberg limits, natural water contents, triaxial shear, sensitivity, dynamic modulus, consolidation strength, and other properties were measured in the laboratory. Pulsating-load tests simulated earthquake loading.

Most of the destructive landslides in the Anchorage area moved primarily by translation rather than by rotation. Thus, all the highly damaging slides were of a single structural dynamic family despite wide variations in size, appearance, and complexity. They slid on nearly horizontal slip surfaces after loss of strength in the Bootlegger Cove Clay. Some failures are attributed to spontaneous liquefaction of sand layers. All translatory slides surmounted flat-topped bluffs bounded marginally by steep slopes facing lower ground. Destructive translatory slides occurred in the downtown area (Fourth Avenue slide and L Street slide), at Government Hill, and at Turnagain Heights. Less destructive slides occurred in many other places—mostly uninhabited or undeveloped areas.

In most translatory slides, damage was greatest in graben areas at the head and in pressure-ridge areas at the toe. Many buildings inside the perimeters of slide blocks were little damaged despite horizontal translations of several feet. The large Turnagain Heights slide, however, was characterized by a complete disintegration and drastic lowering of the prequake land surface. Extensive damage back from the slide, moreover, was caused by countless tension cracks.

An approximation of the depth of failure in the Bootlegger Cove Clay in the various slides may be obtained by using a geometric relationship herein called the "graben rule." Because the cross-sectional area of the graben at the head of the slide approximated the cross-sectional area of the space voided behind the slide block as the block moved outward, the depth of failure was equal to the area of the graben divided by the lateral displacement. This approximation supplements and accords with test data obtained from borings. The graben rule should apply to any translatory slide in which flowage of material from the zone of failure has not been excessive.

Geologic evidence indicates that landslides similar to those triggered by the March 27 earthquake have occurred in the Anchorage area at various times in the past.

INTRODUCTION

Anchorage, "metropolis of the north" and Alaska's largest city, is the rapidly expanding commercial center of the 49th State. Anchorage is about 80 miles west-northwest of the epicenter of the March 27 earthquake (fig. 1).

The Anchorage Lowland, a broad, undulatory glacial plain on which the greater Anchorage area is situated, is roughly triangular. It is bounded on the northwest by Knik Arm of Cook Inlet, on the southwest by Turnagain Arm, and on the east by the abrupt west front of the Chugach Mountains. Anchorage itself is crowded close to Knik Arm, but its perimeter is moving east and south because of suburban growth and soon, no doubt, will reach Turnagain Arm. Expansion north and northeast toward the Eagle River, along the narrow corridor between the mountains and Knik Arm, is largely checked by the military reservations of Fort Richardson and Elmendorf Air Force Base.

ACKNOWLEDGMENTS

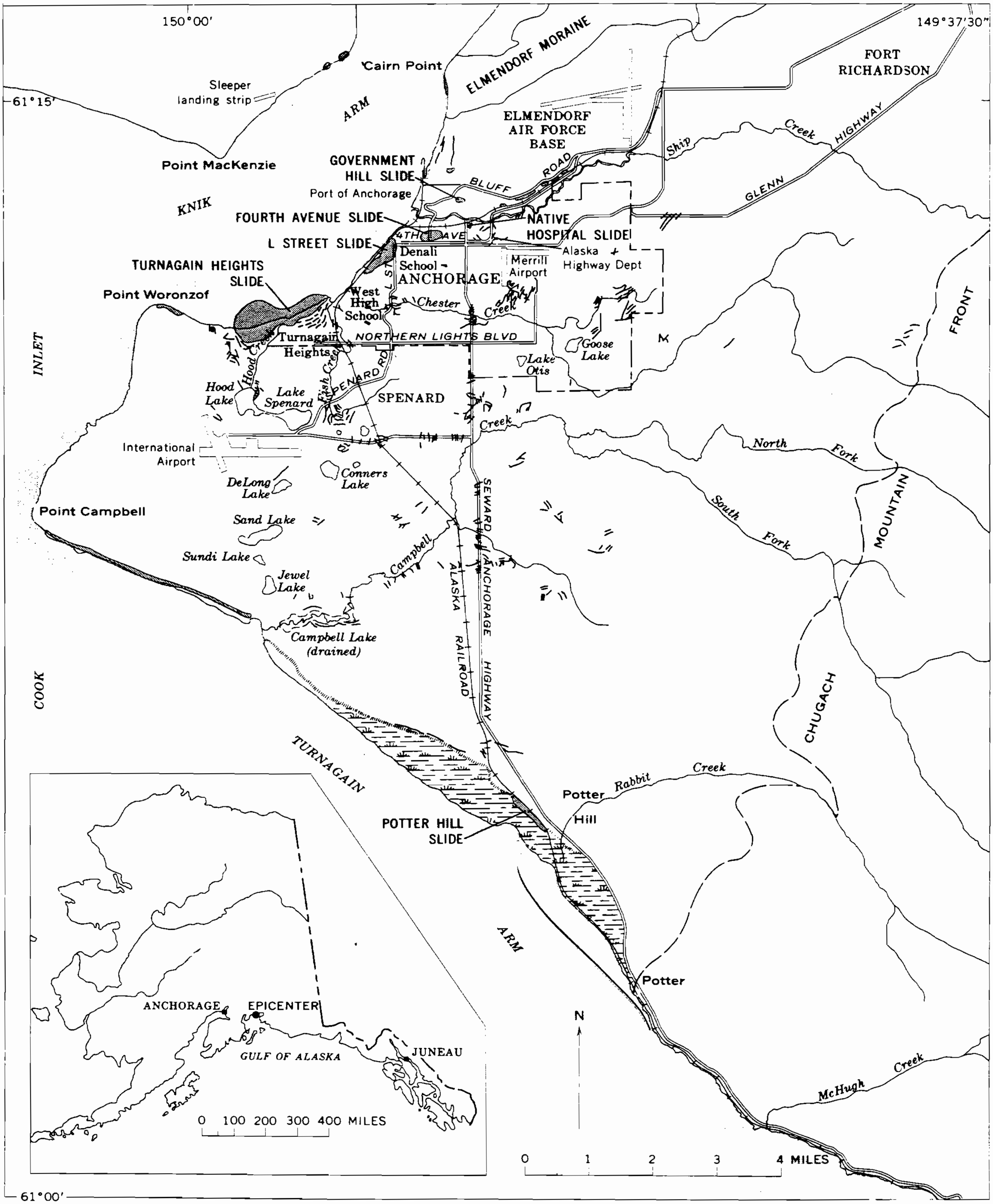
Many ideas and conclusions presented in this report were reached in concert with R. D. Miller and C. A. Kaye, both of whom shortly after the March 27 earthquake joined the author in field investigations at Anchorage. Most of the credit for devising the "graben rule" (see p. A41) must go to Kaye. Discussions with M. G. Bonilla, Ernest Dobrovolny, E. B. Eckel, Reuben Kachadoorian, D. S. McCulloch, and Roger M. Waller consciously or unconsciously led to thoughts incorporated into this report. J. R. Helm, A. B. Dodd, and Arthur Gervais of the Topographic Division, U.S. Geological Survey, ran control for profiles of the various landslides. Helm also prepared the topographic maps of the Native Hospital, Government Hill, and Turnagain Heights slides, using a Wild B-8 plotter. Special thanks are due the U.S. Army Corps of Engineers for the materials and information that they supplied. A report prepared for the Corps by Shannon and Wilson, Inc. (1964) has been freely utilized, and several of its illustrations are reproduced herein. The Engineering Geology Evaluation Group (see p. A10), an ad hoc organization that carried out early investigations at Anchorage at a critical time immediately after the earthquake, is also thanked for its free release of otherwise unobtainable information.

THE EARTHQUAKE AND ITS IMMEDIATE AFTERMATH

At 5:36 p.m. on Good Friday, March 27, 1964, Anchorage and all southern Alaska within a radius of about 400 miles of Prince William Sound were struck by perhaps the strongest earthquake to have hit North America within historic time. The magnitude of this great quake has been computed by the U.S. Coast and Geodetic Survey at 8.5 on the revised Richter scale. Its epicenter was about 80 miles east-southeast of Anchorage near the head of Prince William Sound. Reportedly, the quake was felt throughout most of Alaska, including such remote points as Cape Lisburne, Point Hope, Barrow, and Umiat, 600 to 800 miles north of the epicenter on the Arctic Slope of Alaska, and at Fort Randell, 800 miles southwest at the tip of the Alaska peninsula.

Eyewitness accounts of happenings at Anchorage during and immediately after the earthquake have been reported in many publications, both technical and popular. Much valuable information has thus been gained from the objective observations of individuals who were equal to the task. Calm detachment under such trying circumstances plainly is a rare and admired virtue. Some confusion and contradiction did appear in early accounts, not surprisingly, in view of the distressing conditions under which the observations were made.

The duration of the earthquake at Anchorage can only be surmised owing to the lack of strong-motion seismograph records. Although seismographs have since been installed, none was present in Southern Alaska at the time of the quake. Intense seismic motions seem to have lasted 3 to 4 minutes, possibly longer. Where localized ground displacements occurred, as in or near landslides, strong motions may have lasted appreciably longer, after strong seismic shaking had ceased. According to Shannon and Wilson Inc. (1964, p. 11), the durations at Anchorage, timed on wrist or pocket watches, by several eyewitnesses whom they consider reliable, ranged from 4 minutes 25 seconds to 7 minutes. Even longer durations were reported outside the Anchorage area. Steinbrugge (1964, p. 62) noted that persons at Anchorage were able to accomplish several time-consuming tasks during the shaking, including such things as leaving and reentering buildings more than once, or helping other individuals to escape, despite much difficulty in standing and walking. In some areas, however, people reportedly were thrown to the ground by the force



1.—Map of Anchorage, Alaska, and vicinity showing locations of major landslides and ground cracks, much generalized.

of the acceleration and were unable to regain their footing.

There seems to be general agreement that the first waves to arrive at Anchorage had strong east-west components of movement. Objects reportedly were thrown from shelves and cupboards on east and west sides of rooms, and some individuals were said to have been propelled bodily across rooms in those directions. Figure 2 shows the trace on an asphalt tile floor made by a dresser leg. Because the epicenter was east of Anchorage, an east-west ground motion was expectable. As the shaking continued, however, the motion is said to have shifted to north-south, and tall buildings that had first rocked to-and-fro east and west began to rock north and south as well, in a complex combination of movements. Strain-fracture patterns in the walls of multistory buildings at Anchorage seem to bear out such observations. North-south components of motion, moreover, are theoretically plausible, especially in view of the severe tectonic readjustments southeast and south of Anchorage in the Prince William Sound and Kodiak Island areas. The earthquake seems to have resulted from the rupture of rock at depth beneath a very broad area, extending from some point near the epicenter southwest to the south tip of Kodiak Island (U.S. Coast and Geodetic Survey, 1964, p. 31; Grantz and others, 1964, p. 2); the arrival of seismic waves at Anchorage from such a broad source must have caused a very complex ground motion (fig. 2).

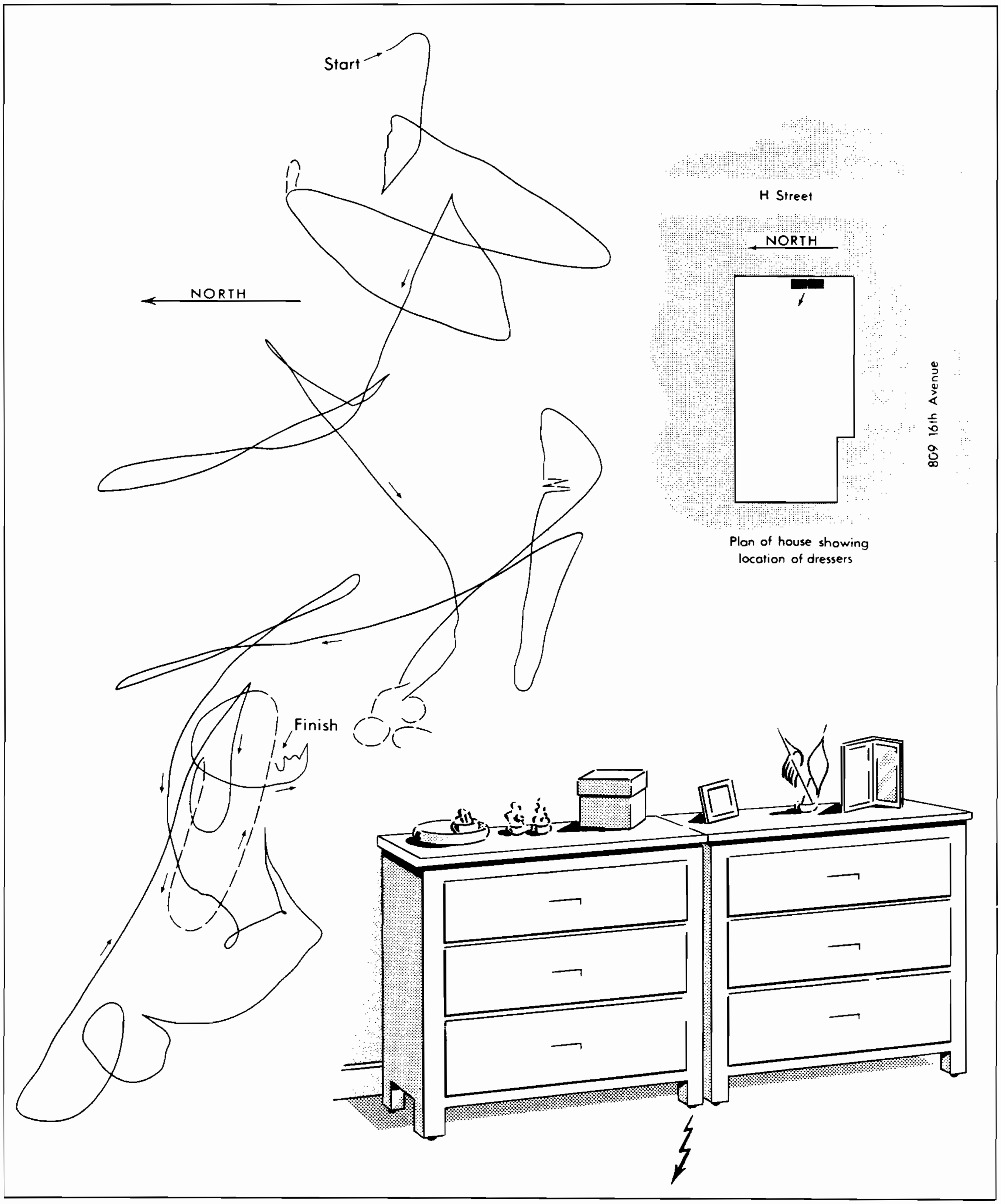
Total earthquake damage to property in the Anchorage area has not been fully evaluated and perhaps will never be fully known. Nine lives are reported to have been lost—five in the downtown area, three at Turnagain Heights, and one at the International Airport. In less than 5 minutes, more than 2,000 people, including apartment dwellers, were made homeless, according to press estimates. The loss of life was less in Anchorage than in some of the small coastal towns, where many people were killed by sea waves. But Anchorage, because of its much greater size, bore the brunt of the property damage and property losses reportedly were greater there than in all the rest of Alaska combined.

Early estimates by the Office of Emergency Planning indicated that about 75 percent of the City's total developed worth was measurably damaged. Early estimates of total damage, however, tended to be larger than later ones. According to the Anchorage Daily Times of April 9, 1964, 215 homes were destroyed in Anchorage and 157 commercial buildings were destroyed or damaged beyond repair (Grantz, Plafker, and Kachadoorian, 1964, p. 14). At Turnagain Heights alone, 75 or more dwellings were destroyed. Estimates by the Daily Times placed the damage at about $200 million. Later, the Office of Emergency Planning estimated the total damage to Alaska at about $537,600,000, of which about 60 percent was sustained by the Anchorage area (National Board of Fire Underwriters and Pacific Fire Rating Bureau, 1964, p. 5). The final total damage estimate for Alaska, exclusive of personal property and loss of income, was about $311 million (Federal Reconstruction and Development Planning Commission for Alaska, 1964, p. 11). Scores of buildings throughout Anchorage sustained damage requiring repairs costing many thousands of dollars.

The school system was hard hit. Early estimates of damage came to about $3.86 million. Classes were not in session, fortunately, so the buildings were empty. Twenty of the 26 schools in Anchorage were soon back in operation. West Anchorage High School, however, was severely damaged by seismic vibration. Government Hill Grade School, astride a landslide (fig. 3), was nearly a total loss, although plans have been made to salvage an intact part of the building outside the slide for use other than as a school. Denali Grade School was damaged by ground cracks and was closed indefinitely, pending damage evaluation and repair.

In downtown Anchorage (the L Street and Fourth Avenue landslide areas), about 30 blocks of dwellings and commercial buildings were destroyed or severely damaged (fig. 4). A new six-story apartment building, the Four Seasons, was razed. A new five-story department store (J. C. Penney Co.) was damaged beyond repair by seismic shaking and had to be torn down (fig. 5). Many automobiles in the downtown area were struck by falling debris. Twin 14-story apartment buildings, though a mile apart, sustained massive, nearly identical vibratory damage, much of it in response to vertical shearing forces caused by oscillation. Many other multistory or large-area buildings were severely damaged.

Water mains and gas, sewer, telephone, and electric systems were disrupted (Waller, Thomas, and Vorhis, 1965, p. 126). Total damage to utilities has been estimated at about $15 million (Alaskan Construction Consultants Committee, 1964, p. 11). Providentially, electric power failed at the very onset of the quake. Although the loss of power might seem to be an added hardship to the stricken city, untold numbers of fires were



2.—Horizontal trace on asphalt tile flooring left by dresser leg during earthquake. Net movement of dresser was about N. 75° W. Other legs left nearly identical traces. Sundry items on dresser tops were little disturbed. Trace reproduced by Oliver V. Kola, published by permission of Dr. M. A. Sozen. Scale of trace, ×½.

3.—Wreckage of Government Hill School. The south wing of the building, shown here, collapsed into a graben at th

head of the slide. Net slip of the graben block is shown by the displacement of the roofline. Photograph by M. G. Bonilla.

4.—View west along Fourth Avenue at head of landslide. North (right) side of street has collapsed into a graben. Vertical displacement is about 10 feet. Photograph by Mac's Foto, Anchorage, Alaska.

probably avoided because of the lack of electric current in all the severed wires—and at a time, too, when water was unavailable for fighting fire. The destruction of San Francisco by fire on the heels of the earthquake of 1906 is forcefully recalled. Fukui, Japan, was similarly destroyed in 1948. Anchorage spent the night of March 27 in total darkness, but the city had no fires and was prepared to deal with the many disrupted electric circuits when service was restored.

Roads and railroad facilities were badly damaged. In the downtown area, many streets were blocked by debris, and in landslide areas, streets and roads were completely disrupted. Differential settlement caused marginal cracking along scores of highway fills throughout the Anchorage Lowland. In The Alaska Railroad yards where landslide debris spread across trackage and damaged or destroyed maintenance sheds (fig. 6), an estimated $2,370,700 damage was sustained (Alaskan Construction Consultants Committee, 1964, p. 74). Cars and equipment were overturned, and car shops were damaged by vibration. Along the main line of the railroad, bridges failed, fills settled, and tracks were bent or buckled; at Potter, near the south margin of the Anchorage Lowland, several hundred feet of track was carried away in an area that has had a long history of repeated sliding.

At the Anchorage International Airport, the control tower failed under seismic vibration and collapsed to the ground (fig. 7), killing one occupant and injuring another. The airport terminal building, although tied structurally to the tower, was only slightly

5.—Wreckage of Penney's department store, Fifth Avenue and D Street, Anchorage. Building failed after sustained seismic shaking. Most of rubble has been cleared from streets. Photograph by George Plafker.

damaged except where it adjoined the tower. A nearby Post Office building was damaged a moderate amount (Berg and Stratta, 1964, p. 14). Almost 20,000 barrels of aviation fuel was lost from a ruptured storage tank. Runways and taxiways were slightly damaged but not put out of commission; air traffic was temporarily controlled first from a parked aircraft and then from a tower at the nearby Hood Lake landing strip (Federal Reconstruction and Development Planning Commission for Alaska, 1964, p. 25).

Facilities at the Port of Anchorage were damaged by seismic vibration and ground cracks. Four cranes jumped their tracks and in so doing, damaged their undercarriages and counterweight arms (Berg and Stratta, 1964, p. 44 and 47). Two steel storage tanks were toppled and destroyed. Nearby oil-storage tanks were damaged superficially, but large quantities of fuel oil were lost. An expected seismic sea wave fortunately did not materialize.

Landslides caused the greatest devastation in the Anchorage area. Great slides occurred in the downtown business section (Fourth Avenue slide), in the lower downtown business and residential area (L Street slide), at Government Hill, and at Turnagain Heights. Less devastating slides occurred in undeveloped or unpopulated areas near the Alaskan Native Service Hospital, at Romig Hill, at the Alaska Highway Department garage, at Bluff Road near the U.S. Army Corps of Engineers Headquarters Building, on the west bluff of Government Hill, at Point Woronzof, Point Campbell, Cairn Point, and on the west side of Knik Arm near Sleeper landing strip in an uninhabited area recently opened to homesteading. The slide at Potter has already been mentioned.

Capricious damage was caused by ground cracks. Such damage was mostly localized, and many areas were spared. Cracking was most common behind the heads of landslides, but it was also prevalent throughout the lowland in areas underlain by clay or silt or artificial fills on muskeg. Differential compaction was a common cause. Cracking was minimal in areas of thick ground moraine.

The flow of streams, such as Ship Creek and Chester Creek, was greatly reduced, temporarily, by percolation of water into cracks. Gages in water wells recorded

6.—Damage to yards of The Alaska Railroad, looking northwest, caused by earthflow at toe of Government Hill slide. Note undamaged water tank in background.

marked fluctuations, not just in Anchorage and southeast Alaska, but at places as distant as Georgia, Florida, and Puerto Rico (Waller and others, 1965, p. 126, 131).

For a city of its size, Anchorage has in residence a large contingent of geologists and other earth scientists. Within hours after the quake, many of these individuals had begun independent investigations of the location, severity, nature, and causes of earthquake damage. On March 29, 2 days after the earthquake, a small group of geologists under the leadership of Dr. Lidia Selkregg of the Alaska State Housing Authority was commissioned by the housing authority and the city of Anchorage to outline "necessary and immediate courses of action" to be taken by the city. This group became known as the Engineer-

7.—Wreckage of control tower at Anchorage International Airport. Six-story tower failed under sustained seismic shaking. Photograph by George Plafker.

ing Geology Evaluation Group (Schmidt, 1964, p. 13). Its membership rose to 40 after a call was made for technical help.

The group immediately began a program of mapping and data gathering. They contracted for aerial photography, started a drilling program in the major slide areas in cooperation with the Alaska Department of Highways, and ran standard laboratory tests on the various soil zones involved in the land failures. Drilling and soil testing were subsequently taken over and greatly expanded by the office of the U.S. Army District Engineer at the request of the city of Anchorage, under a contract dated 25 April 1964 between the District Engineer and Shannon and Wilson, Inc., soil mechanics and foundation engineers, Seattle, Wash. The Corps of Engineers meanwhile had immediately begun disaster relief.

The work of the Engineering Geology Evaluation Group deserves high praise. Its preliminary findings and recommendations were completed on April 12, 1964, 2 weeks after the earthquake, and a final report was completed on May 8, 1964. The findings and conclusions of the group provided the basis for many subsequent investigations by other agencies. Much of the information presented by this paper has been derived from the work of this group.

GEOLOGIC SETTING

Anchorage is near the east border of a deep structural trough filled with moderately consolidated Tertiary rocks that underlie Cook Inlet and extend northeastward toward Mount McKinley at the head of the Susitna Lowland (Capps, 1916, 1940). At Anchorage, Tertiary rocks are covered by Pleistocene deposits, but they have been penetrated by drill holes.

The Chugach Mountains just east of Anchorage consist chiefly of Mesozoic argillite and graywacke, variably deformed and metamorphosed, and intruded by small bodies of igneous rock (Capps, 1916, p. 153). The nature of the contact between the soft Tertiary rocks of the lowland and the harder Mesozoic rocks of the mountains has not been ascertained, owing partly to a lack of exposures and partly to a lack of study, but the very straight trace of the mountain front and the seeming truncation of structural trends by the mountain front suggest that the contact is a fault. On similar evidence, Karlstrom (1964, p. 21) inferred that the Kenai Mountains—a southward structural and topographic extension of the Chugach Mountains—are also bounded by a fault on the west. If these inferences are correct, a major structural displacement separates the Cook Inlet

Lowland from the bordering mountains.

Cook Inlet and the Anchorage Lowland were occupied repeatedly by large piedmont glaciers in Pleistocene time (Karlstrom, 1957, 1964; Miller and Dobrovolny, 1959). Karlstrom has distinguished five major Pleistocene glacial advances in Cook Inlet, although not all reached the Anchorage area. At the site of Anchorage, Pleistocene deposits accumulated to a thickness of 600 feet or more. In general, these deposits seem to thicken westward from the mountain front toward Cook Inlet. They exerted a vital influence on the location and the extent of earthquake damage in the Anchorage area.

Because the Pleistocene deposits of the Anchorage area have been described previously in considerable detail (Miller and Dobrovolny, 1959; Karlstrom, 1964; Cederstrom, Trainer, and Waller, 1964), they will be described but briefly here. They consist chiefly of three categories of material: glacial till, deposited as ground moraine; proglacial silty clays (including the Bootlegger Cove Clay) deposited in estuarine-marine or lacustrine-estuarine environments; and fluvioglacial deposits of several types, but chiefly outwash sand and gravel.

Most of Anchorage lies on late glacial (Naptowne-Wisconsin) outwash deposited in front of the youngest Pleistocene glacier that entered the area. This glacier constructed a large end moraine north of the city at Elmendorf Air Force Base (Miller and Dobrovolny, 1959, p. 59). Remnants of the moraine are also preserved west of Knik Arm about 2½ to 3 miles northeast of Point MacKenzie. Outwash sand and gravel from this glacier spread southward across the Anchorage Lowland and buried ground moraine and Bootlegger Cove Clay alike to depths as great as 60 feet. In general, the outwash thins toward the west and south away from its source. It wedges out completely between Turnagain Heights and Point Woronzof.

East of Anchorage along the foot of the Chugach Mountains, a massive lateral moraine was deposited by one or more of the pre-Wisconsin glaciers that occupied the area (Miller and Dobrovolny, 1959, p. 21). This moraine stands generally 1,000 to 1,200 feet above sea level and has a maximum altitude of about 1,400 feet. Because the area commands impressive views across Cook Inlet toward the snow-swept Alaska Range and Mount McKinley, it has experienced a mild real estate boom, and the next few years should see extensive urbanization of its heights. The moraine fared well in the March 27 earthquake; structures built on it were little, if at all, disturbed.

BOOTLEGGER COVE CLAY

The Bootlegger Cove Clay has become renowned, since the March 27 earthquake, for its part in the devastation. Most of the severe damage in the Anchorage area is traceable to failure within this formation, and an understanding of the mechanisms of landsliding in the Anchorage area requires some knowledge of its character and physical properties. Because of its critical relationship to the landsliding, it has been studied intensively since the earthquake, by the Corps of Engineers and by consultants to the corps.

The Bootlegger Cove Clay was named and first described by Miller and Dobrovolny (1959, p. 35–48). In general the formation consists of three physically distinct but gradational zones—an upper and a lower stiff competent zone and a central weak sensitive zone. Failures occurred chiefly in the central zone. Waves generated by the earthquake led to a drastic loss of strength, a consequent failure of dynamically sensitive saturated sand, silt, and silty clay, and a resultant disruption of the ground surface by landsliding (Shannon and Wilson, Inc., 1964, p. 1–3).

The Bootlegger Cove Clay underlies most of Anchorage and much of the adjacent area (Miller and Dobrovolny, 1959, pls. 3 and 6; Trainer and Waller, 1965). Along Knik Arm, it is exposed almost continuously from a point about three-quarters of a mile east of Point Woronzof northward at least to the Eagle River. Just north of Anchorage, it passes beneath the Elmendorf Moraine. Toward the east and southeast, it laps across older glacial deposits and thins out against higher ground in the morainal belt at the base of the Chugach Mountains. South of Anchorage, it is continuous beneath a cover of sand and muskeg west from the Seward-Anchorage Highway to the east base of the Point Woronzof-Point Campbell highland area and south to Turnagain Arm. It laps up and thins out against the deltal sand and gravel of the Point

Woronzof-Point Campbell highland area.

As described by Miller and Dobrovolny (1959, p. 39), the Bootlegger Cove Clay consists chiefly of silty clay, light gray (N7)[1] when dry and dark greenish gray (5GY 4/1) when wet. Very commonly the upper several inches to several feet immediately beneath the overlying outwash is oxidized to yellowish gray. The clay generally is delicately laminated in layers a fraction of a millimeter to several centimeters thick, although some intervals several feet thick are free of visible bedding. Throughout the clay are scattered layers of sand, some mere partings but others 25 feet or more thick. Such layers are lenticular and can only be traced short distances. In at least one landslide (Fourth Avenue), liquefaction of a sand layer is believed to have been the chief immediate cause of failure (Shannon and Wilson, Inc., 1964, p. 41).

Scattered pebbles are present throughout the Bootlegger Cove Clay. Cobbles and boulders are present also but are rare. These stones contributed nothing to the failure of the clay and had no part in the landsliding, but they do indicate the periglacial environment in which the clay was deposited. Only ice rafting could have emplaced the larger boulders. Pebbles are most numerous in the lower part of the formation. On the west side of Knik Arm opposite Cairn Point, where the base of the formation is above tide water, the clay grades downward into stony till, seemingly without a clear-cut contact.

[1] Numerical designations refer to the Rock-Color Chart number (Goddard and others, 1948).

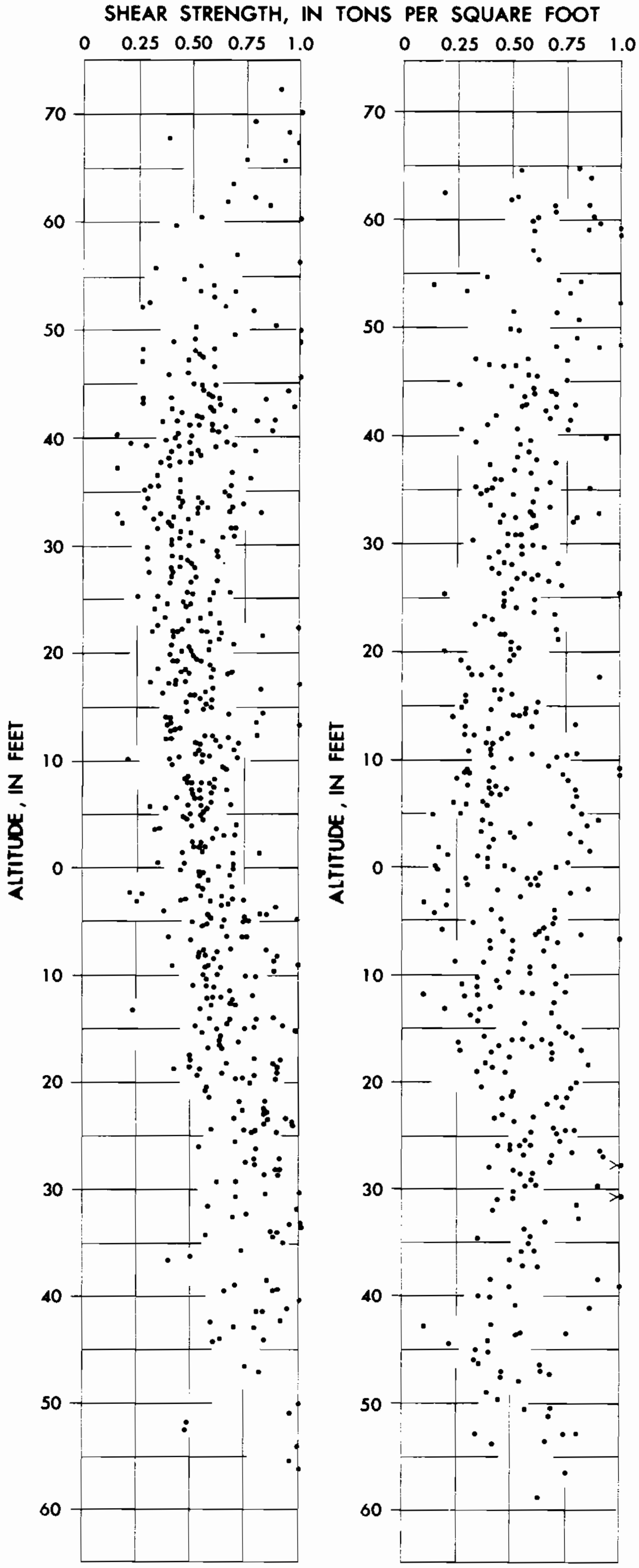
PHYSICAL PROPERTIES

Soon after the earthquake, work started by the Engineering Geology Evaluation Group (1964) and later taken over by the U.S. Army Corps of Engineers (Shannon and Wilson, Inc., 1964) showed that the physical properties of the Bootlegger Cove Clay are far from uniform. Rather, a medial zone of low static shear strength and high sensitivity grades upward and downward into stiffer nonsensitive zones. Sensitivity is defined as the ratio of undisturbed shear strength of a soil sample to remolded (kneaded or squeezed) shear strength of the same sample, regardless of the cause (Terzaghi and Peck, 1948, p. 33).

Numerous physical tests by the Corps of Engineers—both in the field and in the laboratory—were made to analyze the causes of failure in the Bootlegger Cove Clay and find possible remedies. These tests are described in detail by Shannon and Wilson, Inc. (1964, p. 13–30); most of the information in the following paragraphs was abstracted from their report. About 150 borings were drilled within and adjacent to the several landslide areas, either into or through the sensitive zone of the Bootlegger Cove Clay. Most borings were sampled at 5-foot intervals, and continuous samples were obtained at selected locations. Undisturbed samples were collected in 3- by 36- or 3- by 37-inch steel Shelby tubes. Five large-diameter "bucket auger" holes were drilled by the Corps for visual inspection of critical zones in the clay, for in-place shear-strength tests, and for carving out undisturbed samples by hand. Vane-shear tests in standard 3-inch holes were made at 12 locations in the Fourth Avenue, L Street, and Turnagain Heights slide areas to measure in-place strengths and sensitivities of the clay. Similar tests were made on undisturbed samples both in the field and in the laboratory. In the laboratory, samples were also analyzed for Atterberg limits (see p. A14), natural-water content, triaxial shear, sensitivity, dynamic modulus, and consolidation strength.

SHEAR STRENGTH

Shear strength was measured in torsional-vane-shear tests, undrained triaxial tests, consolidated-undrained triaxial tests, and consolidated-drained triaxial tests; strain, stress, volume change, and pore-water pressure were recorded throughout the tests (Shannon and Wilson, Inc., 1964, p. 28). The results of vane-shear tests for the Fourth Avenue and Turnagain Heights slide areas are shown in figure 8. Triaxial tests performed on the clay generally correlated well with the vane-shear tests within the ranges of strengths at which failure was prevalent. Tests correlated well in the range 0.4 to 0.5 tsf (tons per square foot), but for clays of higher strength, the triaxial tests generally yielded somewhat higher values than the vane tests. Most of the samples of clay from critical depths in the landslides had initial shear strengths between 0.2 and 0.7 tsf, and in this range the ratio of torsional-vane-shear-test results to triaxial-compression-test results is generally between 0.7 and 1.5 (Shannon and Wilson, Inc., 1964, p. B8). This ratio indicates a fairly good correlation between test methods. In any event, the vane-shear-test results shown by the graphs of figure 8 clearly demonstrate the generally lower shear strength of the clay at intermediate depths as compared with shallower and deeper layers.



8.—Summary of torsional-vane-shear strengths of selected borings, Bootlegger Cove Clay, from Fourth Avenue slide area (left) and Turnagain Heights slide area (right). Simplified from Shannon and Wilson, Inc. (1964, pl. B15).

Torsional-vane-shear apparatus was also used to measure the strength of the Bootlegger Cove Clay under sustained stress, because it was recognized that the duration of stress influences the resulting shear strength shown by the tested specimen (Shannon and Wilson, Inc., 1964, p. B7). At a given level (0.1 tsf), shear stress was applied for varying lengths of time. These tests showed that the Bootlegger Cove Clay loses shear strength as stress application is prolonged. In other words, the longer the stress is sustained, the lower the shear strengths.

SENSITIVITY

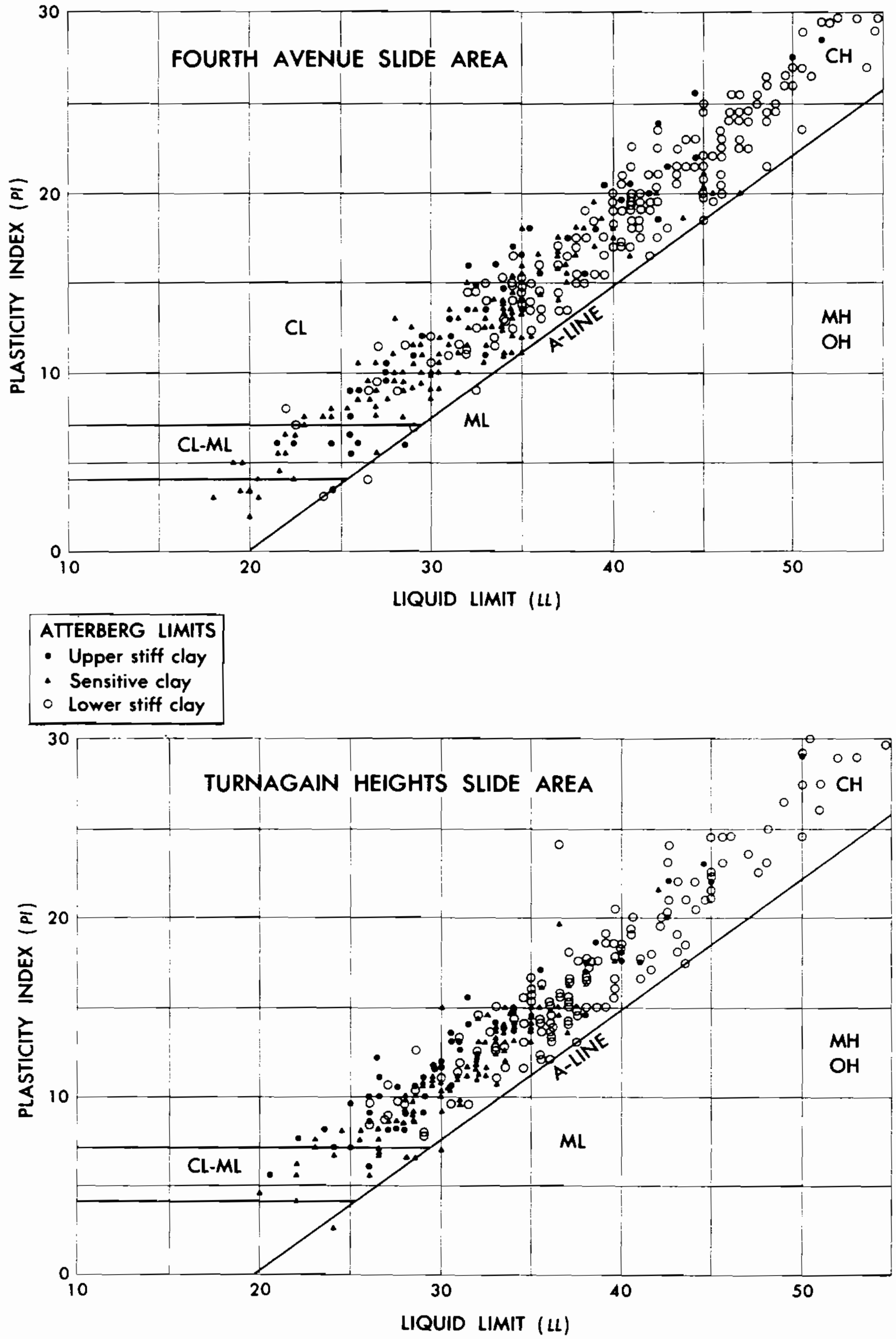
Sensitivity has previously been defined as the ratio of undisturbed shear strength to remolded shear strength. For example, a clay having an undisturbed shear strength of 0.4 tsf and a remolded shear strength of 0.02 tsf has a sensitivity of 20. Sensitivies were measured by the Corps of Engineers on about 2,100 selected samples of the Bootlegger Cove Clay by means of field and laboratory vane-shear devices (Shannon and Wilson, Inc., 1964, p. 27), and a record of sensitivities was plotted on the boring logs. Of the 2,100 samples tested, 302, or 14 percent, had sensitivities greater than 10; 125, or 6 percent, had sensitivities greater than 20; 40, or 1.9 percent, had sensitivities greater than 30; and 11, or 0.5 percent, had sensitivities greater than 40. The highest sensitivity measured was 60.

The sensitive zone generally ranges in thickness from about 20 to 30 feet, but it is as much as 40 feet thick in some places and is nonexistent in others. In most places the center of the sensitive zone is above sea level, generally 10 to 20 feet above, but in parts of the Turnagain Heights area it is below sea level. At no place has sensitive clay been observed higher than about 50 feet above sea level or lower than about 30 feet below sea level. Laboratory tests by the Corps indicate that the shear strength of the sensitive clay ranges from less than 0.15 to about 0.40 tsf, and the sensitivity ranges from about 25 to as high as 60. The maximum shear strengths of the stiffer clays exceed 1 tsf.

ATTERBERG LIMITS

In a laboratory analysis of fine-grained cohesive soils, it is customary to perform tests that yield information on the plasticity of the soils. These tests include measurements of natural water content (*Wn*) liquid limit (*LL*) and plastic limit (*PL*). The liquid limit is the water content in percentage of dry weight at which the soil passes from the liquid state into the plastic state. Similarly, the plastic limit is the water content of the soil at the boundary between the plastic state and the solid state. These limits of consistency are defined by standardized test procedures and are known as the Atterberg limits (Terzaghi and Peck, 1948, p. 32–36). The numerical difference between the liquid limit and the plastic limit is called the plasticity index (*PI*). The plasticity index represents the range of moisture content within which the soil is plastic. Clayey soils thus have higher plasticity indices than nonclayey or silty soils because they remain plastic through a wider range of moisture content (Stokes and Varnes, 1955, p. 85, 110; U.S. Bureau of Reclamation, 1960, p. 8, 28).

The plasticity charts (fig. 9) show the statistical relationship of liquid limit to plasticity index of the Bootlegger Cove Clay in the Fourth Avenue and Turnagain Heights slide areas, as measured for the Corps of Engineers by

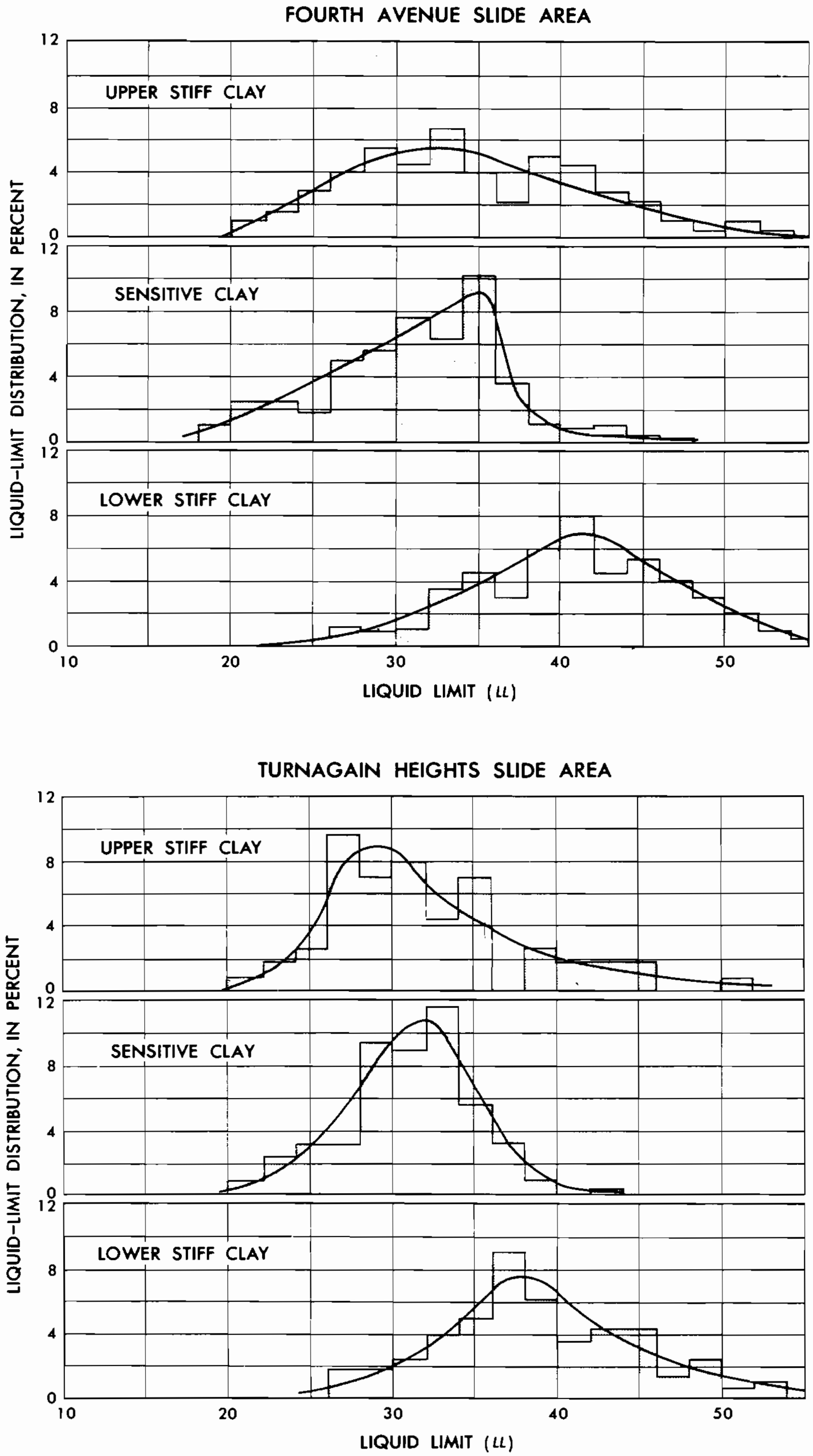


9.—Plasticity charts from selected borings, Bootlegger Cove Clay, Fourth Avenue and Turnagain Heights slide area. Reprinted from Shannon and Wilson, Inc. (1964, pl. B16). CL, inorganic clay of low to medium plasticity; ML, inorganic silts and very fine sand; CH, inorganic clay of high plasticity; MH, inorganic silt; OH, organic clay.

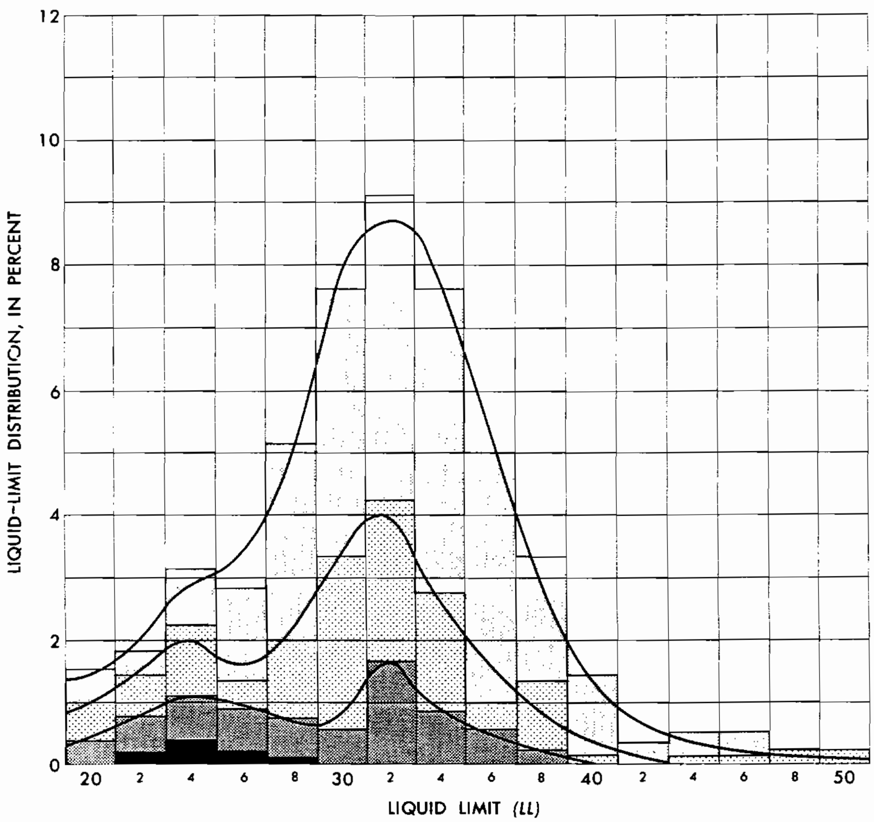
Shannon and Wilson, Inc. (1964). These charts show generally higher plasticity indices and liquid limits for the lower stiff clay than for the upper stiff clay or the sensitive clay. Differences between the upper stiff clay and the sensitive clay are not obvious on these charts, but they are quite apparent on statistical liquid-limit distribution curves (fig. 10). Nearly all points plotted on the plasticity charts lie above the "A" line—the line which separates soils of high cohesion and low permeability above from soils of low cohesion and high permeability below (Casagrande, 1947, p. 801).

The statistical liquid-limit distribution curves (fig. 10) show the distinctions between the three zones of the Bootlegger Cove Clay in the Fourth Avenue and Turnagain Heights areas. These curves bring out the differences between the upper stiff zone and the sensitive zone, and also the higher liquid limit of the lower stiff zone. The statistical relationship of sensitivity to liquid limit is shown by figure 11. In general, about 80 percent of the soil samples tested had liquid limits within the range of 22 to 48 percent, regardless of their vertical position in the formation. About 80 percent of the liquid limits of the sensitive clay range from 22 to 38 percent, and about 80 percent from the lower stiff clay range from 32 to 50 percent. As sensitivity increases, the upper limit of the liquid limit decreases. Most clay samples of sensitivities greater than 10 had liquid limits in the range of 20 to 40 percent, and no sensitive-clay sample had a liquid limit above 45 percent. For sensitivities greater than 40, the range of liquid limits was only 22 to 28 percent (Shannon and Wilson, Inc., 1964, p. B11). Several samples analyzed had a natural-water content greater than their liquid limits.

To relate the limit values of a soil to its natural-water content, the water-plasticity ratio, or liquidity index, is sometimes used (Casagrande and Fadum, 1944, p. 341). This ratio is equal to the natural-water content of the soil minus the plastic limit divided by



10.—Statistical liquid-limit distribution curves from selected borings, Bootlegger Cove Clay. Reprinted from Shannon and Wilson, Inc. (1964, pl. B16).



11.—**Liquid-limit sensitivity distribution curves, Bootlegger Cove Clay. Reprinted from Shannon and Wilson, Inc. (1965, pl. B17).**

the liquid limit minus the plastic limit; that is $(Wn-PL)/(LL-PL)$. For the Bootlegger Cove Clay, it was found that in most soils having a water-plasticity ratio greater than 1.0 the liquid limit ranged from 20 to 40 percent, and in soils having a water-plasticity ratio greater than 2.0 the liquid limit lay in the range 19 to 27 percent. A direct correlation also was found between sensitivity and the water-plasticity ratio (Shannon and Wilson, Inc., 1964, p. B11, pl. B17; see fig. 12, next page). Thus in the Bootlegger Cove Clay a high water-plasticity ratio is related to low shear strength, high sensitivity, and high susceptibility to failure.

VIBRATION TESTS

Representative samples of sand and clay from the various slides were subjected to vibration tests from which the dynamic Young's modulus and the dynamic shear modulus of elasticity could be computed. These tests were designed to simulate in part the vibratory effects of the earthquake on the sand and clay. The vibration apparatus, developed by Shannon and Wilson, Inc. (1964, p. E3), subjected a cylindrical specimen to a sinusoidal vibration in a longitudinal or torsional mode in such a way that the frequency corresponded to a state of resonance. Both longitudinal and torsional resonance frequencies increased as straight-line functions of increased confining pressure.

The shear modulus of the sands was found to be dependent on the confining pressure. The shear modulus of the stiff clay was about 5,000 psi (pounds per square inch), and that of the soft sensitive clay was about 1,500 psi (Shannon and Wilson, Inc., 1964, p. 28). In nearly all tests, the value of the shear modulus and of Young's modulus increased as straight-line functions of increased confining pressure.

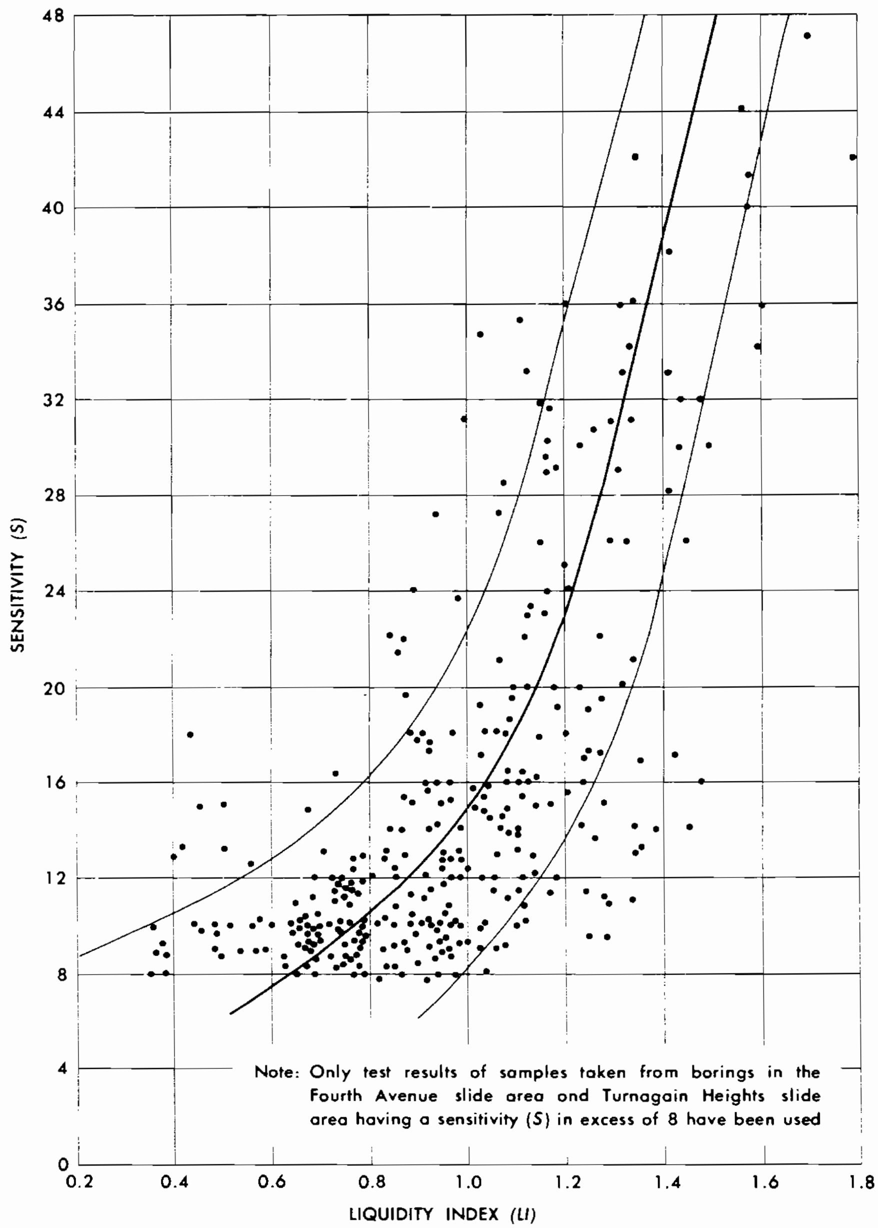
PULSATING-LOAD TESTS

Pulsating-load tests on saturated fine silty sand and on undisturbed samples of silty clay and clayey silt were run for the Corps of Engineers by H. B. Seed and K. L. Lee of the University of California at Berkeley (*in* Shannon and Wilson, Inc., 1964). These tests demonstrated the susceptibility of the Bootlegger Cove Clay and associated sands to loss of strength under pulsating-load conditions such as would be induced by an earthquake. In other words, the shear strength of the samples was markedly less under pulsating-load conditions (simulating an earthquake), than under static load.

Seed's apparatus was so designed that applied deviator (axial) stress could be increased and decreased in one direction or could be changed from vertical to horizontal during each stress cycle, and thus the direction of shear stress in the specimen could be reversed. The latter test is called a "stress-reversing test."

Because it is not feasible to collect undisturbed sand samples, sand specimens were artificially deposited in water in the laboratory under conditions that simulated natural sedimentation. The sand was then subjected to pressures comparable to the in-place overburden pressures of the natural environment, about 1.8 kg per cm². Under stress-reversing conditions, sand specimens showed negligible strains, but they did show progressive increases in pore-water pressures with increasing numbers of stress cycles. Failure abruptly followed a sudden increase in pore-water pressure equal in magnitude to the applied confining pressure, accompanied by large strains and complete liquefaction. Liquefaction followed 50 cycles of pulsating deviator stress of about 0.5 kg per cm² applied at a frequency of about 2 cycles per second. This stress level is only about 10 percent of the strength of an identical specimen tested under static-loading conditions.

Clay samples were found to fail at levels of pulsating deviator stress substantially lower than the undrained compressive strength. Undrained compressive strengths of the specimens tested ranged from about 0.4 to 1.8 kg per cm², and sensitivities ranged from about 15 to 60. Moreover, under stress-reversing conditions, the level of failure was substantially lower than under non-stress-reversing conditions. Thus, under a non-stress-reversing pulsating load at a frequency of 2 cycles per second, failure generally followed 50 cycles of maximum stress of 80 to 100 percent of the static undrained strength, after a gradual buildup of strain, whereas under stress-reversing conditions, failure occurred abruptly without prior significant strain after a maximum stress of only about 55 percent of the static undrained strength. The natural vibration of an earth-



12.—Relation between sensitivity and liquidity index (water-plasticity ratio), Bootlegger Cove Clay. Reprinted from Shannon and Wilson, Inc. (1964, pl. B17).

quake, according to Seed, probably has a pulse form somewhere between the two types used in these laboratory tests. It is concluded, therefore, that under earthquake conditions, failure would occur at a stress level below the static undrained strength of the clay, that is, somewhere between the limits indicated by the two types of pulsating stress applied in the laboratory.

SPECIFIC GRAVITY

Specific gravity was measured by Shannon and Wilson, Inc. (1964, p. B11–B12) on 26 test specimens from various depths at seven different slide areas. Of the specimens tested, 1 had a specific gravity of 2.67, but the other 25 ranged from 2.71 to 2.82, the average being 2.78. No correlation between specific gravity and shear strength or sensitivity was apparent.

PHYSICO-CHEMICAL TESTS

Physico-chemical analyses of six samples were made for the Corps of Engineers by J. K. Mitchell (*in* Shannon and Wilson, Inc., 1964, p. I3). Four of the samples—two from Turnagain Heights and two from the Fourth Avenue area—were cuts from core intervals that also were analyzed mineralogically. The other two samples came from the L Street slide. Each sample was tested for pH, salt content, conductivity, and base-exchange capacity. Values of pH ranged from 8.2 to 10.3. These relatively high base values, according to Mitchell, are "consistent with high sensitivity because of the tendency of a high pH to promote dispersion of remolded samples" (*in* Shannon and Wilson, Inc., 1964, p. I3).

The salt content of the samples, expressed in terms of equivalent NaCl, ranged from 1.03 grams per liter to somewhat less than 6 grams per liter. "By way of comparison," Mitchell points out, "the salt content of sea water is about 34 grams per liter. Thus if the Anchorage clays are of marine origin then the results of these tests would suggest that they have been leached of salt since their original deposition. Such leaching could be important in generating a high sensitivity and stable suspensions after remolding." In support of this view, a marine depositional environment of near normal salinity, at sample depths spanning the zone of landslide failure, is indicated by the microfaunal assemblages in the clay (Patsy J. Smith, written commun., 1964). The paleontology of the clay is discussed on page A20.

In measuring electrical conductivity, Mitchell (*in* Shannon and Wilson, Inc., 1964, p. I4) found that remolded clay had consistently greater conductivity at all frequencies than undisturbed clay. Remolded clay generally also had a greater variation in conductivity as frequency was varied. Mitchell attributed these differences to greater dispersion (parallelism of clay particles) and smaller average effective particle size in the remolded samples.

Base-exchange capacities of the six samples studied by Mitchell ranged from 4.6 to 10.9 milliequivalents per 100 g of dry clay. Of the three cations determined—sodium, calcium, and magnesium—Mitchell found that calcium was the most abundant in solution in all but one sample. Sodium was next and magnesium was least abundant. This ratio also suggests that authigenic brine was leached from the clay by the action of ground water; hence that leaching may have been a cause of the sensitivity.

MINERALOGY

Not much mineralogic work has been done on the Bootlegger Cove Clay. Seven samples collected by Miller and Dobrovolny (1959, p. 42–43) from four localities and from seven different parts of the section were analyzed by X-ray diffraction in two fractions each—clay ($<2\mu$) and silt (2 to 74μ). Both fractions of each sample contained quartz, feldspar, mica, and chlorite as the chief constituents. Some contained montmorillonite, mixed layered chlorite-montmorillonite, or mixed layered chlorite-montmorillonite-hornblende (Miller and Dobrovolny, 1959, p. 42). Quartz predominated in most silt fractions, followed by feldspar, but it was subordinate in the clay fractions, in which chlorite predominated. These seven samples resembled, in composition and general proportions of minerals, samples of other silty clay collected from lake, estuarine, and till deposits of the same general vicinity but of different ages. Their compositions, therefore, probably reflect parent material more than diagenesis.

A single clay sample collected at Turnagain Heights after the earthquake was found to contain chlorite, illite, quartz, and feldspar in its clay fraction (Grantz, Plafker, and Kachadoorian, 1964, p. 26). Its composition by size was 25 percent clay, 61 percent silt, and 14 percent fine sand.

Samples of Bootlegger Cove Clay from four drill holes, two from the Fourth Avenue area and two from Turnagain Heights, were analyzed by X-ray, differential thermal, thermogravimetric, and petrographic means by T. S. Shevlin for the Corps of Engineers (Shannon and Wilson, Inc., 1964, p. H3). Two of the samples were sensitive; two were stiff. Shevlin found that quartz predominated in all samples; its peak intensities (X-ray) were greater than any other mineral in all size fractions analyzed. Feldspar was present in finely divided form in all samples, and kaolinite, illite, and chlorite were also present. One of the stiff samples was gritty and much siltier than the others, but there was no consistent variation either in size or mineral content between the sensitive and the stiff samples. Shevlin's table 1 is reproduced in part at top of page 20 to show the distribution of particle sizes.

Sample [1]	Depth (feet)	Classification	Percent finer than particle size (microns)					
			38	10	8	4	2	1
A117AX	77.6–78.3	Sensitive	99.6	81.5	78.9	64.7	49.0	36.0
A117AX	124.5–125.2	Stiff	96.9	96.7	92.2	73.4	52.7	38.5
C108B	73.2–73.9	Sensitive	96.8	64.0	60.3	46.6	33.3	24.4
C117	116.7–117.4	Stiff	81.8	26.7	22.3	12.9	8.1	5.5

[1] Samples prefixed with A are from Fourth Avenue area; samples prefixed with C are from Turnagain Heights.

DEPOSITIONAL ENVIRONMENT

Recent paleontologic studies by Patsy J. Smith (written commun., 1964) indicate that the Bootlegger Cove Clay is of marine origin throughout but that it was deposited in environments of variable salinity. The depositional environment bears on the origin and distribution of sensitivity in the clay. Clay deposited in saline waters tends to flocculate. Flocculation occurs when negative particle surfaces are attracted to positive particle edges by electrostatic forces (Lambe, 1958, p. 8; Rosenqvist, 1962, pl. 5; Meade, 1964, p. B4–B5); flocculated clay in turn acquires sensitivity when the interparticle bond is destroyed, as when ground water leaches the cation and thereby diminishes the electrolyte concentration (Mitchell, 1956, p. 693). Such leaching is promoted when marine clay beds are elevated above sea level. According to Bjerrum (1954, p. 49; 1955, p. 108), the reduced electrolyte concentration, by decreasing the activity of the clay minerals, leads to a lowering of the Atterberg limits, which in turn reduces the shear strength of the clay by as much as 30 percent.

Until recently the depositional environment of the Bootlegger Cove Clay was in serious doubt, if not dispute (Miller and Dobrovolny, 1959, p. 44; Schmidt, 1963, p. 350; Karlstrom, 1964, p. 35; Cederstrom, Trainer, and Waller, 1964, p. 30). A brief review of past thinking, therefore, seems warranted. The environment has been considered to have been lacustrine, estuarine, or marine, or some combination thereof. Marine organisms had been noted in the Bootlegger Cove Clay by several investigators, although some doubt had existed as to whether the specimens noted were in place or had been cast ashore by storm waves (Miller and Dobrovolny, 1959, p. 45). Trainer (*in* Miller and Dobrovolny, 1959, p. 45) found estuarine mollusks that he was convinced were in place. Schmidt (1963, p. 350) verified Trainer's find and added an abundant microfauna in confirmation. Miller and Dobrovolny (1959, p. 46) reasoned that varvelike beds and laminations high in the clay near Cairn Point indicated probable fresh-water deposition, a view subsequently shared by Cederstrom, Trainer, and Waller (1964, p. 32) but on slightly different evidence—the presence of well-sorted interbedded sands. Karlstrom (1964, p. 38) expressed the opinion that "the Bootlegger Cove Clay records proglacial-lake sedimentation * * * with an intervening interval of marine deposition"—in other words, fresh-water deposits separated by salt-water deposits.

Smith's conclusions (written commun., 1964) are based on studies of microfossils from drill samples collected by the Corps of Engineers. These continuously cored samples afforded a chance not provided by surface exposures to examine the depositional environment of the clay from the top of the formation down through and below the sensitive zone. Smith concluded that the formation was marine throughout the interval studied but that the upper part was deposited in a deltaic environment of low variable salinity such as now exists in the Yukon and Koskokwim deltas. Fossils from the lower part of the formation, including the sensitive zone, indicated a shallow (25± meters) marine environment probably of near-normal salinity. Vertical variations in the depositional environment, as indicated by fossils, thus may provide an explanation for the zonal character of the sensitivity: the stiffer clays accumulated in water of low salinity and the more sensitive clays accumulated in waters of near-normal salinity. Since deposition, reduction of the brine concentration by leaching probably has altered the plasticity of the clay and increased the sensitivity.

Fossils identified by Smith, and their relative abundance, are shown in tables 1 and 2.

TABLE 1.—*Foraminifera from Bootlegger Cove Clay, Anchorage, Alaska, Shannon and Wilson, Inc. (1964) boring A120 A*

[Abundances estimated: A, abundant; C, common; F, few; R, rare]

Species	Sample interval (feet)																					
	29.9–32.0	34.9–37.0	39.7–42.0	44.7–46.0	54.7–56.5	64.0–66.5	69.0–71.5	71.5–74.0	79.0–81.5	80.0–86.5	89.9–90.5	92.3–93.9	96.4–98.9	101.6–103.8	106.3–108.8	115.9–118.4	120.9–122.5	125.0–127.5	130.0–132.5	136.8–137.9	137.5–142.0	142.2–144.3
Protelphidium cf. *orbiculare* (Brady)	F	F		F	R	F	R	R	A		F	F	R	C	A	A	F	F	F		C	F
Elphidium clavatum Cushman						F							R		F							
frigidum Cushman				R																		
bartletti Cushman														F		F						
Pateoris hauerinoides Loeblich and Tappan														F	F	R						
Polymorphina sp									R						F			R				
Fissurina sp	R													R	R							
Buliminella curta Cushman																			R			
Ostracodes								R	R			F			A	R						

TABLE 2.—*Foraminifera from Bootlegger Cove Clay, Anchorage, Alaska, Shannon and Wilson, Inc. (1964) borings A 110 A and B*

[Abundances estimated: AA, very abundant; A, abundant; C, common; F, few; R, rare]

Species	Sample interval (feet) and boring letter																		
	27–38 (B)	28–28.7 (B)	28.7–29.5 (B)	29.1–30 (B)	30–30.9 (B)	30.9–31.8 (B)	32–32.8 (B)	32.8–33.6 (B)	33.6–34.4 (B)	38.3–39 (A)	44.6–45.5 (A)	45.5–46.5 (A)	48.5–49.5 (A)	49.5–50.5 (A)	50.5–51.5 (A)	51.5–52.5 (A)	53.3–54.5 (A)	54.3–55.5 (B)	54.9–56 (A)
Protelphidium cf. *orbiculare* (Brady)	R			R	F	R	AA	C	C	AA	B	AA	A	A	A	A	A	A	A
orbiculare (Brady)														F	C				
Elphidium clavatum Cushman			F	F		R													
subarcticum Cushman										C									
bartletti Cushman													F		F	C	C	C	C
incertum (Williamson)																R			
Elphidiella groenlandica (Cushman)								R		C		F	F	C					
Quinqueloculina cf. *seminula* (Linné)						R							C	C	C				
Pateoris hauerinoides Loeblich and Tappan										C				C	C	C	C	F	C
Polymorphina sp										F			C	C		F	R	R	F
Fissurina sp				R			R				R	R	R	F	F	A	C	F	F
Dentalina sp														R	F	F		R	R
Cassidulina islandica Nørvang			R							R									
Buliminella curta Cushman										R		R							
Bolivina pseudopunctata Höglund												R				R			
Buccella frigida (Cushman)				R								R							
inusitata Anderson								R								C		R	
Rosalina sp																R			
Globigerina bulloides d'Orbigny				R								R							
pachyderma (Ehrenberg)									R			F				R	R		
Ostracodes	R	R	R		R	R	C	F		F			C	C	C	C	C	C	C
Mollusk fragments													R						

EARTHQUAKE EFFECTS

DIRECT SEISMIC EFFECTS

Most of the more spectacular structural damage in the Anchorage area resulted from secondary causes such as landslides and ground cracks, themselves triggered by seismic vibration. Structural damage due directly to seismic vibration was subordinate in terms of total property damage and financial loss and, for the most part, was relatively unspectacular. Nevertheless, the extent of such effects must not be underestimated. The cumulative damage was impressive, and had there been no landslides at Anchorage, vibratory effects of the earthquake undoubtedly would have received more attention by investigators and by the press.

Vibratory damage to buildings and other structures is not strictly a geologic effect, but the extent and amplitude of vibration is dependent partly on geologic factors such as foundation and subfoundation conditions. Other things being equal, vibration is more intense and damage is greater in areas of thick unconsolidated deposits than in areas of bedrock. Examples of this rule are too numerous and too widely recognized to need documentation. All the severe vibratory damage in the Anchorage Lowland was in areas underlain at some depth by Bootlegger Cove Clay. Design and construction practices, on the other hand, are critical nongeologic factors that influence building performance and vibratory damage. As pointed out by Berg and Stratta (1964, p. 58), structural weaknesses are quickly ferreted out by earthquakes.

Inasmuch as this report emphasizes primarily the geologic effects of the March 27 earthquake in the Anchorage area, only the highlights of the vibratory effects are summarized here. Detailed analyses of structural response to vibratory stress have been presented by several authors (Berg and Stratta, 1964; McMinn, 1964; National Board of Fire Underwriters and Pacific Fire Rating Bureau, 1964; Steinbrugge, 1964), and the following summary is based largely on their findings.

The distribution and character of direct seismic damage were unusual to say the least. Probably few buildings in the Anchorage area were totally undamaged, but many blocks of homes and small commercial buildings received cursory damage at most and sustained virtually no damage to structural members, frameworks, or foundations. Many chimneys toppled, and quite a few fireplaces separated from adjoining exterior walls. Indoors, plaster cracked and objects were thrown to the floors of many homes. On many buildings however, fragile decorative facings, pillars, and cornices were unscathed. Church steeples remained standing. Headstones in cemeteries—long used as indicators of earthquake intensity—stood untoppled throughout the area. Large elevated well-guyed water tanks were undamaged (fig. 6), although trees whiplashing back and forth touched the ground on either side according to eyewitnesses. On the other hand, multistory buildings and buildings having large floor areas commonly sustained significant structural damage; several such buildings were total losses (figs. 5 and 13), and many required major repairs.

Thus, direct seismic damage was highly selective. Aside from variations in design, construction practice, and workmanship, large buildings were more severely damaged than small ones. Inertia was a factor, of course; other things being equal, heavy structures are more susceptible to vibratory damage than light ones.

Steinbrugge (1964, p. 58–72) has described vibratory damage effects in the Anchorage area in terms of engineering seismology. He ascribed (p. 59) the selectivity of the effects to the great magnitude of the quake and the distance of Anchorage from the epicenter: "The ground motion at Anchorage did not appear to contain significant short period motion which has been commonly observed in epicentral regions of destructive shocks." He added (p. 71): "The earthquake damage in Anchorage was selective * * *. This damage pattern appears to be attributable to the distance that Anchorage was from the epicenter, with the longer period ground motion having a dominant effect at this distance." According to White (1965, p. 91), attenuation of sinusoidal seismic waves at low frequencies should vary as the square of the frequency. Where seismic vibrations have long periods and large amplitudes, distortion and strain are far more severe in large structures than in small ones. The reason for the selectivity of the March 27 earthquake at Anchorage is thus apparent. The prolonged duration of the shaking must also have heightened all other effects; as mentioned before, strong motion lasted 3 to 4 minutes, possibly longer.

In general, most buildings outside of landslide areas withstood well the effects of the earthquake. The low loss of life is, in some measure, a tribute to the structural soundness of most of the larger buildings but in part also is the result of the low susceptibility of smaller structures to the long-period vibrations that racked the area. All the high-rise structures in Anchorage (10 stories or more) sustained moderate to heavy damage without collapse or loss of life. Four structures of medium height—four to nine stories—collapsed or were damaged beyond repair. Several low buildings of less than four stories were totally destroyed, but most such structures were damaged little or not at all.

In the general downtown area, outside the Fourth Avenue and L Street slides, a 5-story department store (J. C. Penney Co.) was a total loss, 4 buildings 1 or 2 stories high were heavily damaged, 9 buildings 1 to 14 stories high sustained moderate damage, and about 22 buildings received light damage (National Board of Fire

Underwriters and Pacific Fire Rating Bureau, 1964, p. 26). In the same area, however, many single family dwellings, stores, and small commercial buildings were damaged little if at all. Steinbrugge (1964, p. 71) found no evidence to indicate that one construction material was superior to another, given comparable attention to design and construction.

A few buildings seem to have been designed without regard to earthquake stresses. The Hillside Apartments and various warehouses that collapsed were examples. In other buildings, inadequate connections between structural parts were the most common causes of failure; improperly welded joints, inadequate ties in reinforced-concrete members, and improperly spliced reinforcing rods all were loci of failure. Welded precast- and prestressed-concrete structural members seemed to be particularly susceptible to joint failure, although some buildings so framed were undamaged. In some buildings, construction joints were inadequately keyed. Poured nonmonolithic concrete joints were sites of shear failures, and some concrete appeared to be substandard. Most poured-in-place concrete structures, however, fared well. In some multistory buildings, provisions for vertical shear were inadequate to withstand a stress of the intensity generated by the March 27 earthquake (Berg and Stratta, 1964, p. 58).

Small wood-frame buildings outside areas of ground displacement were generally little damaged (Steinbrugge, 1964, p. 63–64). Unreinforced masonry walls and chimneys were usually intact, and most interior wooden stud walls sustained nothing more than minor nonstructural cracking. Foundations of poured concrete or hollow concrete block (some apparently unreinforced) generally were intact. Usually windows were unbroken, and objects remained on shelves indoors.

Of the many buildings damaged or destroyed by direct seismic vibration in Anchorage, several structures described briefly below are most significant for the damage or lack of damage they sustained, relative to design and construction practice. These structures, therefore, have received the most attention from investigators. The following synopsis is based mostly on the reports of Berg and Stratta, Steinbrugge, and the National Board of Fire Underwriters and Pacific Fire Rating Bureau.

ALASKA PSYCHIATRIC INSTITUTE

The three-story steel-framed Alaska Psychiatric Institute, southeast of the main part of Anchorage, is notable because it sustained so little damage. There were minor cracks in stair wells and broken pipe hangers and machinery mounts in the penthouse. The nearby new Providence Hospital and the Alaska Methodist University also were little damaged. Significantly, perhaps, these three buildings are outside the area underlain by Bootlegger Cove Clay.

THE ALASKA RAILROAD MARSHALLING YARDS

Several buildings in the Alaska Railroad marshalling yards were damaged. Warehouses collapsed and shops were slightly damaged. The steel-framed wheel-shop building partly collapsed and has since been torn down. Some buildings were in or near the toe area of the Fourth Avenue landslide and, hence, may have been subjected to ground displacements as well as to seismic vibration.

ALASKA SALES AND SERVICE BUILDING

The one-story Alaska Sales and Service Building on East Fifth Avenue at Medfra Street was under construction but was structurally almost complete at the time of the earthquake; it was a total loss. Collapse is attributed chiefly to failure of welded connections between T-shaped precast-concrete columns and roof beams, caused either by the breaking of welds or tearing out of bar inserts. The exterior precast-concrete walls of the building partly collapsed when the roof gave way.

ANCHORAGE INTERNATIONAL AIRPORT

The Anchorage International Airport control tower, a reinforced-concrete structure, collapsed to the ground (fig. 7), killing one occupant and injuring another and damaging the connecting walls of the adjacent terminal building. The terminal building was otherwise little damaged. At the airport post office building, a rear wall pulled away from the roof trusses and leaned outward; moderate nonstructural damage was sustained indoors.

ANCHORAGE WESTWARD HOTEL

The steel-framed Anchorage Westward Hotel complex is just west of the Fourth Avenue landslide area facing Third Avenue between E and F Streets and is within the area of peripheral cracking. Several cracks passed through the building foundations and first floors. Some cracks had vertical offsets. There was little visible damage to the flexible exterior metal skin of the 14-story tower, but there was significant structural damage (since repaired) to the more rigid interior structural elements. Reinforced-concrete columns were buckled, and reinforced-concrete shear walls were damaged. An interior east-west shear wall had failures above all door openings in every story. At the eight-story level there was shear-wall damage along a horizontal joint between older and newer sections of the building. Considerable damage was caused by pounding between the 14-story tower and the adjacent 3- and 6-story wings.

CORDOVA BUILDING

The Cordova Building is a six-story structure at the northeast corner of Sixth Avenue and Cordova Street. The damage was mostly light and was caused chiefly by an east-west movement of the building. This direction is parallel to the shorter dimension of the floor plan and also is the direction of greatest strength in the steel support columns (Berg and Stratta, 1964, p. 37). Damage was mainly in the first story. The southeast-corner support column failed below the second-floor beam, and the exterior reinforced-concrete curtain walls sheared at the top of the basement; the corner broke away at the first and second stories. The center column in the south wall buckled. The reinforced-concrete stair and elevator shaft sheared at the base of the first story. The penthouse collapsed. Repairs have since been completed.

ELMENDORF AIR FORCE BASE

At Elmendorf Air Force Base the reinforced-concrete control tower was damaged by cracks extending from its base to a height of about 15 feet. A warehouse (No. 21–884) partly collapsed when roof-beam anchor bolts pulled away from poured-in-place concrete piers. The Elmendorf Field House was slightly damaged at the roof corners and at the junctions of the side walls with the steel frame. Elmendorf Hospital was damaged by X-shaped shear fractures in block-panel walls at the second- through the fifth-story levels, by interior failures at a few reinforced-concrete columns and at the elevator shaft, and by some movement between different sections of the building. Damage has since been repaired, reportedly at a cost of more than $1 million (R. M. Waller, written commun., 1965).

FIFTH AVENUE CHRYSLER CENTER

The one-story Fifth Avenue Chrysler Center just north of Merrill Field was a total loss. It had a precast- and prestressed-concrete T-beam roof supported by concrete-block walls. The front of the building collapsed and the T-beam roof fell in on the showroom. The hollow-core concrete-block side walls failed at the rear corners of the building.

FIRST FEDERAL SAVINGS AND LOAN BUILDING

The three-story First Federal Savings and Loan Building is located at the northwest corner of Fifth Avenue and C Street. Ground cracks opened parallel to its east and south walls evidently in back fill. Even though the east and south faces of the building are mainly glass, there was little glass breakage. The west and north walls and the brick panels in the east wall sheared horizontally through the second story. The building has been repaired.

FOUR SEASONS APARTMENT BUILDING

The Four Seasons Apartment Building at West Ninth Avenue and M Street was a spectacular total loss (fig. 13). It was designed for year-round luxury living, but didn't survive its first season. The building was nearing completion at the time of the earthquake, but fortunately it was not yet occupied—it collapsed in a pile of rubble. It was a six-story lift-slab reinforced-concrete building with two central poured-in-place cores, one for a stairwell and one for an elevator. The floor slabs were torn loose from shear heads on the structural-steel columns and pulled away from keyways in the stairwell and elevator cores. Some dowels connecting the slabs to the cores broke; others were pulled free.

The collapse of the Four Seasons Apartment Building was witnessed from 1000 Tenth Avenue by Glen L. Faulkner, a consulting geologist, who reported (oral commun., 1964) that the building collapsed just before the end of the quake, after shaking violently for perhaps 2 to 3 minutes. Just before it fell, it seemed to start crumbling near the second-floor level in the area of its northeast corner. Then with a slight tilt northward it collapsed vertically in a great cloud of dust. The two central cores were left leaning sharply northward. The steel support columns fell to the north also, and the slab floors were stacked one on another like pancakes.

The effects of seismic vibration in areas of cracked ground cannot always be separated from the effects of ground displacement. The collapse of the Four Seasons Apartment Building (fig. 13), a case in point, has been attributed to seismic vibration. Perhaps vibration was the dominant factor in the building's failure. The building, however, was close to the L Street graben at West Ninth Avenue and M Street, and a large subsidiary fracture, trending about west-southwest, passed directly under the building. In the chaotic pile of rubble left after the

13.—Wreckage of six-story Four Seasons Apartment Building, Anchorage, Alaska. Canted elevator shaft at center. Large crack in foreground, filled in on M Street to restore traffic, passed beneath building.

quake, the fracture may have been overlooked by early investigators. Furthermore, part of the fracture was quickly bulldozed over to restore traffic along M Street. In any event, the fracture was mapped by the Engineering Geology Evaluation Group, and its extensions east and west were still well preserved 2 months or more after the quake. Near the building the fracture had a vertical displacement of about 2 feet; the north side dropped down—a displacement that few manmade structures could be expected to withstand. The actual displacement beneath the building, however, is uncertain. The fact that the elevator shaft and stair well of the building fell northward suggests that they tilted in that direction in response to the northward displacement along the fracture. Support pillars collapsed in the same direction.

The building reportedly withstood most of the earthquake, although it shook violently, but it collapsed suddenly shortly before the end of the quake. Eyewitness accounts of the L Street slide (Shannon and Wilson, Inc., 1964, p. 53) indicate that sliding there occurred in the latter part of the quake. The timing seems more than coincidental; the crack beneath the Four Seasons Apartment Building must have attained offset at about the same time. It could not have had offset until the slide itself had begun.

HILL BUILDING

The Hill Building, an eight-story office building at the southeast corner of Sixth Avenue and G Street, was slightly damaged. There apparently was no damage to the steel frame, and there was little exterior damage. A reinforced-concrete column failed at its junction with the canopy roof of the service entrance. The center cores, which were intended to resist horizontal forces, dropped as much as 5 inches at one point and 3 inches at another where they were partly shattered at their connections with the footings. The core walls were cracked also, and the beams between the cores were sheared by vertical stresses.

HILLSIDE APARTMENTS

The Hillside Apartments were on the south side of Sixteenth Avenue between G and H Streets on a bluff overlooking Chester Creek. They were damaged beyond repair and have since been dismantled. This was a split-level building, five stories high on the south side and three stories high on the street side. It had a post-and-lintel frame with steel-pipe columns, rolled-steel beams and concrete floor slabs on steel joists.

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Walls were unreinforced hollow concrete block. The building was sheared in an east-west direction at the third-story level on the south side and in the lower two stories on the north side—the upper stories lurched west relative to the lower stories. Seemingly, no provision had been made for resistance to strong lateral seismic stress.

KNIK ARMS APARTMENT BUILDING

The six-story Knik Arms Apartment Building, of poured-in-place reinforced-concrete framing, is on the west side of L Street between Sixth and Seventh Avenues, on the L Street landslide block between the headward graben and the bluff overlooking Knik Arm. It is especially noteworthy for its apparent lack of damage, despite a lateral displacement of about 10 feet west-northwest as it was carried along with the underlying slide block. Many smaller residential buildings on the slide block were also undamaged or little damaged.

MOUNT McKINLEY BUILDING AND 1200 L STREET APARTMENT BUILDING

The Mount McKinley Building and the 1200 L Street Apartment Building are twin 14-story reinforced-concrete apartment buildings about a mile apart and facing in opposite directions. They sustained similar damage, almost matching crack for crack, although damage in general was somewhat more severe in the Mount McKinley Building. The Mount McKinley Building is on Denali Street between Third and Fourth Avenues; the 1200 L Street Apartment Building is across town near Inlet View School.

The most obvious damage to both buildings was X-shaped shear cracks in spandrel beams, caused by vertical shear between exterior support piers as a result of lateral swaying. Spandrels in general were most heavily damaged in the middle third of the floor levels. Interior walls were less damaged than exterior ones.

In the Mount McKinley Building, vertical piers sheared horizontally at the third-story level on the north side of the building (at a construction joint) and at the second story on the south side. An exterior column fractured diagonally at the first-story level. A television-antenna tower on top of the building was undamaged.

Failures similar to those in the Mount McKinley Building occurred in the 1200 L Street Apartment Building in piers at the south face of the building at the second- and third-story levels. Corner spandrels were more heavily damaged than those in the Mount McKinley Building.

An account of the earthquake as related by an occupant of the 12th story of the 1200 L Street Apartment Building was provided by William G. Binkley (written commun., 1964). Mr. Binkley, a geologist, noted wryly that he was living in a sort of oversized seismograph. The quake was first felt in the building as a light tremor that intensified rapidly until objects began to fall to the floor. About 30 seconds of trembling motion was followed by perhaps 2 minutes of violent jarring, in which the building seemed to sway 8 to 10 feet horizontally and 1 to 2 feet vertically. Bookcases were overturned, and fallen books and furniture were thrown back and forth across the room. Violent shaking was accompanied by a deep rumble and by higher pitched sounds of shattering plaster and falling dishes and furniture. In the kitchen, everything from the cabinets and refrigerator crashed to the floor, where it was "churned into a melange of broken dishes and glass, catsup and syrup, flour, beans, pots and pans, eggs, lettuce, and pickles." Mr. Binkley was able to crawl from the living room to a hall where he braced his feet against one wall and his back against another. Then the building stopped "jumping," the noise stopped abruptly, and the motion of the building diminished gradually with a subsiding, weaving tremble.

Most people at ground level experienced far less violent shaking than those in tall buildings. Many individuals were unaware of the catastrophic proportions of the quake until much later when the reports began to come in. In the heights of the 1200 L Street Apartments, however, the accelerations undoubtedly were greatly magnified.

PENNEY'S DEPARTMENT STORE BUILDING

The Penney's Department Store Building (fig. 5) was a total loss and has since been dismantled. Photographs of its wreckage have been widely distributed in magazines and newspapers. Penney's was a five-story reinforced-concrete structure having shear walls on three sides and a curtain wall of precast panels on the north. Failure is attributed to torsion caused by rotational displacements, in turn caused by an eccentric position of the center of rigidity of the structure. This rotational motion sheared off the west support wall at the second-story level, causing the wall and all overlying floors to collapse. The floor slabs also sheared at their connections to the next adjacent column wall to the east. The northeast corner of the building collapsed (fig. 4), and most of the precast panels on the north face fell to the street. Berg and Stratta (1964, p. 35) reconstructed the sequence of events as follows: The northeast corner of the building collapsed following failure of the east shear wall at that point; the west shear wall then failed at the second-story level at the north end of the building. The east shear wall failed at the south end of the building, and the precast curtain-wall panels collapsed at the north face of the building.

The north-facing curtain wall must have fallen rather late in the quake. Some motorists reportedly were able to start their parked cars and move them away from below the wall before it fell. One motorist who ran to her car in an attempt to move it, however, was trapped and killed (R. M. Waller, written commun., 1965).

PORT OF ANCHORAGE AREA

Much of the damage in the Port of Anchorage area was caused by ground displacements along fractures, but some damage is attributable to direct seismic shaking. The main pier lurched laterally 5 to 19 inches. Large longitudinal cracks and several transverse ones opened up, and the walls of several buildings were cracked. All four gantry cranes were damaged. Steel piles penetrated the deck of a subordinate pier. Approach roads and railroads settled as much as 18 inches. Two cement-storage tanks were toppled, one at the property of the Permanente Cement Co. at the entrance to the U.S. Army Dock and one at the Alaska Aggregate Corp. facility just north of Ship Creek. Oil-storage tanks in the dock area were mostly superficially damaged, but some tanks were bulged outward at the bottom, probably by rocking and pounding back and forth as the contents sloshed to and fro.

WEST ANCHORAGE HIGH SCHOOL

West Anchorage High School is on the south side of Hillcrest Drive a few blocks west of Spenard Road and just south of a bluff overlooking Chester Creek. Structurally separate parts of the school building reacted differently to the vibrations. The two-story classroom section of the building was heavily damaged, especially the second story. Exterior columns failed at connections with the roof and with the second-floor spandrels. Extensive damage was caused by pounding between the gymnasium section and the classroom section. There was extensive ground cracking along the bluff above Chester Creek and a small rotational slump formed due north of the school.

GROUND DISPLACEMENTS OTHER THAN LANDSLIDES

GROUND CRACKS AND COMPACTION

The distribution of ground cracks in the Anchorage area (generalized in fig. 1) was mapped by the Engineering Geology Evaluation Group (1964, pl. 1) shortly after the earthquake. Most of the observations on which this map is based were made on traverses adjacent to streets and highways but some were obtained from a scrutiny of aerial photographs. Many unobserved cracks probably formed in the less-accessible snow-covered undeveloped areas. Frozen muskeg in and bordering swamps, for example, was very susceptible to cracking. The map, nevertheless, shows clearly the widespread distribution of cracking, and it emphasizes the high susceptibility to cracking of lowland areas underlain by silty clay and outwash, as opposed to highland areas underlain by ground moraine. Made land in former muskeg areas, reclaimed either by draining or by filling directly over muskeg, was also very susceptible to cracking.

Reportedly, many ground cracks opened and closed with the rhythm of the earthquake. Such action may help explain the extensive damage some cracks caused. A pulsating fracture would cause more damage to a superencumbent structure than a fracture that merely opened.

The most severely cracked ground was adjacent to landslides (fig. 40) where cracks were caused by tension directly related to sliding. The cracks in turn caused much structural damage in built-up areas. Some cracked ground undoubtedly would have developed into landslides had the earthquake lasted longer, as for example at Turnagain Heights, back of the landslide (pl. 1). Here, the pattern of fracturing was concentric to the head of the slide, and many crescentic tension cracks extended as far as 2,200 feet from the slide proper, to the vicinity of Northern Lights Boulevard. In the downtown area, many streets and buildings were damaged by cracks behind the periphery of the Fourth Avenue slide. Major cracking and ground adjustments extended a city block or more back from the slide. In contrast, the L Street and Government Hill slides broke away clean; they were much cracked themselves, but few cracks extended behind them.

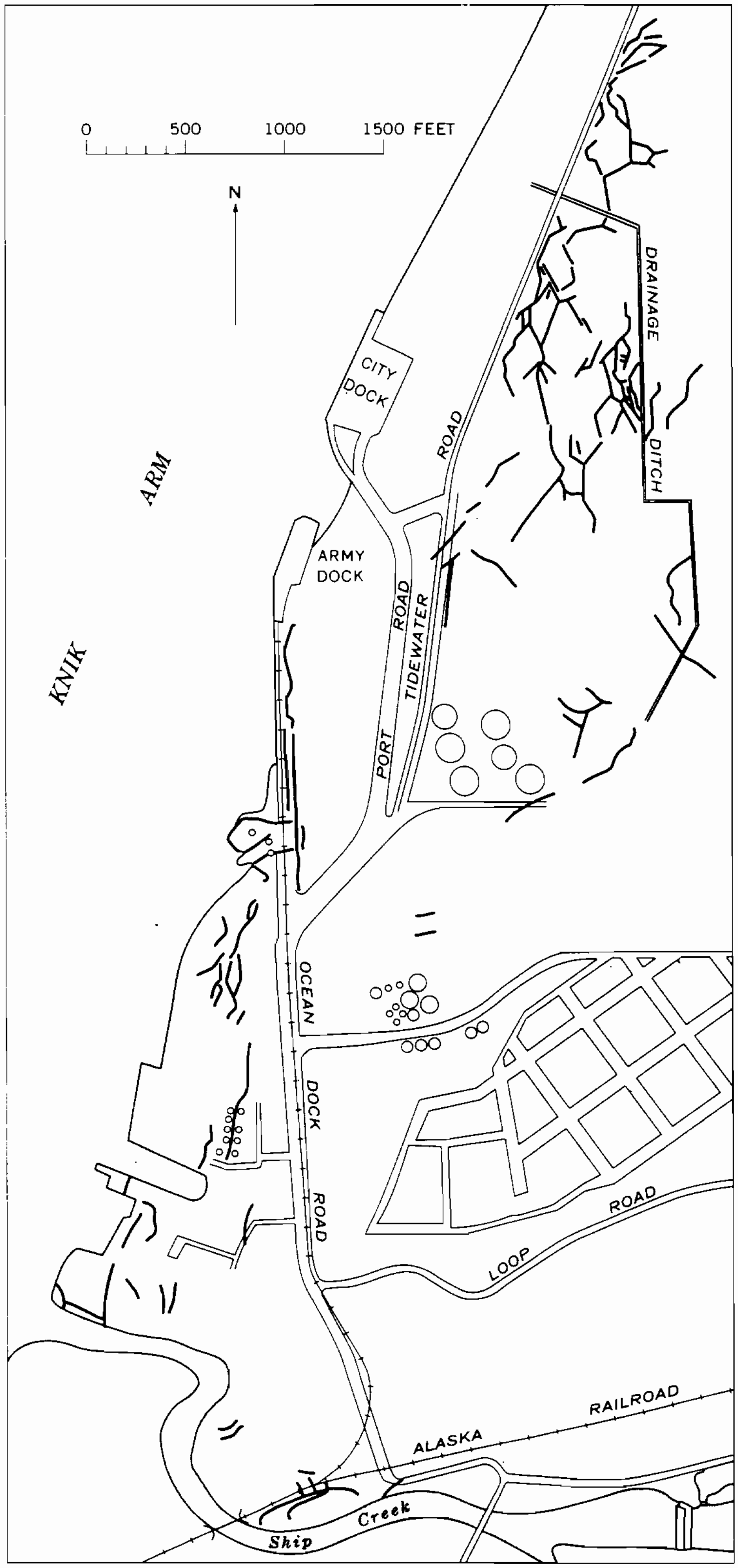
Preexisting zones of weakness in the ground were particularly susceptible to cracking. Some cracks followed backfilled utility trenches, for example, or backfills around building foundations. Old frost cracks in pavement—caused by contraction at low temperatures—were reopened by the earthquake; at Turnagain Heights, the pattern of cracking was very striking—ground cracks veered along concrete curbings, crossed streets at right angles along old contraction cracks, and then resumed direction on the other side. In filled areas, cracks commonly formed near the contact between filled and preexisting ground. Some buildings unfortunately constructed partly on fill were sundered by cracking between the fill and the adjacent ground.

Trees were split by cracks that passed through their root systems (fig. 40), their roots being held fast by the frozen ground.

On sloping ground, cracks generally formed parallel to the slope, owing to differential compaction, lateral shifting (lurching) under the influence of gravity, tension, and rupture.

Alternate cuts and fills along streets, highways, and railroads demonstrated a predictable pattern of fractures. Fills that compacted under the vibratory stress of the quake were cracked adjacent to the fill-cut contact. Many cracks of this sort crossed the Seward-Anchorage Highway, some fills dropping several inches. One in particular formed a very sharp pavement break on the north approach to the Rabbit Creek crossing near the south margin of the Anchorage Lowland. Similarly, an Alaska Railroad embankment across Fish Creek east of Turnagain Heights cracked to pieces and collapsed.

Extensive ground cracks formed in the Port of Anchorage area on the flats extending north from the mouth of Ship Creek to and beyond City Dock between the bluff on the east and the shore of Knik



14.—Ground cracks, Port of Anchorage and vicinity. Anchorage, Alaska. Many of these cracks spouted mud, particularly those east of City Dock. Base by city of Anchorage, Office of City Engineer.

Arm on the west (fig. 14). This area is underlain by estuarine silt, peat, muskeg, and artificial fill (Engineering Geology Evaluation Group, 1964, p. 22), all highly susceptible to cracking under the stresses of prolonged vibration and compaction. Great quantities of mud were extruded from large polygonal fractures in the flats northeast of City Dock. To the south, longitudinal cracks 300 feet or more long extended the full length of the Army Dock embankment. At the south end of the embankment, in an area of transverse cracks, a steel cement-storage tank was overturned. Collapse of the tank, however, may have been caused by inadequate anchoring of base plates to support columns (Berg and Stratta, 1964, p. 46).

Facilities south of the Army Dock to the west of Ocean Dock Road were also damaged by ground cracks. Cracks extended through the tank farm of the Union Oil Co. of California and damaged the adjacent Alaska Fish and Farm Products lease and the Cook Inlet Tug and Barge lease. There was ground cracking in the dock area at the mouth of Ship Creek, where another cement storage tank toppled and where the right abutment of the adjacent Alaska Railroad bridge across Ship Creek was damaged by subsidence and cracking.

Incipient and minor slumping along the bluff east of the dock area also produced many cracks, some of them large. All these, however, were in undeveloped woodland areas, mostly on military land.

15.—Part of a sand boil, Turnagain Heights slide area. Ridges 2 to 3 feet high and 100 feet long or more were formed as fountains of water were ejected through frozen outwash.

SAND BOILS AND MUD FOUNTAINS

Sand boils and mud fountains are transient or short-lived features commonly produced by strong earthquakes where ground breakage occurs. They were wide spread in the damage zone of the March 27 Alaskan earthquake (Grantz, Plafker, and Kachadoorian, 1964, p. 6). At Anchorage they have been reported at several localities and probably occurred in many others.

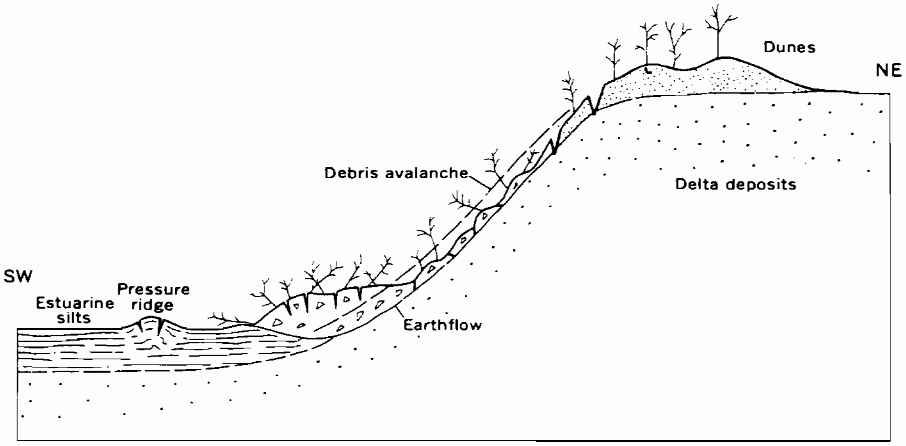
Mud fountains were produced where the water table was shallow and frozen ground overlying saturated, unconsolidated sand or silt was cracked by the earthquake. Intense shaking possibly accompanied by settlement and compaction forced muddy water out through the cracks.

At the Port of Anchorage, mud was ejected from many cracks east and northeast of the City Dock and from a few cracks between the Army Dock and Ship Creek. The ground was snow covered at the time, and consequently the ejected mud shows plainly on aerial photographs taken shortly after the earthquake.

A mile above the mouth of Chester Creek, where Spenard Road crosses it, large mud fountains were reported by a motorist whose passage was blocked by cracks in the road. The fountains reached higher than his car (R. D. Miller, oral commun., 1964).

At Turnagain Heights, large sand boils, unlike any others in the area, resulted from agitation and settlement of slump blocks within the landslide (fig. 15). These boils built ridges of sand 2 to 3 feet high, 3 to 6 feet wide, and about 100 feet long. Some ridges had hollow cores. The sand generally was finer grained than the subjacent outwash, from which it must have come, and it may, therefore, have been sorted by natural elutriation in a pulsating water column.

A large kidney-shaped boil near Turnagain Heights, but outside the landslide area, spread over an area of about 3,200 square feet. This boil was just west of a collapsed Alaska Railroad embankment across Fish Creek one-fourth mile north of Northern Lights Boulevard (Engineering Geology



16.—Diagrammatic section near Point Campbell, showing mode of failure along bluff line overlooking Turnagain Arm.

Evaluation Group, 1964, pl. 8). Two other large boils poured out of the opposite side of the same embankment. They may have been caused by spontaneous liquefaction of subjacent silty sand, that was pumped to the surface when the embankment fill collapsed. Several similar but smaller boils formed in the downtown area in the toe of the Fourth Avenue slide, east and just below the intersection of E Street and West Second Avenue (Engineering Geology Evaluation Group, 1964, pl. 5).

Most of the lakes in the Anchorage area—solidly frozen over at the time of the quake—developed distinctive patterns of peripheral ice cracks extending generally 50 to 100 feet out from shore but in places much farther. The fracture zones in the larger lakes were wider than those in the smaller ones, and some small lakes had no cracks at all. Most of the cracks ejected fountains of water during the quake, and where the bottom was shallow enough they ejected mud, particularly in toward shore. Large mud fountains formed at Lake Otis (on the northeast shore), Lake Spenard, Hood Lake, and Connors Lake, to name a few.

The earth dam impounding Campbell Lake, near Turnagain Arm, broke during the earthquake. As the water level fell, ice on the lake broke into blocks, and fountains of muddy water squirted up through the cracks. Reportedly, these fountains reached heights of 20 feet or more.

LANDSLIDES

DEBRIS SLIDES, AVALANCHES, AND ROTATIONAL SLIDES

Landslides precipitated along the bluffs of Knik and Turnagain Arms and along the benches above Ship Creek were the most spectacular and awesome manifestations of earthquake damage in the area. The slides took several forms and occurred in several types of earth material. The most destructive slides, in terms of property damage, resulted from failures in the Bootlegger Cove Clay. Serious failures, however, also occurred in glacial till, delta deposits, and dune sands—the only criteria being unconsolidated surficial deposits having insufficient shear strength to resist the accelerations of the quake in potentially unstable topographic settings. A landslide occurs where the ratio of shearing resistance of the ground to the shearing stress on the potential slide surface decreases from a value greater than one to unity. An earthquake can cause an almost instantaneous decrease in this ratio (Terzaghi, 1950, p. 110).

SLIDES ALONG TURNAGAIN ARM

Structurally, the simplest slides were in the bluffs facing Turnagain Arm southeast of Point Campbell (fig. 16). There, a thin cover of wind-blown sand and slope wash had been loosely anchored to the steep face of the bluff by an overgrowth of trees, shrubs, and grasses. The entire superficial mat, probably frozen at the time, slumped downward a few inches to several feet along the full length of the bluff from Point Campbell to Campbell Creek, a distance of about 4 miles. Locally the slumping was more intense, and the face of the bluff was laid bare from top to bottom. Southeast of Campbell Creek, minor or incipient slumping and cracking extended along the bluff to the tracks of The Alaska Railroad near Potter, where slides destroyed trackage and embankments.

Between Point Campbell and Campbell Creek, the earth slid either in a mass, as a shallow nonrotational glide, or it disintegrated into blocks and fragments, as a debris slide or avalanche (Varnes, 1958, pl. 1). All intermediate steps are represented. The velocity of motion has not been ascertained, but it probably was rapid because of the height and steepness of the slope and the granular incoherent character of the material.

Trees carried along in the slides were tilted at all angles of disarray, many were rotated outward, some inward, and some were overridden by surging at the base of the slope. In some places the debris surged out into lobate forms on the estuarine muds and swamps at the foot of the bluff, and in many places one, two, or three pressure ridges were formed at the foot of the slope or out on the flats. Some ridges were more than a thousand feet from the bluff. The flats below the bluff were snow covered at the time and probably frozen. Pressure probably was transmitted directly from the slide through the frozen layer to the points of failure on the flats.

The high, steep part of the bluff nearest Point Campbell consists chiefly of loose deltaic sand and gravel veneered with dune sand. The bluff supported little or no vegetation prior to the earthquake because of more or less continuous buffeting by storm waves. It had, therefore, practically no cohesion with the slope and failed chiefly by avalanching.

Many gaping cracks remained along the edge of the bluff between Point Campbell and Campbell Creek after the temblor ceased. The sliding, however, was entirely superficial and did little to alter the firmness of the land surface back from the bluff. Nevertheless, by partial removal of the stabilizing vegetative mantle, the vulnerability of the slope to continued sapping has been heightened, and intermittent sluffing will probably take place for a long time to come. Slow, gradual backwasting, in fact, should be regarded as normal.

What effect tectonic subsidence—about 2 feet in the Anchorage area—will have on the stability of the bluff line is still undetermined. Certainly the part of the bluff near Point Campbell will be even more exposed to the attacks of tide and wave than it was before the quake. Other points along the bluff farther southeast probably will be attacked by occasional storm waves also, especially when strong winds coincide with highest tides. It seems likely that, wherever the foot of the bluff is accessible to storm waves, new slope profiles will have to be established before equilibrium can be fully regained.

POTTER HILL SLIDES

Destructive slides carried away several hundred feet of track and right-of-way along The Alaska Railroad between mile posts 103 and 104 about 2½ miles northeast of Potter (fig. 17). These slides, known as the Potter Hill slides, were mapped and examined in detail after the earthquake, on April 20, 1964, by D. S. McCulloch and M. G. Bonilla of the U.S. Geological Survey on behalf of The Alaska Railroad. The following summary is based on their findings.

The Potter Hill area has a history of landsliding that extends back at least 35 years. Bert Wennerstrom, chief accountant of The Alaska Railroad recalled that "in the late twenties and early thirties, heavy rains caused sliding along this section (mile 103–104). The largest slide took out about 1,000 feet of track." Wennerstrom remembered that the failures involved cuts in the natural bank material (McCulloch and Bonilla, written commun., 1964).

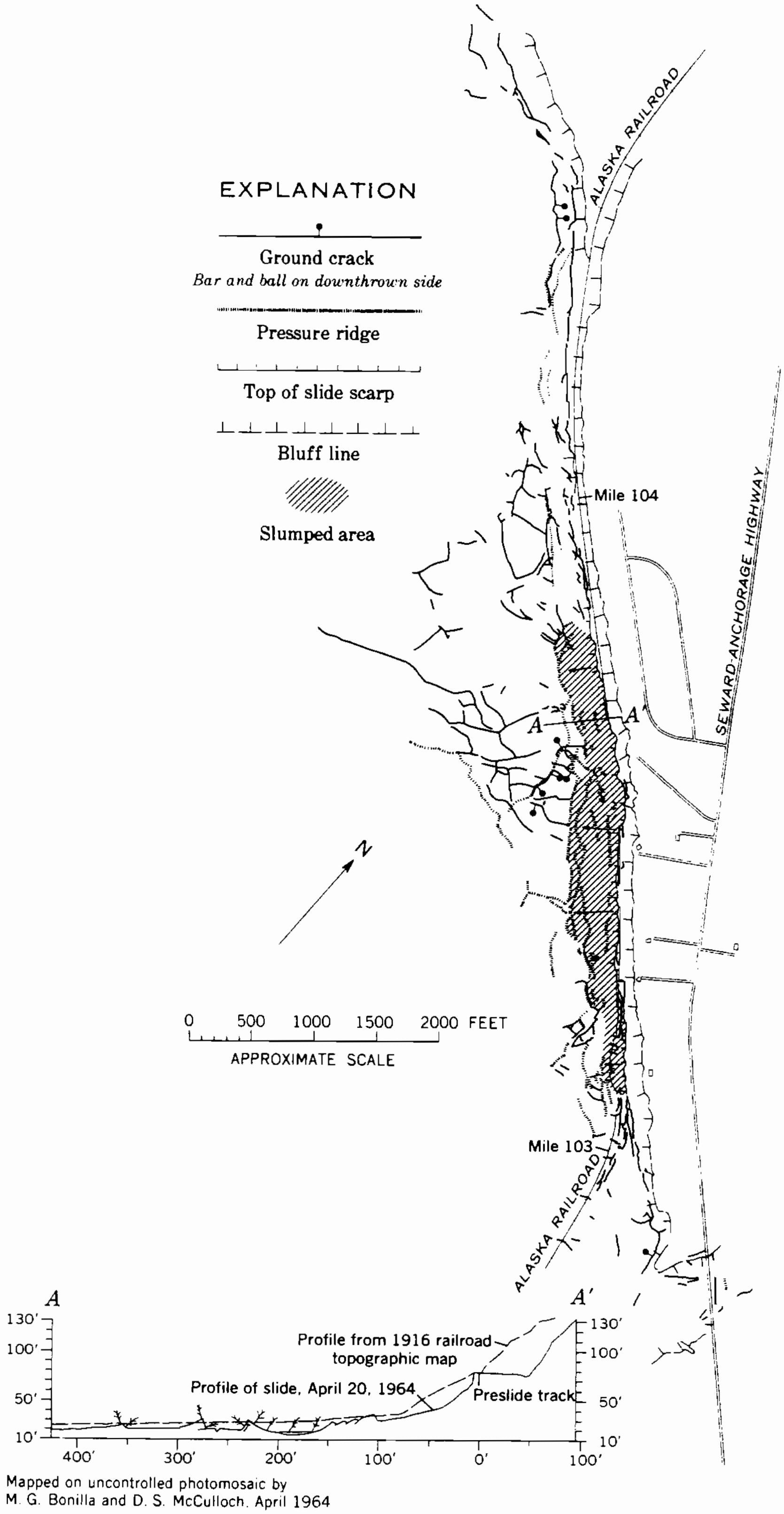
On October 3, 1954, landslides caused by an earthquake again destroyed trackage in the same area. One slump left 140 feet of track suspended 15 to 20 feet in the air (Miller and Dobrovolny, 1959, p. 105). Although it is not clear just what materials were involved, failure probably occurred in natural bank material as well as in fill.

The slides of March 27, 1964, affected cuts in both natural bank material and fill. The ground was frozen at the time of the earthquake and, presumably, there was little surface runoff. Sand boils, however, indicate that the material in the slides had a high water content.

Several kinds of material were involved. Glacial till forms the top of the bluff back of the slides and most of the exposed face of the bluff in the northern part of the slide area. The till rests on outwash with a contact that dips gently north. The outwash in turn lies on a sequence of blue clay, silt, and fine sand, which perhaps is equivalent to the Bootlegger Cove Clay. This sequence forms the lower part of the slope. At the south end of the slide area, a coarse gravel overlies the till. The materials in the bluff abut against and probably pass beneath intertidal silts in the flats at the base of the bluff.

The bluff is saturated at its base, and, at several levels above its base, ground water escapes laterally through permeable beds in the outwash. In wells east of the slide area, ground-water levels reportedly are 30 to 240 feet above sea level. During the earthquake, sand spouts issued from cracks 400 feet east of the edge of the bluff.

All the slides along Potter Hill had a similar form. They con-



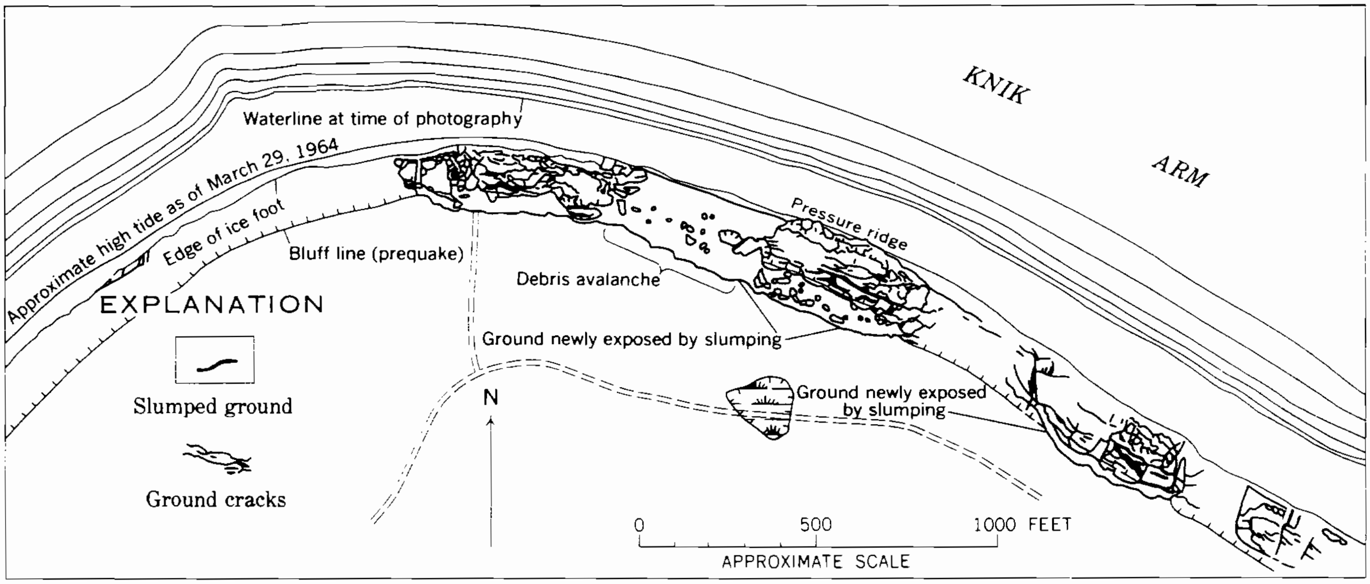
17.—**Map and profile of Potter Hill landslide area near Anchorage, Alaska.**

sisted of elongate fragmented slump blocks rotated backward and broken into many pieces toward the base of the slope. In places at the toe, they turned into earth and mudflows derived partly from the intertidal silt of Turnagain Arm and partly, perhaps, from the clay-silt-sand sequence in the lower part of the bluff.

Many pressure ridges formed on the flats below the slides; some smaller ones were as far as a third of a mile beyond the toe, but most of the larger ones were within 400 feet. They were formed by compressive stresses probably transmitted horizontally from the toe of the slides through the frozen upper layers of the estuarine silt.

Marginal fractures bounded the slides roughly parallel to the face of the bluff. They cut across natural bank material and fill alike as high as the level of the tracks, in the roadbed itself, and in the drainage ditch beside the tracks. Near Rabbit Creek, where the tracks turn southward across the flats, cracks reached nearly to the top of the bluff. At one point they extended as far back as 400 feet from the edge of the bluff. Sand that was ejected from these cracks during the quake spread out onto the surface of the snow. But in general, cracking was confined to the face of the bluff itself.

The mechanics of sliding has been investigated by D. S. McCulloch and M. G. Bonilla (written commun., 1964). They concluded that sliding was initiated by failure and flowage of material from the base of the slope. Higher parts of the slope slumped and distintegrated as a consequence. Flowage may have occurred in the modern estuarine silt at the base



18.—Photogeologic sketch map of Point Woronzof landslide area, Alaska.

of the slope, which carried part of the weight of the roadbed, or in the fine silt and sand in the lower part of the bluff. Perhaps it occurred in both.

Measurements by McCulloch and Bonilla indicate that the volume of the slide material that accumulated at the base of the slope and on the adjacent flats was less than that of the material removed from the slope by sliding. They concluded, therefore, that this difference in volume was compensated partly by flowage in the estuarine silt and partly by lateral translation through the silt to pressure ridges on the flats.

SLIDES AT POINT WORONZOF

Point Woronzof is underlain by loose unconsolidated sand and gravel. Because it projects north into Knik Arm, the point is exposed continuously to the ebb and flow of the strong tidal currents and to the pounding of storm waves propelled across the arm by northerly winds sweeping down across the Susitna Lowland. The maximum tidal range at Anchorage is about 38 feet. Point Woronzof, therefore, has been subjected to more or less continuous and vigorous shoreline erosion, and, since 1909 at least, has retreated southward at a mean rate of about 2 feet per year (Miller and Dobrovolny, 1959, p. 89).

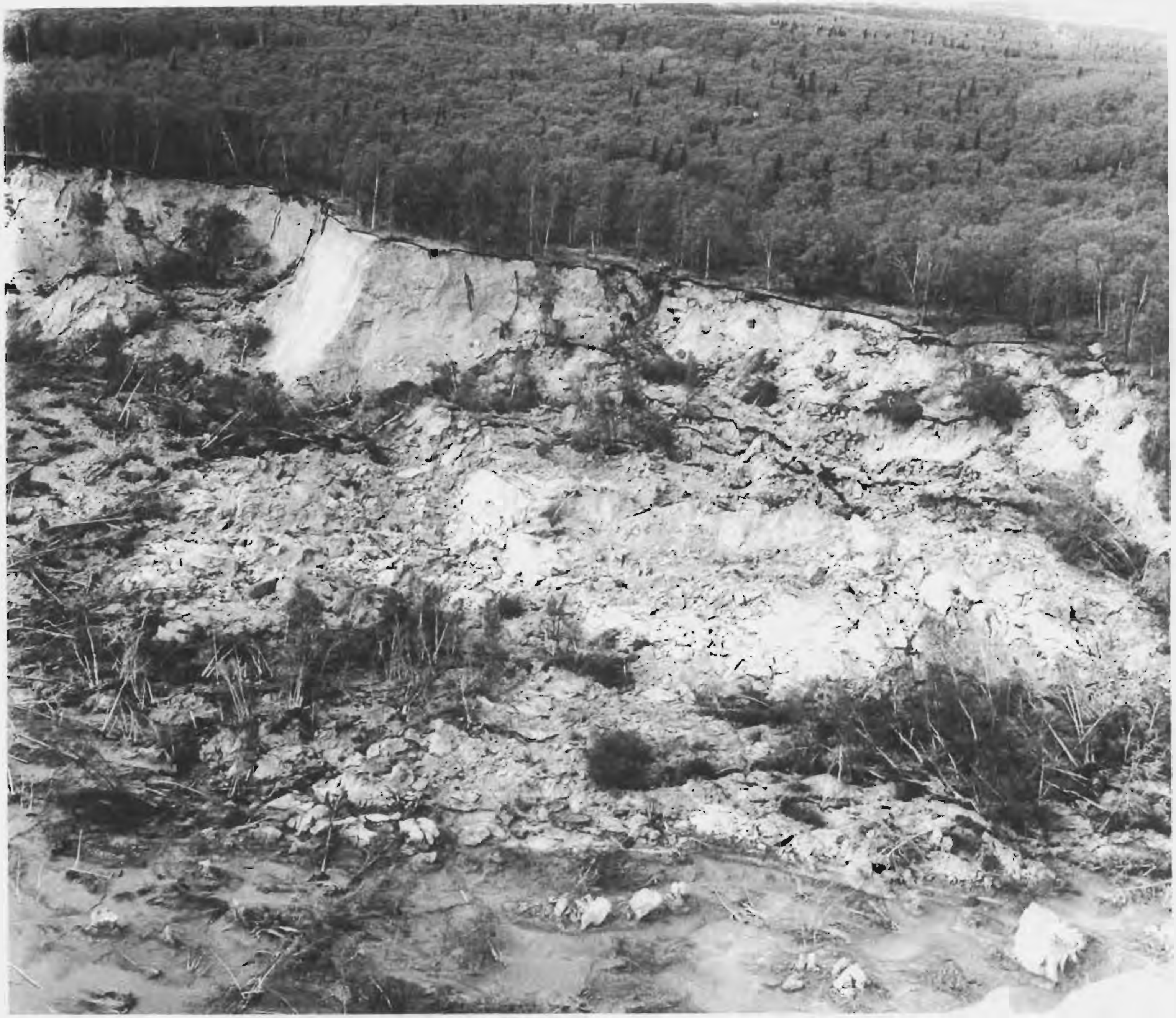
Waves and tides, by periodically removing sluffed increments of sand and gravel from the foot of the point, have kept the slopes above in a state of precarious repose. It is not surprising, therefore, that on March 27 large volumes of material slumped down the face of Point Woronzof under the driving force of the earthquake. As shown by figure 18, the bluff caved away in three separate parts, mostly by modified rotational slumping. The largest mass, directly beneath Point Woronzof, extended about 1,500 feet along the bluff and about 50 to 100 feet back behind the old bluff line. Part of the slumped mass disintegrated into a debris avalanche, but most of it slid down as an intact though much fractured block. It probably moved very rapidly.

The next smaller slide was 300 feet east. It extended about 500 feet along the bluff and caved back about 30 to 40 feet behind the old bluff line. It, too, was a much fractured but basically intact slump. The smallest slide, 150 feet farther east, was about 200 feet wide. It broke away from the slope entirely below the old bluff line.

SLIDES NEAR CAIRN POINT

Several small landslides were generated by the earthquake along Knik Arm just south of Cairn Point at the west end of Elmendorf Air Force Base. Cairn Point, like Point Woronzof, projects out into Knik Arm and is subject to vigorous erosion by waves and tidal currents. It, too, therefore, has a past history of instability and slumping.

At Cairn Point an exceptionally thick sequence of Bootlegger Cove Clay is overlain by silty till of the Elmendorf Moraine. The clay is at least 126 feet thick, and the till is at least 110 feet thick (Miller and Dobrovolny, 1959, p. 39). The quake-induced landslides probably



19.—Rotational slump near Sleeper landing strip, on the west side of Knik Arm opposite Cairn Point. At its foot, the slump passed into an earthflow. Tidewater in foreground. Scarp to left of upper center is about 80 feet high.

involved both materials, but they broke away mostly from the lower half of the slope and, hence, were mostly in the Bootlegger Cove Clay. Morphologically, the slides were rotational slumps, modified by disruption and flowage.

The largest slide was about 450 feet wide at beach line, and it surged out at least 200 feet onto the tidal mudflat. From its appearance after the quake, it must have disintegrated into many blocks as it moved downslope, the parts that moved farthest disintegrating most completely. At its toe, it passed into an earthflow. The several other slides, although smaller, were identical to the largest in form.

SLIDES ON WEST SIDE OF KNIK ARM

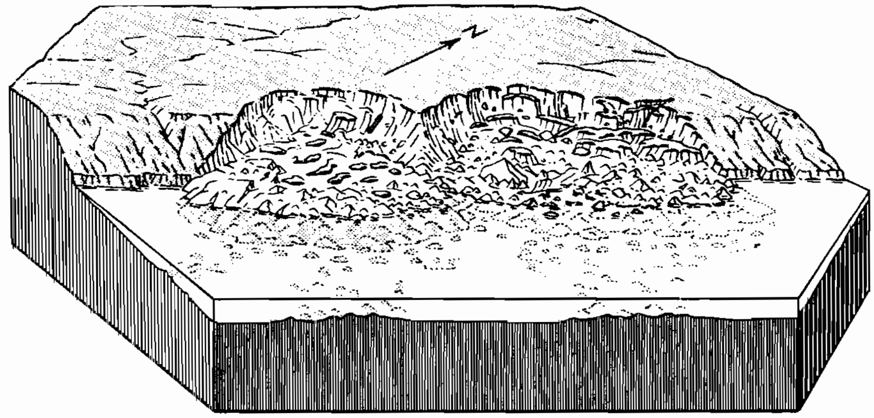
Two large rotational slumps were generated by the earthquake on the west side of Knik Arm about a mile northeast of Sleeper landing strip. The larger of the two (fig. 19) was just below hill

204 in sec. 35, T. 14 N., R. 4 W. (This hill is shown on the U.S. Geological Survey 1962 map of Anchorage and vicinity.) The smaller slide was about a quarter of a mile southwest. The Bootlegger Cove Clay rests on glacial till at about the level of mean high tide, with a contact that appears to be gradational. The clay contains scattered pebbles throughout, especially near the base; some are as large as 4 inches across.

Bootlegger Cove Clay extends nearly the full height of the bluff, to an altitude of about 190 to 195 feet, where it is topped by a veneer of sand, silt, and peat. Sliding, therefore, was largely in the clay. Till may have been involved near the foot of the slides, but if so, it was pushed out into Knik Arm below tidewater. Both slides surged out onto the tidal mudflat in broad earthflow lobes that reached an undetermined distance into the water below low tide.

Before the earthquake, the top and slope of the bluff supported a forest of birch and cottonwood in the slide areas. These trees were rotated backward in an arc of about 30° to 60°, the angle of tilt being progressively greater farther downslope. In the earthflow lobes, many of the trees lay flat, their crowns pointed toward the bluff and their roots toward Knik Arm.

Detailed land measurements have not been made of either slide. Estimates based on air and ground reconnaissance and on photographs indicate that the larger slide was about 700 feet wide along the bluff line. It worked headward about 200 feet into the bluff, and it surged out perhaps 500 feet onto the mudflat for a total length of at least 700 feet from crown to toe. It was a compound slide in that it extended back into the bluff in two large joined arcuate alcoves (fig. 20). The slide was complex in that the slumped mass broke into several slices below the crown and into countless jumbled blocks toward the foot, where it passed into an earthflow. Its main scarp had a pronounced concave-outward profile and a maximum height of about 80 feet from the crown to the jumbled blocks below.



20.—Block diagram of compound slump near Sleeper landing strip. Two lobes of slide surged out onto mudflat below high tide.

The smaller slide was similar to the larger one in morphology and habit, but its main scarp formed a simple arcuate alcove about 90 feet high. Gouging and slickensides on the face of the scarp indicate that the slumped mass slid diagonally downward from upper left to lower right (as viewed from the crown) rather than directly down the fall line; the reason for the diagonal slipping is unclear, unless the mass slumped almost instantaneously and at the same time lurched laterally in response to the ground motion of the quake. The slumped mass itself was rotated backward toward the bluff, but it was broken into helter-skelter blocks in various attitudes of disarray, particularly out toward the toe.

Near the foot of the smaller slide there were sand boils 3 to 4 feet high. Because the bluff had only a thin veneer of sand at the top, the boils presumably were derived from the subjacent beach.

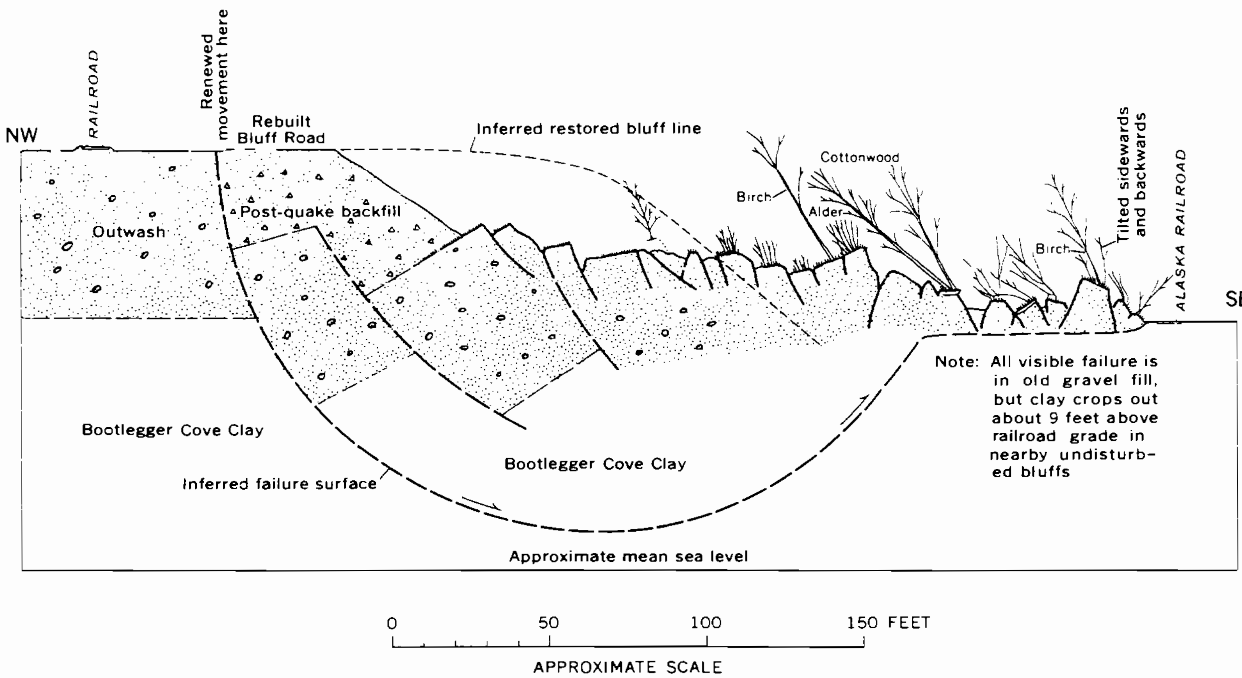
The slip plane of an older slide was exposed in section on the south side of the main scarp of the smaller slide. The older slide—no longer preserved as a topographic form—was truncated by the slip plane of the younger one.

Further slumping at a gradually diminished rate is forecast for both slides. The toes of both slides project far out into Knik Arm where they are under relentless attack by tidal currents and waves and where removal of material will reduce whatever buttressing effect may have existed immediately after the quake. Loss of buttress support, in turn, will encourage further rotation of the main slide masses; this rotation will remove support from the main scarps. These scarps already are precarious, and will remain so until headward slumping has reduced their height and pitch to a more stable configuration. In brief, therefore, erosion at the toes and slumping at the heads will continue indefinitely at a gradually diminishing rate until something approaching the prequake profile of the bluff has been restored.

21.—Chaotic pressure ridge at toe of rotational slide, Bluff Road.. In part, toe surged forward as an earthflow. Compare with figure 22.

ROTATIONAL SLIDES ALONG BLUFF ROAD

Several rotational slides gave way along the south-facing bluff line of Ship Creek, between Bluff Road and the tracks of The Alaska Railroad (fig. 1). The most widespread failure in that area extended east along the bluff between the district offices of the Corps of Engineers and the steam generating plant of Elmendorf Air Force Base. About 1,300 feet of bluff gave way in a compound displacement consisting of four separate but connected slumps. Collapse at the heads of the slumps extended back horizontally into the bluff as much as 120 feet. An additional 400 feet of bluff line east and west of the main slide mass cracked and began to slump. Altogether, about 650 feet of Bluff Road was destroyed, dropping as much as 40 feet below the prequake gradeline and rotating backward 15° to 30°.



22.—Sketch section through rotational slide at Bluff Road.

All four slumps were modified by multiple cracking, internal breakup, and flowage. None moved as a simple block. Within each slump, slices rotated on individual slip surfaces, their motions modified by interference or shoving of one slice against another. There was evidence, moreover, of flowage beneath each slump, as well as of extrusion of material at the toe.

At the toe, each of the four slumps bulged out into a chaotic lobe of jumbled blocks (fig. 21). These lobes were transitional in form between pressure ridges and earthflows. Three of them stopped short of The Alaska Railroad tracks; the most easterly lobe, however, advanced across two sets of tracks, but without damaging the tracks or the roadbed. Each slump had a length of about 300 feet from crown to toe.

In general, the rotation of blocks within the slides diminished downslope from the head of each slide toward a position about halfway down, below which rotation either increased again or the attitude of the blocks became chaotic (fig. 22). In other words, the old ground surface was tilted most near the head and least about halfway to the toe. This differential tilting (in addition to rotation) seems to have been in response to a withdrawal of material from beneath the headward parts of the slumps, either by flowage and extrusion, by lateral spreading, or by a combination of the two. If by a combination, these slides represent a transitional form between the simple slumps of the Point Woronzof and Cairn Point areas and the block-glide displacements in the downtown Anchorage area.

Most of the visible landslide material was coarse outwash gravel. Some of it had been disturbed prior to the quake, and some plainly had been backfilled artificially. Bootlegger Cove Clay, however, was exposed along a line of springs 9 feet above the base of the slope of the bluff just west of the slides, and it must have extended into the slide area. Its loss of strength under the vibratory motion of the quake very likely contributed to the failure of the slope and to the flowage at the toes and beneath the heads of the slumps.

East from the steam generating plant, cracks and smaller separate slumps extended along the bluff a total distance of about 3,000 feet. The bluff beyond was appreciably lower and presumably more stable. Some of these slumps showed clear evidence of movement prior to the March 27 earthquake, particularly the slump just east of the steam plant and the one in the next bend of Bluff Road farther east. Old settlement cracks in the road at these places had been filled and patched before the quake.

Emergency repairs along Bluff Road did little to stabilize the sliding triggered by the quake. These repairs consisted simply of backfilling collapsed areas with gravel to reestablish the grade profile, without regard for the unbalancing effect that the backfilling

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might have on the slide itself. At several points along the bluff, fill material dumped onto the slides probably caused continued movement by overweighting the head and altering whatever balance had been achieved naturally. By mid-May of 1964, new cracks with small vertical displacements had already formed in the repaired fills. Unless remedial procedures are altered, therefore, continued slow movements at the heads of the slumps seem inevitable. A comprehensive stabilization program for these slides, on the other hand, may be less feasible economically than intermittent road repair and maintenance.

TRANSLATORY SLIDES

All the highly destructive landslides in the built-up parts of Anchorage were of a single structural-dynamic family, despite wide variations from slide to slide in size, appearance, and complexity. All moved chiefly by translation rather than rotation. They slid laterally on nearly horizontal slip surfaces following drastic loss of strength in previously weak sensitive zones of the Bootlegger Cove Clay. Slides in which the slid mass was practically intact are classed as block glides; those in which the slid mass underwent appreciable disruption are probably best classed as failures by lateral spreading (Varnes, 1958, pl. 1). Between these limits, all gradations of form were represented—not only from place to place but also in time. Structurally, the Fourth Avenue slide was the simplest and the Turnagain Heights slide was the most complex. The Turnagain Heights slide, however, must have begun as a simple though highly transient block glide, or perhaps as several such block glides, arising independently along the bluff line. As sliding progressed, the Turnagain Heights slide deteriorated rapidly into a complex failure involving simultaneous motions in several directions. Its predominant motion, however, remained translatory. Destruction to property in the several slides was caused by tilting, wrenching, warping, and disruption of structures over the cracked, collapsed, and compressed zones of the slides. Structures in undistorted parts of some slides were little damaged despite horizontal ground translations of several feet.

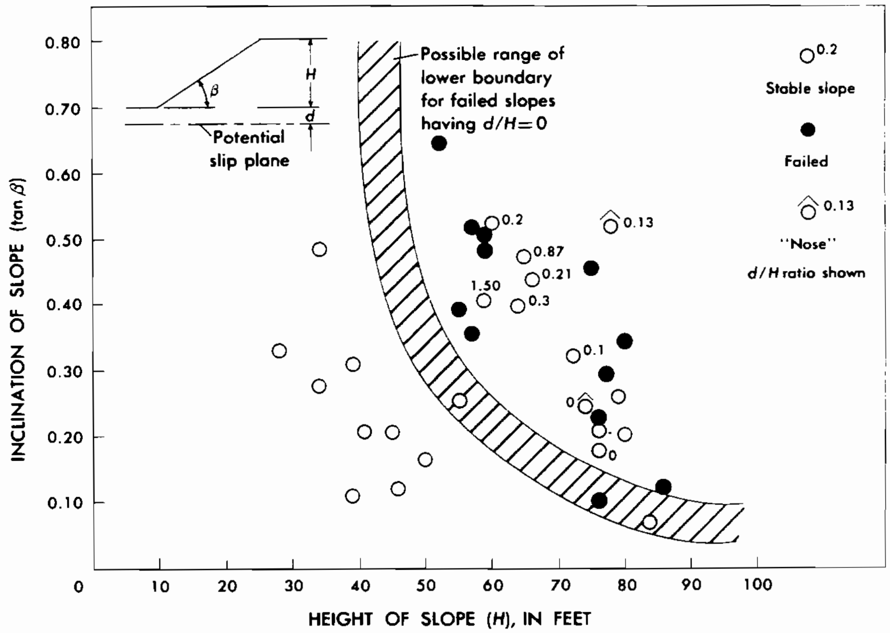
Translatory slides are less common than rotational slides. They have, therefore, received less attention in the literature. Examples similar in many respects to those at Anchorage, however, have been reported and described from Scandinavia where the Pleistocene history has been comparable in some ways to that at Anchorage. Notable translatory slides occurred at Skottorp, Sweden, in 1946 (Odenstad, 1951) and at Bekkelaget, Norway, in 1953 (Eide and Bjerrum, 1955, p. 88–100; Rosenqvist, 1960, p. 10). The Skottorp slide, particularly, resembled the Turnagain Heights slide, except that it was smaller; the Bekkelaget slide was similar to the Fourth Avenue slide. Translatory slides in the conterminous United States somewhat like those in Anchorage have been described by Crandell (1952, p. 552; 1958, p. 73) and by Varnes (1958, p. 26–32). All these slides—in Scandinavia and in the United States—moved in response to gravitational stress, without the intervention of earthquakes.

Earthquake-triggered landslides accompanying the great New Madrid, Mo., earthquakes of 1811 seem to have been very similar to the slides at Anchorage. The physical setting of the slides was analogous to that at Anchorage. Fissures, sand blows, and various other features related to sliding are described by Fuller (1912, p. 48, 59–61) on the basis of early accounts and on observations made nearly 100 years after the quakes. Fuller's descriptions might well apply to the Turnagain Heights slide. Grabens and tension cracks, for example, were formed where clayey alluvium, afloat on quicksand, glided laterally.

The Chilean earthquake of May 22, 1960, triggered three large landslides at Lago Riñihue, 65 kilometers east of Valdivia in central Chile. The descriptions of Davis and Karzulovíc (1963, p. 1407) indicate that these slides resembled the Turnagain Heights slide in form, size, and mode of failure. Translatory movement predominated; rotational movement was subordinate. Significantly, these slides occurred in an area of previous landsliding and were partly superimposed on preexisting landslides.

GEOLOGIC SETTING

All areas of translatory sliding in Anchorage had the same general geologic environment. All were underlain at various depths by Bootlegger Cove Clay that had zones of low shear strength, high water content, and high sensitivity. All surmounted flat-topped bluffs bounded on one side by steep slopes. In all areas, except the westernmost part of the Turnagain Heights slide, the Bootlegger Cove Clay was overlain by outwash sand and gravel. These deposits thinned markedly over a distance of a few miles from north to south and from east to west; grain size diminished concomitantly from gravel to sand. Near the Alaska Native Service Hospital the outwash is about 55 feet thick, near Government Hill



23.—Slope stability chart. Reprinted from Shannon and Wilson, Inc. (1964, pl. 7.1).

School about 40 feet, in the downtown area about 30 to 40 feet, and at Turnagain Heights from about 25 feet to zero.

Outwash deposits had no critical part in the sliding. Failure was confined to thin zones of sensitive clay, silt, and sand within the Bootlegger Cove Clay.

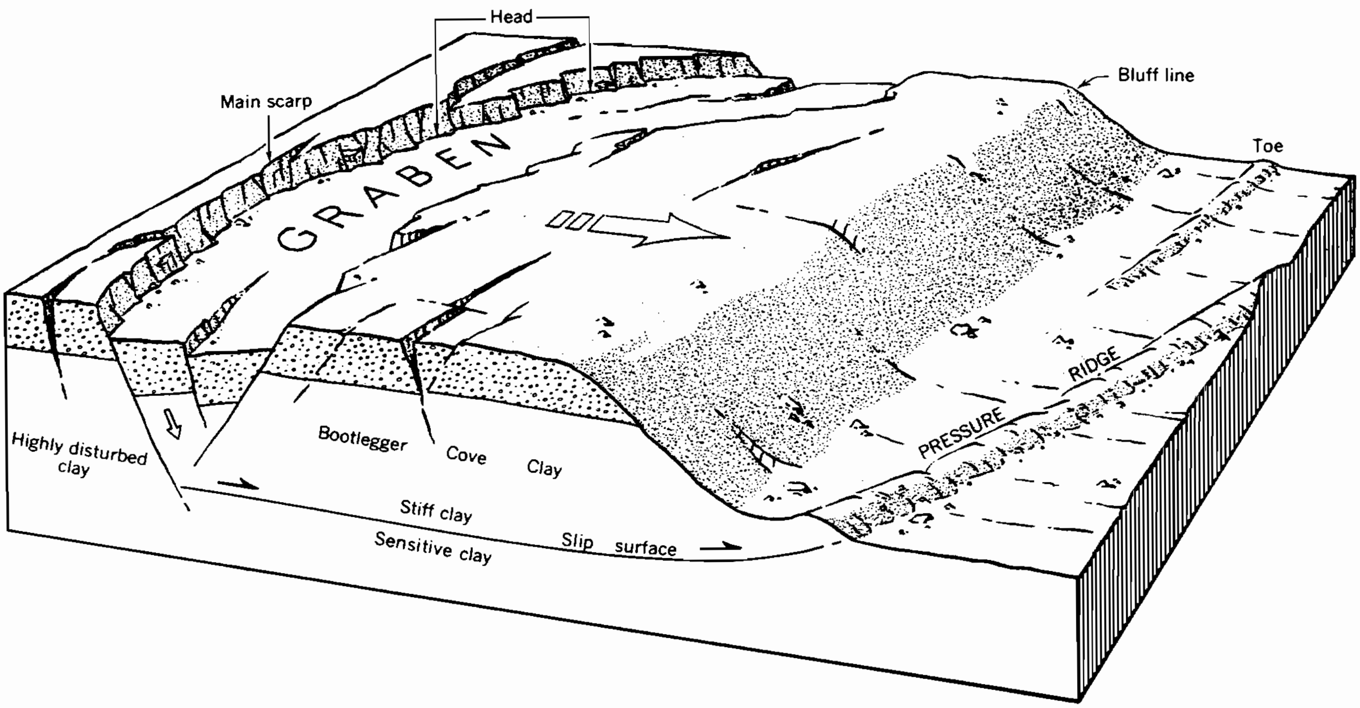
Several geologic factors acting in concert with earthquake shaking probably caused failure in the several slide areas. All these factors are themselves variables at each site. They include: (1) the topographic elements of bluff configuration—the height of the bluff above its base, the slope angle or declivity of its face, and perhaps the ground-plan configuration, (2) the soil-strength profile, including consistency, dynamic shear strength, and sensitivity, and (3) water content and liquid limit of the soil at the critical depth below the ground surface. Some of these factors seem to be interdependent—the height, slope, and ground-plan configuration, for example, may have influenced the water content of the soil, which in turn influenced the consistency.

The influence of topography, particularly the bluff profile, on failure susceptibility seems obvious. The ultimate driving force of landsliding was gravity, operating on a nearly horizontal shear surface in the direction of least shear resistance (the free face of the bluff) and acting on a dynamically sensitive soil that had undergone a severe loss of strength by earthquake shaking. Other factors being equal, the higher and steeper the bluff profile, the greater is the shearing stress, the less effective is the shear resistance of the clay, and the greater is the susceptibility of the bluff to failure.

The full significance of ground-plan configuration is uncertain and may involve several variables difficult to evaluate. Most failures took place along fairly straight or recessed bluff lines, for reasons not entirely clear, although possibly because such places are apt to be well watered (Shannon and Wilson, Inc., 1964, p. 33). Salients or cuspate areas ("noses") generally were spared failure, again possibly because they are apt to be well drained, although the slide at the Native Hospital failed precisely at just such a salient.

Using factors of height and slope, only, Shannon and Wilson, Inc. (1964, pl. 7.1), constructed a slope stability chart that grouped failed and stable slopes and showed lower limits of slope-to-height ratios below which failure is improbable (fig. 23). The higher the slope, the flatter the angle at which the slope is susceptible to failure, other variable factors such as soil strength and water content being equal. The fact that several stable slopes, including salients or "noses," have slope-to-height ratios within the same area of the chart as slopes that failed seems to point up the modifying influences of other factors such as soil-strength profile and water content. A low slope underlain by soil having a high water-plasticity ratio, for example, may be more susceptible to failure—may have a lower safety factor—than a high slope underlain by soil having a low water-plasticity ratio. In a unit of time, other factors being equal, a salient should drain and desiccate more rapidly than a recess; hence it should be more stable.

The length of time a bluff has been stationary and unmodified by erosion may perhaps influence its dynamic stability by affecting the water content of the underlying Bootlegger Cove Clay. Far more information on the water content of the clay at critical localities than is now available would be needed to test this hypothesis, but presumably a long-stationary bluff line would have a better chance of desiccation, and hence of



24.—**Block diagram of a translatory slide.**

dynamic stability, than an actively retreating one. Permeability is extremely low in most of the Bootlegger Cove Clay, and a long period of immobility would be required to promote even a low degree of drainage and desiccation. At Turnagain Heights, for example, the bluff line facing Knik Arm, which failed so dramatically during the earthquake, was in a state of relatively rapid regression owing to active shoreline processes prior to the quake, whereas the bluff line facing Fish Creek was relatively stationary before the quake and did not fail during the quake.

GEOMETRY AND MODE OF FAILURE

Translatory slides of the Anchorage area varied widely in size, shape, and internal complexity, but they all conformed to a single basic geometric format. Figure 24 is a simplified block diagram showing the essential structural elements, including the head, toe, bluff line, slip surface, graben, and pressure ridge. Sections through both the Fourth Avenue and the L Street slides closely approximate the idealized section at the front of the block diagram. The so-called slip surface probably is not generally a surface at all, at least at the onset of sliding, but rather is a narrow planar zone of failure.

Geometric development of the translatory slide is reconstructed in the following sequence: Earthquake shaking drastically reduces the shear strength of saturated sensitive zones in the Bootlegger Cove Clay. The strength of the clay falls below the level of shear stress caused by the weight of material in the bluff and the accelerations of the quake. Under the influence of gravity, therefore, a prismatic block of earth begins to move laterally on a nearly horizontal slide surface toward the free face of the bluff. In effect, the block is afloat on a zone of disturbed clay whose strength properties are those of a confined viscous liquid. As the block starts to move, tension fractures form at the head of the slide and widen as movement progresses. These tension fractures dip toward the slide block at commonly observed angles of about 60° to 70°. As the fractures widen, their hanging wall (on the moving block) loses support and collapses along one or more antithetical fractures to form a graben.

The term "graben," since the earthquake, has become a household word to the populace of Anchorage. The downward piston-like movement of the graben is synchronous with the lateral slippage of the slide block. Continued slippage places more ground under tension behind the slide, and additional fractures form as the slide retrogresses headward. In the more complex slides such as the Government Hill slide, and, especially, the Turnagain Heights slide, this process occurred repeatedly, and a series of alternate horsts and grabens developed regressively, parallel to the direction of slippage. (See figs. 37, 44, and pl. 2).

As the block moves outward, tension at the head of the slide is partly countered by compression at the toe, and pressure ridges form in the flats below. This step may be transient, because continued compression leads to rupture and overthrusting. Both steps occurred in each of the translatory slides and caused extensive local damage. The Turnagain Heights slide not only sheared off at the toe, but it slid under gravity down the mudflat into Knik Arm—at one point it slid more than half a mile. But if resistance in the toe builds up to a level equal to the thrust of the moving block plus the shear resistance of the clay at the slip surface, motion will stop. A modification of this principle was utilized by the Corps of Engineers in designing remedial buttresses. The Corps, however, visualized the driving mechanism of the slide as the "active earth pressure" of the earth mass behind the head of the slide (Shannon and Wilson, Inc., 1964, p. 32) plus the acceleration of earthquake ground motion. But inasmuch as the head of the slide block is plainly under tension during sliding, there obviously can be no "active earth pressure" from the mass behind it. Furthermore, inasmuch as regression and sliding continued after earthquake ground motion had ceased (p. A64), the accelerations of the quake cannot be the driving mechanism either. Rather, sliding was precipitated by gravity as soon as the accelerations of the quake had reduced the shear resistance of the soil to the point of failure. The earthquake was the "trigger"; gravity was the "propellant." Sliding continued as long as the gravitational component on the sliding surface exceeded the shear resistance of the soil. At Turnagain Heights, sliding probably continued a full minute or more after earth shaking had stopped (see page A64).

In plan, each slide was bounded laterally by a series of crescentic tension fractures, which also marked the outer wall of the outermost graben. Subordinate crescentic tension fractures—bounding potential slide blocks—commonly extended headward, outside the slide proper, scores or hundreds of feet beyond the bounding fractures. At Turnagain Heights, numerous crescentic tension fractures extended back through the subdivision as much as 2,200 feet beyond the head of the slide (pl. 1); had the earthquake lasted longer, sliding undoubtedly would have retrogressed back into that area. "Retrogression," as applied to landslides, means a headward expansion of the slide. (See Varnes, 1958, p. 31.)

Extensive internal fracturing also characterized all slides, although internal fracturing was minimal in the L Street slide. The Native Hospital slide contained abundant radial fractures; as noted before, this slide occupied a salient or "nose" on the bluff line, and the radial fractures undoubtedly were tensional responses to lateral spreading.

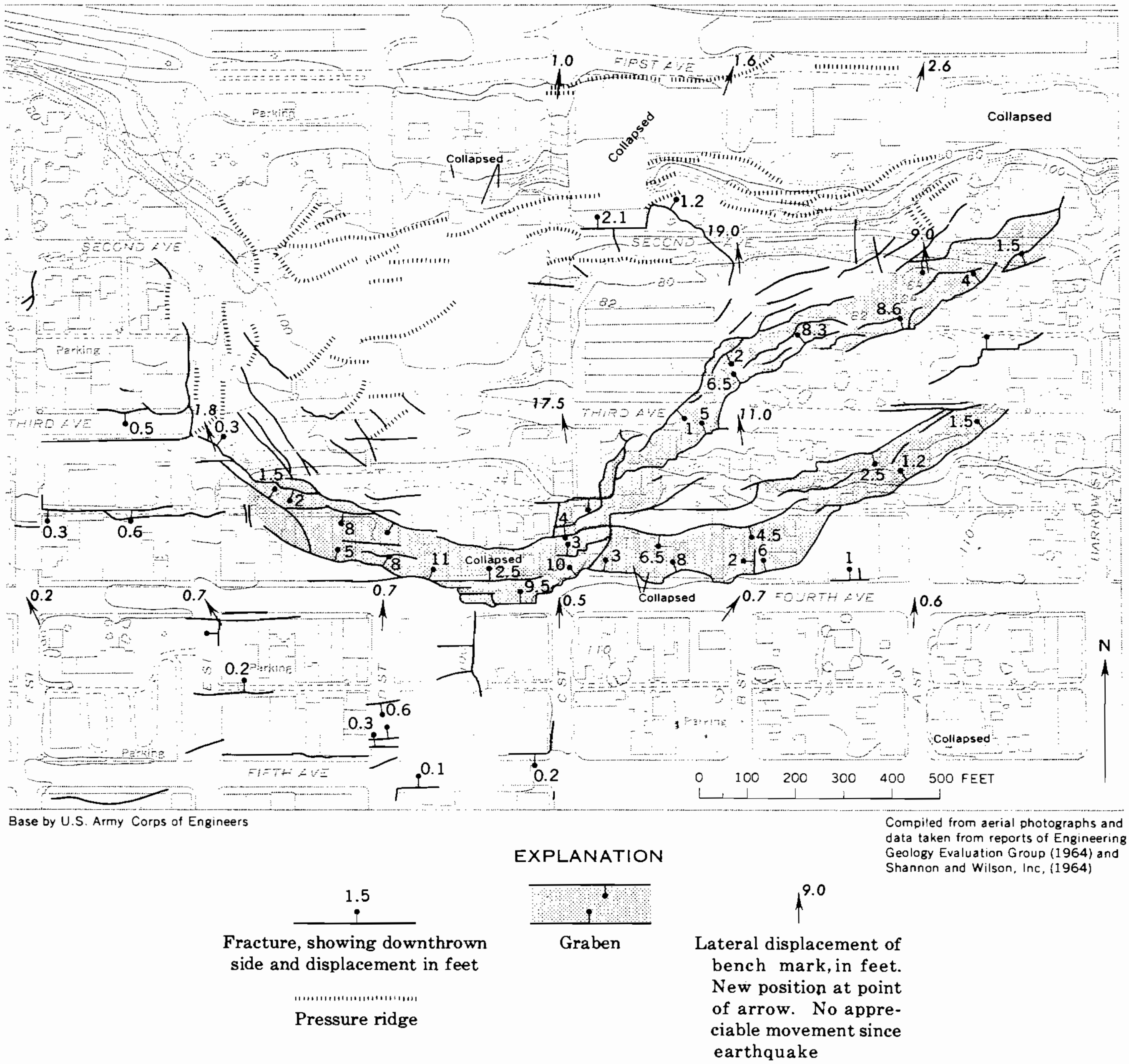
THE GRABEN RULE

A close approximation of the depth of failure is prerequisite to planning remedial or stabilization procedures. A significant geometric relationship, here called the "graben rule," affords a rapid, yet reliable, estimate. The graben rule should apply to any translatory slide in which flowage of material from the zone of failure has not been excessive. Because the cross-sectional area of the graben trough approximates the cross-sectional area of the space voided behind the block as the block moves outward, the depth of failure can be estimated from the simple relationship $D=A/l$ where D is the depth of failure, A is the cross-sectional area of the graben, and l is the lateral displacement of the block. For example, the Fourth Avenue graben, on the average, was about 11 feet deep, 100 feet across, and had an area (A) of about 1,100 square feet. Its maximum lateral displacement (l) was about 17½ feet, as determined by postquake resurveys. The calculated depth of failure (D) was about 63 feet, or about 43 feet above mean sea level. Subsurface exploration, though somewhat indecisive in its results, yielded a nearly identical figure.

INDIVIDUAL TRANSLATORY LANDSLIDES

Fourth Avenue Slide

The Fourth Avenue slide involved all or parts of 14 city blocks in a roughly oval area of about 36 acres, containing perhaps 2 million cubic yards of earth, centered at the north side of downtown Anchorage (fig. 25). It was bounded headward on the south by Fourth Avenue, on the west by E Street, on the north approximately by First Avenue, and on the east somewhat indefinitely by Barrow Street. Its length north to south in the direction of slippage was about 1,050 feet; east to west it was about 1,800 feet across. Strong fracturing and related ground displacements extended 1½ blocks (about 450 feet) or so south of the slide proper, where considerable damage was inflicted on buildings,



25.—Fourth Avenue landslide area, Anchorage, Alaska.

streets, and sidewalks. Minor displacements extended as far south as 600 feet. Eyewitnesses reported that sliding began about 2 minutes after the earthquake started and stopped about the same time as the earthquake (Grantz, Plafker, and Kachadoorian, 1964, p. 15).

Most of the damaging tensional fractures within the slide mass were between Second and Fourth Avenues in the east part of the slide and between Third and Fourth Avenues in the west part. North from those areas, downslope, compressional movements in the foot of the slide caused numerous pressure ridges.

Ground dislocations were most severe in the graben areas at the head of the slide along the north side of Fourth Avenue (fig. 4). Ironically, these areas also had the highest property evaluations. Many small business and commercial buildings, apartment houses, and residences were destroyed or badly damaged. Vacant lots and parking spaces within the slide block to the north were little disturbed. Just east of C Street the main graben bifurcated, and two long belts of severe damage extended northeast toward Barrow Street.

Between B and D Streets, where the destruction was total, the main graben had a width of 100 to 150 feet and a depth of as much as 11 feet. Lateral displacements near the center of the slide just north of the graben were as much as 17½ feet toward the north, according to surveys by the City Engineer's office. Near the foot of the bluff on Second Avenue a block east of B Street, a complex of pressure ridges was pushed up by 19 feet of lateral displacement—apparently the greatest local slippage in the slide. Although several days elapsed after the quake before resurveys were started, there appears to have been no movement since the main quake. None of the strong aftershocks caused measurable movement.

Subsurface exploration of the Fourth Avenue slide was started by the Engineering Geology Evaluation Group soon after the earthquake, and was expanded and concluded by the Corps of Engineers. The findings of these groups indicate that failure occurred at the top of the sensitive zone of the Bootlegger Cove Clay at a depth of about 60 feet (48 feet above mean sea level). Additional failures may have occurred at greater depths, inasmuch as many pressure ridges formed at the surface in the toe of the slide at altitudes below 48 feet, unless, as suggested by their sharply peaked crests and steep sides (fig. 26), the ridges were caused by simple surface translations of the rigid frozen surface layer of the soil.

Shear-strength profiles measured near the head of the slide by Shannon and Wilson, Inc. (1964, p. 41) show that the clay decreases in strength from about 1.0 tsf at altitude 70 (top of clay) to as low as 0.25 tsf between altitudes 45 and 20, then increases in strength to about 0.75 tsf at 15 feet below sea level. The zone of maximum sensitivity coincides with the zone of minimum strength, and failure appears to have followed spontaneous liquefaction of sand layers as well as loss of strength in silty clay.

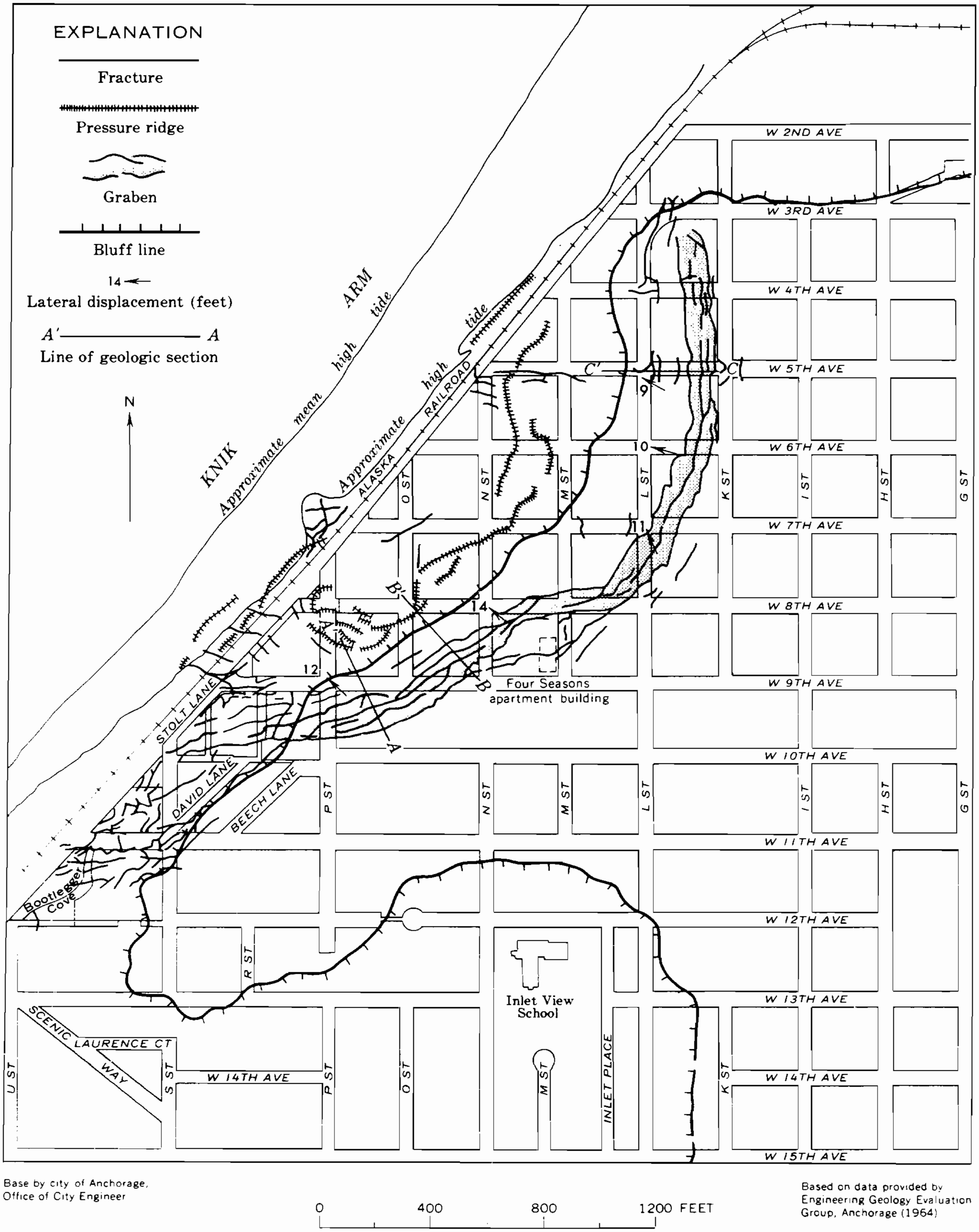
L Street Slide

The L Street slide involved all or parts of about 30 city blocks in the northwest part of Anchorage adjacent to Knik Arm (fig. 27). It extended northeast about 4,800 feet along the bluff and had a maximum breadth northwest across the bluff of about 1,200 feet, parallel to the direction of slippage. It reached about a block and a half back from the bluff line into thickly settled residential and commercial neighborhoods of Anchorage. In all, about 72 acres were included between the graben at the head of the slide and the outermost pressure ridges at the toe. The total volume probably approached 6 million cubic yards though an accurate estimate is difficult. Much of the 72-acre area, however, was little if at all damaged, despite lateral shifting of as much as 14 feet. Most of the damage was concentrated along the graben, which with marginal fractures covered about 14 acres, and along the pressure ridges, which were mainly linear features but which involved properties of a total area of perhaps 7 to 8 acres. There was very little fracturing much beyond the bounding fractures of the graben, and there was not much fracturing within the slide mass itself except for the graben area. Eyewitness reports indicate that slippage began during the latter part of the earthquake (Shannon and Wilson, Inc., 1964, p. 53).

The whimsical pattern of destruction in Anchorage was perhaps best exemplified by the L Street slide; here wrecked buildings inside or astride the graben faced almost undamaged adjacent properties on either side (fig. 28). The irregular trends of the bounding fractures further compounded the fateful selectivity of the slide. Damage was equally capricious, moreover, in the compressed areas at the toe of the slide; here individual dwellings were buckled or shoved by pressure ridges that as often as not left adjoining buildings undisturbed (fig. 29).

The graben itself formed a broad arc in plan, concave toward the slide block. It extended south from West Third Avenue between K and L Streets and curved westward to R Street between Ninth and Tenth Avenues. It was about 3,600 feet long and had a maximum width of about 250 feet; generally, its width was between 150 and 200 feet (fig. 30). Its depth—that is, the displacement of the down-dropped block—reached a maximum of about 10 feet; the displacement was greater south of Sixth Avenue than north. Overall, the graben looked like a dry canal or a streambed, and, when contrasted with the lack of damage on either side, it stirred considerable speculation in the minds of early viewers. One popular magazine account stated that it resulted from collapse of an old

26.—Sharp-crested pressure ridge at Second Avenue and C Street, Anchorage. Probably caused by shallow translation of frozen surface layer. Note smaller pressure ridge in street part way up block and toppled chimney of building behind car. Photograph by Mac's Foto, Anchorage.



Base by city of Anchorage, Office of City Engineer

Based on data provided by Engineering Geology Evaluation Group, Anchorage (1964)

27.—L Street slide area, Anchorage, Alaska. Geologic cross sections are shown in figure 30.

28.—Wrecked dwelling astride boundling fracture, L Street graben at Eighth Avenue and N Street. Damage caused entirely by ground displacement along fracture.

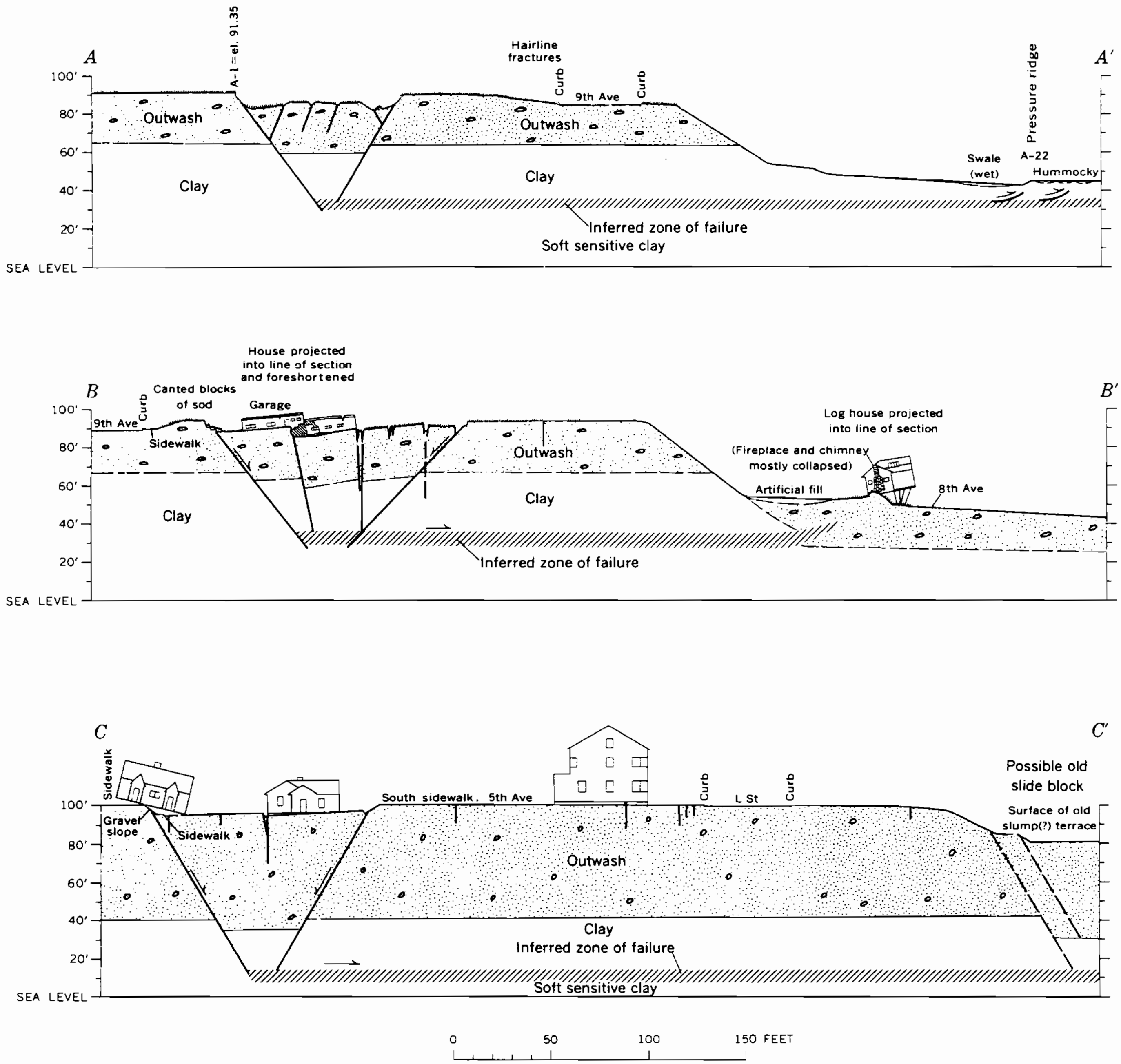
29.—House pushed off foundation by pressure ridge at toe of L Street slide. Well-framed house not otherwise visibly damaged. Push must have been shallow and nearly horizontal.

buried but melted-out ice-filled channel!

Many buildings on the slide block, including a six-story apartment building carried 10 feet laterally, sustained little or no damage, but utility service to the slide block was curtailed. Overhead wires and buried water, gas, and sewer lines all were disrupted where they crossed the graben. In most places, entirely new emergency connections had to be made before service could be restored. Many buildings, therefore, which were not themselves damaged were nevertheless evacuated.

Pressure ridges below the slide block were concentrated mostly in a zone about 200 feet wide close to the foot of the bluff at 40 to 55 feet above sea level (figs. 27, 30). A few extended beyond The Alaska Railroad tracks, which were damaged by lateral shove, and even onto the tidal flat of Knik Arm about 500 feet from the foot of the bluff and only about 15 feet above mean sea level. Individual ridges ranged in length from a few tens of feet to more than 600 feet. They ranged in width from sharply peaked ridges 10 to 15 feet across to broad gentle bulges several tens of feet across. Most were less than 3 feet high, but a few were as high as 7 feet. Many were ruptured and overthrust toward Knik Arm.

The available evidence, both surface and subsurface, favors the view that the pressure ridges were caused chiefly by shallow compressional translations of the frozen superficial soil mantle, which in turn was shoved laterally by the sliding block behind. This mantle was only a few feet thick, but it was highly competent in its frozen state. Some indication of its strength was provided at one point where a small chunk of partly buried concrete caught in a surface fracture was so firmly held in place that it parted in half instead of breaking free from the frozen soil. In further support of the shallow depth concept, the altitude of the



30.—Geologic section through the L Street slide. In section *C–C′* the toe of the slide is to the right of *C′*.

pressure ridges above sea level was very close to the inferred altitude of failure beneath the main slide block.

Damage caused by pressure ridges was extensive in the L Street slide although it was generally less devastating and less spectacular than that caused by tensional cracking and displacement in the graben area. Even so, losses to many small dwellings totaled thousands of dollars; more expensive losses were sustained by larger structures. Some buildings that showed little evidence of damage on the outside sustained severe structural damage within. Utilities were badly damaged also.

Subsurface exploration of the L Street slide, as of the Fourth Avenue slide, was started by the Engineering Geology Evaluation Group soon after the earthquake, to learn the causes of sliding, to predict future behavior of the slide, and to recommend courses of action to be taken by the city. Subsequently, this work was transferred to the Corps of Engineers.

Extensive drilling and sampling by the Corps, augmented by surface studies, indicated a complex and varied stratigraphy, modified by erosion downslope from the bluff line and by prior landsliding of undetermined age. Shannon and Wilson, Inc., (1964, p. 54) reported the following general sequence of stratigraphy:

1. At the top, outwash sand and gravel, generally 40 to 60 feet thick.
2. Stiff, silty clay, about 30 to 50 feet thick, and interbedded layers of sand and silt. This clay had a static shear strength generally greater than 0.5 tsf.
3. Silty clay, sensitive, 20 to 30 feet thick, having very sensitive layers of clayey silt, silt, and fine sand. Static shear strengths ranged from about 0.2 tsf to more than 0.5 tsf, and sensitivity ranged from about 5 to 30.
4. Stiff silty clay containing scattered sand grains, pebbles, and lenses of sand; thickness undetermined but greater than 50 feet. Static shear strength in this unit generally exceeded 0.5 tsf and increased with depth.

Despite numerous borings, the position of the zone of failure in the L Street slide was not definitely established, although zones of very low strength and high sensitivity were clearly indicated. Presumably, slippage occurred at the top of the sensitive zone, where the strength was lowest and the shearing stress was highest (Shannon and Wilson, Inc., 1964, p. 55). This position varied appreciably—from about 55 to 85 feet below ground surface (45 to 15 feet above sea level)—in different drilling locations. Estimates based on the graben rule place the zone of failure within that range also. Pressure ridges at the foot of the slide lie within the same range of altitudes. Therefore, if failure beneath the block was within that range, the pressure ridges—as noted above—must have been caused by shallow translations of the relatively thin frozen surface layer—a sort of shove effect—rather than by deeper surfaceward shear. If so, some of the remedial buttresses proposed by the Corps of Engineers (Shannon and Wilson, Inc., 1964, pl. 9.6) would require design changes to be effective in the event of future movement.

Failure by liquefaction may also have occurred in saturated sand layers within the clay because several such layers were penetrated during the drilling program. Remolding would not necessarily lead to strengthening of these sand layers, unless repacking led to consolidation accompanied by escape of excess pore water. Unless these strengthening conditions have been met, the sand could fail again under similar circumstances as before.

Native Hospital Slide

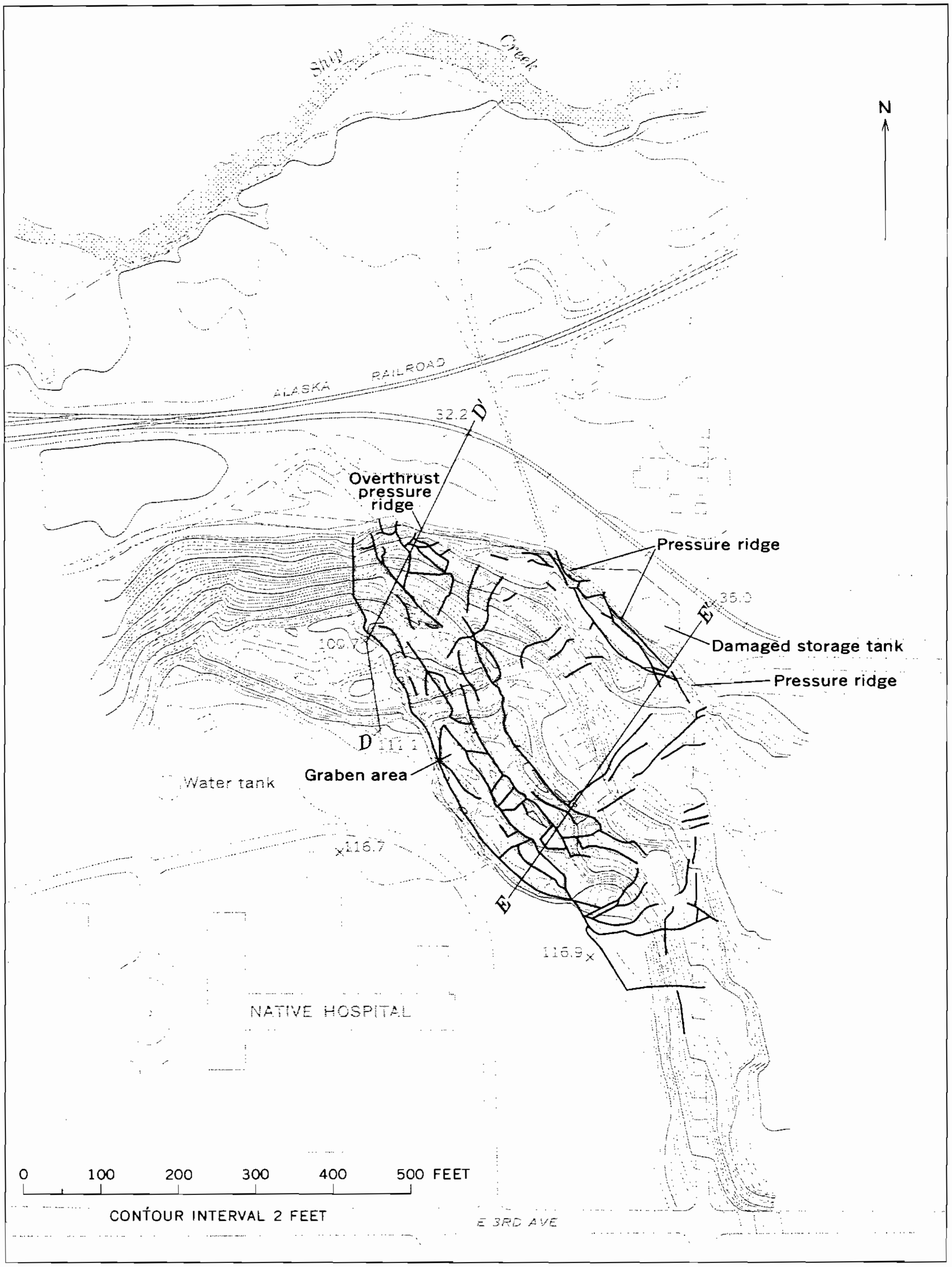
The Native Hospital slide, or First Avenue slide [2] as it has occasionally been called, disrupted part of the grounds of the Alaska Native Service Hospital and wrecked a fuel-storage tank at the foot of the bluff (figs. 31, 32). Although it was a small slide and not a very destructive one, it was of unusual scientific interest because of its clear portrayal of repeated translatory landsliding in the same area. The slide of March 27, 1964, transected an earlier slide of identical habit and exposed the older graben in full cross section (fig. 32) in the headward scarp of the present graben. This site of failure, therefore, seems to answer the question as to whether natural remolding and consolidation of the clay after sliding is sufficient in itself to forestall further sliding—it obviously was not, in the Native Hospital slide. Most other slide areas at Anchorage and vicinity also presented evidence of multiple sliding, but none did it as clearly as the Native Hospital slide. Multiple sliding is discussed further on pages A66–A67.

The Native Hospital slide involved only slightly more than 4 acres of ground and perhaps 360,000 cubic yards of earth. From flank to flank—northwest to southeast—it was about 650 feet across; from head to toe it was about 350 feet. Most of the disruption was in the slopes of a cuspate salient on the bluff line behind the hospital, but about three-fourths of an acre of upland behind the hospital collapsed into the graben as the main headward fracture opened up 120 feet or so back from the rim. Part of the parking lot of the hospital and an area of lawn-covered grounds were destroyed. Fractures that extended back from the slide damaged the hospital building itself.

The graben was exceptionally large for the size of the slide. Its disproportionate size is attributed to the large apparent lateral slippage of the slide. Arcuate in plan, it was about 600 feet long; it had a mean width of about 120 feet and a downthrow (depth) of as much as 25 feet which averaged about 20 feet. Its depth diminished rapidly toward the south, where horizontal slippage died out also.

Near the center of the slide, the downdropped graben wedge broke off precisely at the bluff line and tilted headward as it collapsed, so that the displacement at the headwall of the graben was greater than at the front wall and gave

[2] Inasmuch as First Avenue does not extend into this part of Anchorage, the name "First Avenue slide" is misleading.

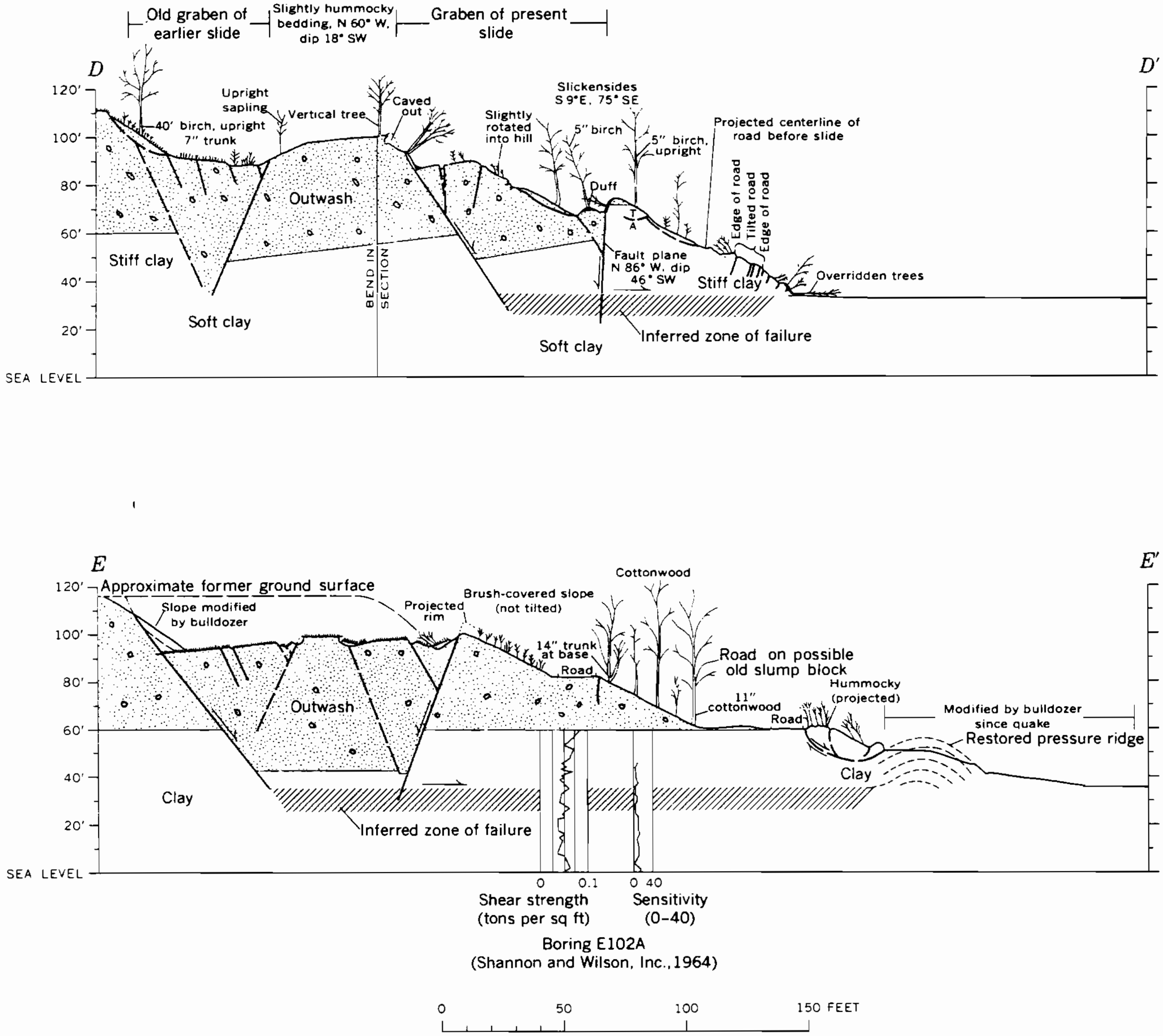


Topography sterocompiled by J. R. Helm,
U.S. Geological Survey

Graben indicated by shading

31.—Native Hospital slide area, Anchorage, Alaska. Graben indicated by shading.

32.—Air view (looking south) of Native Hospital slide showing graben and pressure ridge. The scar of an older landslide is transected by the slide of March 27, 1964. Photograph by U.S. Army.



33.—Geologic sections through the Native Hospital slide, Anchorage, Alaska. Section *D–D'* transects the graben of an earlier landslide (upper left). Boring data from Shannon and Wilson, Inc. (1964).

the slide a pseudorotational appearance; however, the causative failure and dominant movements were horizontal, as verified by the perfectly upright attitude of trees in the main slide block. Longitudinally, the graben wedge was rent by great gaping tension cracks. The total relations are well illustrated by the aerial view (fig. 32) and by the cross sections (fig. 33).

Inasmuch as the slide moved outward from a salient, it was unconfined laterally. Accordingly, it spread as it moved outward, and tension cracks opened in a fanlike arrangement at the periphery. Some of these tension cracks are visible in figure 32. They have been plotted in figure 31.

A large shallow-rooted pressure ridge about 500 feet long at the toe of the slide absorbed much of the thrust of the slide. At its greatest development the ridge was about 15 feet high and 40 to 50 feet wide. It wrecked the fuel-storage tank at the foot of the bluff (fig. 32) and overrode overturned trees where it thrust forward as much as 12 feet (fig. 34).

Exactly how much the Native Hospital slide block was displaced laterally has not been determined, owing to a lack of good prequake horizontal control. Lateral offsets in roads and trails that crossed the block, however, afford a basis for reasonable estimates. These estimates range from 17 feet of slippage near the north flank of

34.—Overthrust toe of Native Hospital slide. Road is bulged up and displaced about 12 feet laterally. Note overridden trees at left.

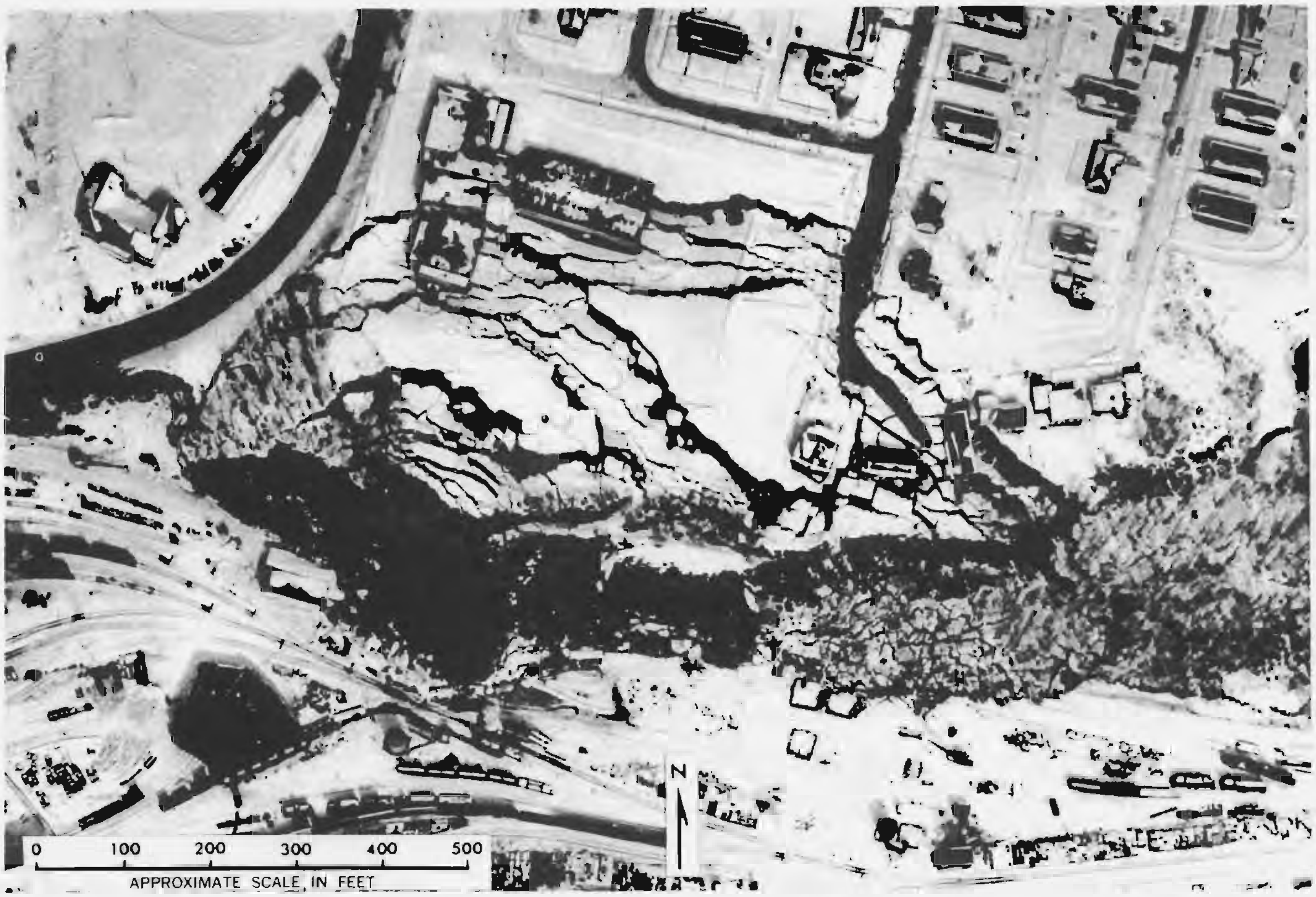
the slide to 25 feet near the center. Furthermore, at least 15 feet of slippage can be accounted for in the pressure ridge at the toe of the slide. Projections to the theoretical slip surface of the slide, based on these figures and using the graben rule, place the slip surface at a depth of 85 to 95 feet or 25 to 35 feet above sea level. This altitude is very close to the height of the flat on which the pressure ridge formed at the foot of the slide and to the top of the sensitive zone of the Bootlegger Cove Clay beneath the bluff. It is a probable altitude, therefore, for the zone of failure. Thus lateral slippage of 17 to 25 feet within the slide block is geometrically reasonable. Any appreciably smaller slippage would require an inordinately deep slip surface to compensate for the relatively large sectional area of the graben.

Figure 33 shows cross-sectional reconstructions of the Native Hospital slide based on both surface and subsurface data. Although subsurface exploration by the Corps of Engineers did not disclose the surface of rupture, it did clearly define zones of stiff clay above and below a zone of weaker clay. Most of the clay sampled had sensitivities below 10, but the sensitivity of some of it exceeded 30. Several thin sand layers may have contributed to failure by liquefaction.

Government Hill Slide

The Government Hill slide caused severe dislocations in the south-facing bluff on the north side of Ship Creek (fig. 35). Altogether, about 11 acres of land was involved, including about 2¼ acres of bottomland below the bluff where the slide passed into an earthflow and spread out in the yards of The Alaska Railroad. The volume of earth involved was about 900,000 cubic yards.

From flank to flank the slide



35.—Air view of Government Hill slide, Anchorage, Alaska. Graben plainly discernible. Compare with figure 38. Photograph by Air Photo Tech, Anchorage, Alaska.

had a width of 1,180 feet. From head to toe its greatest length, in the direction of slippage, was about 600 feet. The head of the slide regressed back about 400 feet behind the prequake bluff line, where it intersected the Government Hill Grade School. The slide devastated all but one wing of the school, destroyed two houses, damaged a third, left a fourth (since removed) perched precariously above a cliff, wrecked a shed in the railroad yards at the foot of the bluff, and did extensive damage to railroad equipment and trackage.

If any good fortune accompanied the March 27 earthquake, it was its timing; had school been in session, the disaster would have been unthinkable. The south wing of the school dropped as much as 20 feet vertically into a graben after being sheared cleanly in half (fig. 3). Electric wall clocks stopped at 5:36 p.m. The east wing, also astride a graben, collapsed after being split longitudinally. The playground was a mass of chaotic blocks and open fissures (fig. 36).

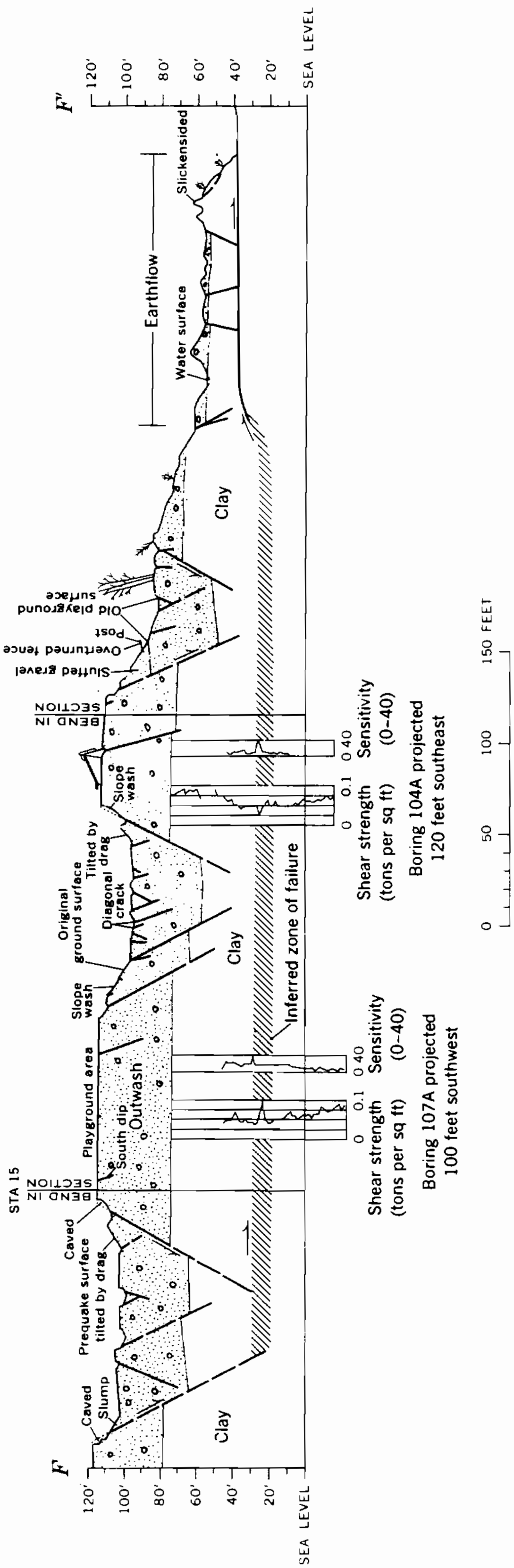
The slide was more complex than the Fourth Avenue, L Street, or Native Hospital slides. Its complexity was a step further toward the total disruption shown by the Turnagain Heights slide (fig. 37). As shown by the map (fig. 38), a complex of arcuate horsts and grabens, more or less concentrically disposed, retrogressed headward as the slide pulled away from the bluff. Two or three partly integrated grabens formed side by side along the crown of the bluff; these were complicated by internal collapses, cracking, and slumping along the face of the bluff. Two additional grabens formed behind the crown on the upland surface. The head of the slide broke away clean; back from the head there was very little cracking and virtually no damage to property.

Each graben was about 100 feet across. The two upland grabens, however, merged laterally, and at their junction the downdropped block was 200 feet across. The outer grabens were deepest—they exceeded 20 feet in depth; the

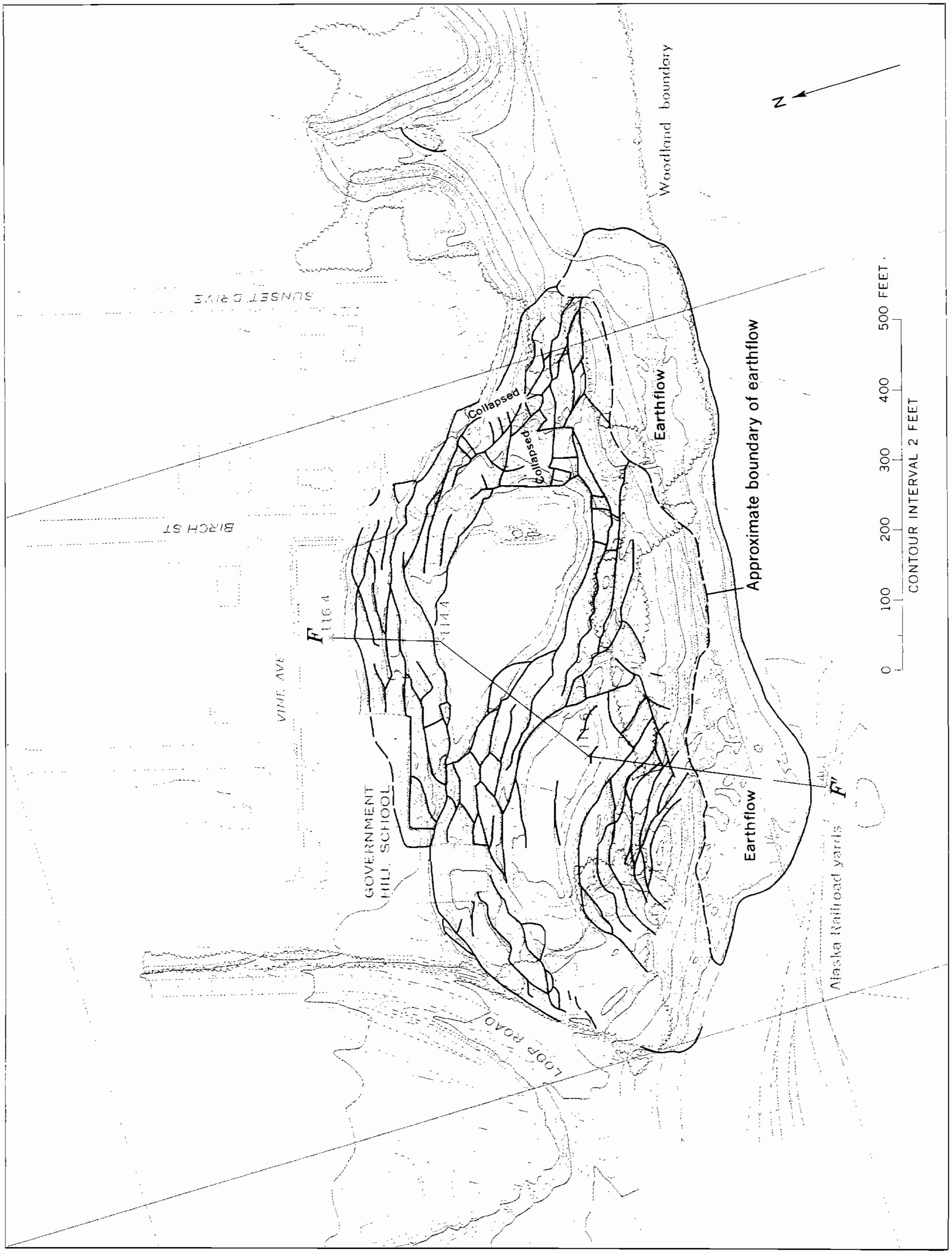
36.—Wreckage of Government Hill School as viewed from the playground, looking west. Graben in foreground is about 12 feet deep. Note undamaged water tower.

medial graben was 14 to 16 feet deep, and the inner graben averaged 12 to 14 feet deep. The intervening horsts glided laterally with slight vertical displacement.

Lateral displacements on the slide varied greatly from place to place inasmuch as the movement of the outermost horst block must have equalled the sum of movements of all blocks behind it plus its own differential movement. Although precise figures for displacement are unavailable, careful comparisons of prequake and postquake aerial photographs give reasonable approximations. The south wing of the elementary school shifted laterally 6 to 7 feet to the southwest. A house at the end of Birch Street, on a horst detached from the main bluff, but otherwise little disturbed, moved about 35 feet southwestward. On the playground south of the school building and across a graben from the building, a children's slide and whirligig moved about 35 feet southwestward. South of the whirligig across another graben, an old concrete blockhouse moved about 65 feet. The outermost points on the toe of the slide moved as much as 150 feet; these points, however, moved partly by flowage.



37.—Geologic section through Government Hill slide. Boring data from Shannon and Wilson, Inc. (1964).



Graben indicated by shading

Topography sterocompiled by J. R. Helm and Gaylord Johansen, U.S. Geological Survey

38.—Map of Government Hill slide, Anchorage, Alaska. Many small fractures have been omitted. Graben areas indicated by shading. Compare with aerial photograph, figure 35.

39.—Homes devastated by Turnagain Heights slide; deep within slide area, upper; at main scarp, lower. About 75 homes were destroyed.

Surface inspection of the bluff line and subsurface studies by the Corps of Engineers indicate that the top of the Bootlegger Cove Clay is somewhat uneven in the Government Hill slide area but averages about 70 feet above sea level (40 to 55 feet below the original ground surface of the bluff). Overlying the clay is outwash sand and pebble gravel.

Shear strength profiles of the Bootlegger Cove Clay measured by Shannon and Wilson, Inc. (1964, p. 94) show the characteristic weak zone, in which static strength ranges from about 0.35 to 0.5 tsf, separated by stiffer clay zones above and below. Scattered throughout the section are lenses of sand and silt. In general, the strength profiles show a gradual decrease in strength to a depth of 85 to 90 feet (altitude 25 to 30 feet) followed by a gradual increase in strength down to the lowest depth tested. The top of the weak zone (shear strength less than 0.5 tsf) generally lay at an altitude of about 40 feet. Most sensitivities were relatively low, perhaps owing partly to consolidation of the clay after remolding, but some were within the range of 20 to 40.

Most of the clay contained a percentage of water less than the liquid limit, but some clay—particularly at critical depths near the indicated depth of failure—had natural water contents equal to or greater than the liquid limit. Thus, the zone of lowest strength and highest sensitivity in the clay coincided with a zone of critically high water-plasticity ratios.

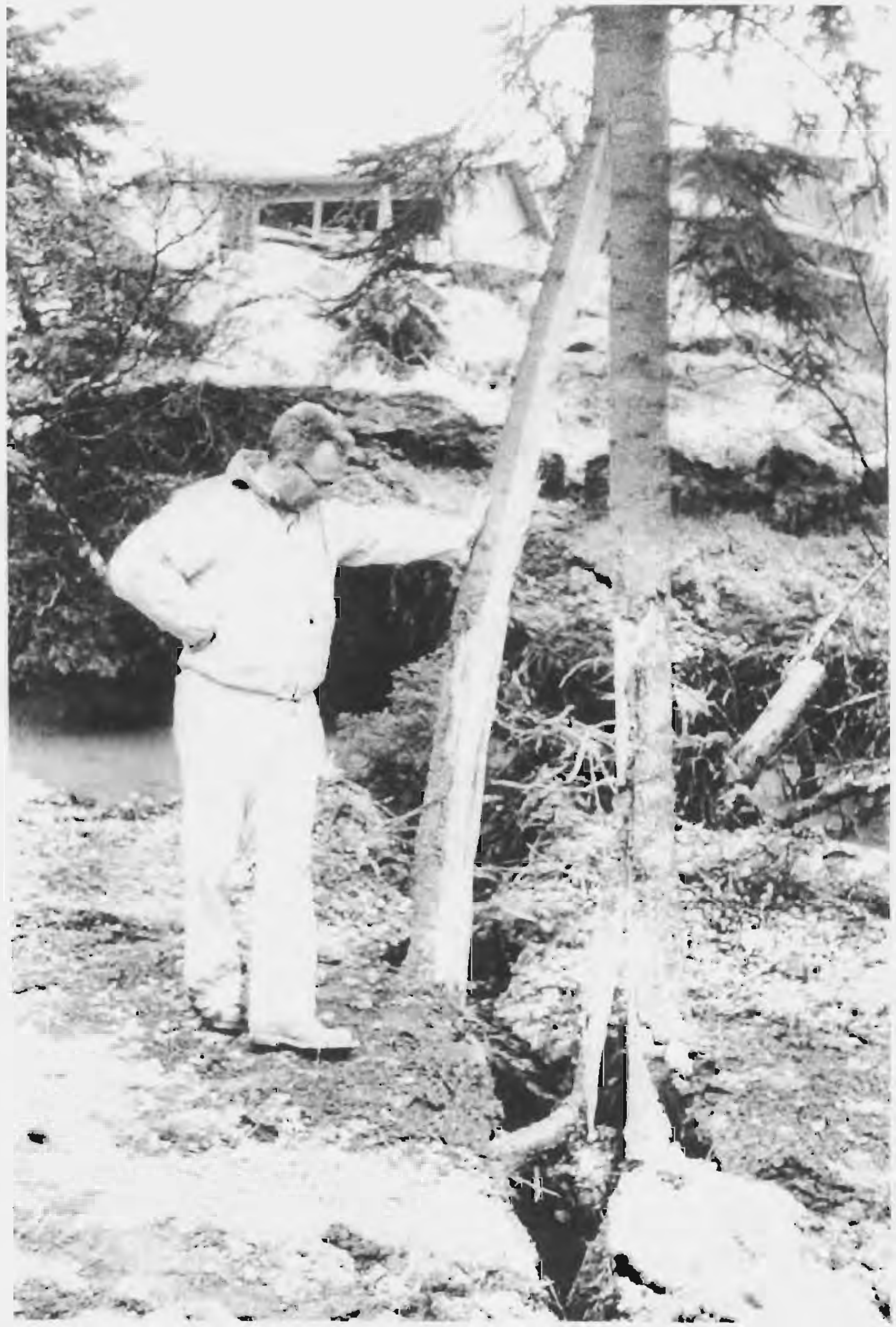
Failure appears to have occurred somewhere between altitudes of about 20 to 40 feet—depths somewhere between 70 and 90 feet below the prequake bluff line. This depth is very close to the top of the weak zone of the clay, and is near the altitude (40 feet) of the flat at the base of the bluff in The Alaska Railroad yards. It also coincides closely with the minimum depth of failure estimated according to the graben rule by using a horizontal translation of 35 feet divided into a graben sectional area of 2,500 square feet.

Several independent factors operating in concert helped precipitate failure. These factors included low shear strength, high sensitivity, and water contents exceeding liquid limits, combined at a depth where static shear stress was critically high, and where added stress conditions were introduced by the accelerations of the earthquake. An additional factor—artificial modification of the toe of the bluff by excavations at the base of the slope—may have contributed to failure by increasing the static shear stress on the slope. Aerial photographs taken in 1962 show plainly that the slope had been modified artificially prior to the earthquake. Shannon and Wilson, Inc. (1964, pl. 13.3), also presented clear evidence that the face of the slope had been modified by removal of material from the toe.

Turnagain Heights Slide

The Turnagain Heights slide was the largest, most complex, and physiographically most devastating landslide in the Anchorage area (figs. 39, 40; pl. 1). It extended west to east along the bluff line about 8,600 feet. Its maximum headward retrogression from the bluff was about 1,200 feet; its average retrogression into the heavily populated residential section of Turnagain Heights, where 75 homes reportedly were destroyed (fig. 39), was about 500 feet. A total area of about 130 acres was completely devastated by displacements that broke the ground into countless deranged blocks, collapsed and tilted at all odd angles. The ground surface within the slide area behind the prequake bluff line was lowered an average of about 35 feet below the old prequake level. The volume of earth within the slide was about 12½ million cubic yards.

Lateral spreading extended the slide seaward, as the leading edge glided down the mudflat into Knik Arm beyond the low tide line. The leading edge of the slide extended farther out beyond the old bluff line than the head of the slide regressed behind; in most places it reached much farther, even twice as far, before passing below tidewater. Maximum lateral slippage exceeded 2,000 feet. Thus, the



40.—Tree trunk split by tension fracture, Turnagain Heights slide. Many trees were similarly damaged because their roots were firmly embedded in the frozen ground.

landslide, on coming to rest, occupied an area considerably more than twice as large as the original undisturbed area.

As seen in plan, the Turnagain Heights slide was composite. It consisted of two main lobes: a West Turnagain lobe and an East Turnagain lobe, each in front of a separate headwall. Each lobe probably started as a separate landslide, but the two lobes merged laterally at a northward-projecting salient that was spared of sliding on the new bluff line. This salient centered on the mouth of

41.—Furrowed, slickensided clay ridge, Turnagain Heights slide. Ridge is about 20 feet high. Tilted collapsed block at left. Compare with figure 42.

Hood Creek and extended southward along the Hood Creek ravine. For a distance of 150 to 300 feet from the creek this shallow ravine formed an immobile, apparently stable, crack-free buttress that projected into the heart of the slide and divided the slide into east and west counterparts (pl. 1).

Each lobe removed about 4,300 feet of bluff line. The West Turnagain lobe, however, retrogressed headward much farther and extended farther seaward than the East Turnagain lobe, but the fractured area behind the East Turnagain lobe was much broader and more intensely broken than the area behind the West Turnagain lobe. If the East Turnagain lobe had retrogressed to the limit of intensive fracturing, it would have become a nearly exact counterpart of the West Turnagain lobe. Apparently the West Turnagain lobe had approached a state of dynamic stability by the time shaking stopped. In other words, the lobe probably had retrogressed to a point of near equilibrium under the dynamic conditions of the time; it probably would not have reached much farther back if shaking had continued. The ground behind the East Turnagain lobe, on the other hand, was left in a state of precarious equilibrium when shaking ceased; it undoubtedly would have retrogressed much farther—possibly as far as Northern Lights Boulevard—if shaking had continued.

In addition to the slide proper, hundreds of tension fractures (fig. 40; pl. 1) opened behind the head of the slide—as far away as 2,200 feet from the main scarp of the East Turnagain lobe (Engineering Geology Evaluation Group, 1964, pl. 8b). These fractures were disposed concentrically about the two main centers of regression—one south of the East Turnagain lobe between Fish Creek and Hood Creek, and one south of the West Turnagain lobe west of Hood Creek. Besides causing great structural damage to housing in the Turnagain Heights subdivision, these fractures totally disrupted all underground utilities and seriously damaged streets and curbings. Even more ominous, they outlined a potential headward expansion of the landslide that was forestalled only by the cessation of ground shaking. Continued shaking undoubtedly would have involved much more property in the landslide.

The ground behind the East Turnagain lobe had in fact begun to move, and the displacement increased toward the new bluff line. Preliminary studies reported by the Engineering Geology Evaluation Group (1964, p. 17) indicated that the area along Turnagain Parkway between the head of the slide and Northern Lights Boulevard was lengthened by more than 3 feet. Differential movement in this distance was taken up by the opening of tension cracks. There also was subsidence of 6 inches or more within a city block of the new bluff line. Detailed measurements by Dickinson Oswald and Associates, Anchorage, for the Engineering Department, City of Anchorage (summarized by Shannon and Wilson, Inc., 1964, p. 64) indicate that movement occurred chiefly before resurveying began—probably during the strong-motion period of the quake—and that no appreciable movement occurred during a 3-month period following the quake when detailed measurements were still being made.

It thus appears that the weak zone in the Bootlegger Cove Clay under the intensely fractured area between the slide proper and Northern Lights Boulevard did indeed fail, but sufficient shear resistance remained in the block after shaking stopped to prevent further sliding. The area at that time was buttressed, of course, by the passive pressure of the slide debris in front of it.

At the east margin of the slide, the bluff line facing Fish Creek held firm despite a sharp local re-

42.—Sharp-crested clay ridge, Turnagain Heights slide. Collapsed blocks tilted toward ridge on both flanks. Note horizontal stratification in ridge. Compare with figure 41.

lief of 40 feet or more. The slide, in fact, moved northwest almost diametrically away from Fish Creek, leaving behind a somewhat cracked but otherwise little-damaged salient or riblike buttress 70 to 400 feet wide projecting north along Loussac Drive. Conditions at Fish Creek, thus, were comparable to those at Hood Creek at the far side of the East Turnagain lobe.

One can but conclude that the shallow valleys of Fish Creek and Hood Creek were instrumental in limiting ground failure in the Turnagain Heights slide. Their function is unclear, but they may have acted as natural sumps that partly dewatered, and hence stabilized, the adjacent ground in the slopes of the ravines and in the bordering uplands some distance back from the ravines. This supposition gains support from subsurface data at Fish Creek where a boring by the Corps of Engineers penetrated clay of high strength, low sensitivity, and low water-plasticity ratios at the critical depths where failure had occurred nearby. The boring nearest Hood Creek (200 feet southeast of the creek) showed high sensitivity and low shear strength at the critical depth, but it showed high strength and low sensitivity immediately above. Fish Creek (altitude near mean sea level) had eroded its channel to the depth of the sensitive zone, but Hood Creek—a smaller stream—had not. But by cutting through the overlying clays, Hood Creek must have effectively lowered the water table in the adjacent ground, and in so doing further stiffened the upper clays. A drill hole closer to Hood Creek might have provided more conclusive data.

Hundreds of sharp-crested clay ridges alternating with collapsed troughs, and oriented normal to the direction of slippage, distinguished the disruption pattern of the Turnagain Heights slide from all other slides at Anchorage. The ridges, however, were exact homologs of the horsts of the Government Hill slide; the troughs were exact homologs of the grabens. The chief distinction of the Turnagain Heights slide, therefore, other than its size, was the utter totality of its disruption. Its pattern is well portrayed by the topographic map (pl. 1) and by the strip map and section (pl. 2).

Most of the clay ridges ranged in height from about 10 to 15 feet, but a few were more than 20 feet high. They were as much as 300 feet long and were spaced 50 to 150 feet apart. Their steep sides, which sloped 60° to 70°, were furrowed and grooved by slippage of one surface against another (figs. 41, 42). On the average, the ridges were sharper crested and more closely spaced in the West Turnagain lobe than in the East Turnagain lobe.

Stratification was greatly disturbed in the collapsed areas between ridges, but it was little disturbed in the ridges themselves; the ridges were displaced virtually without rotation. There was, moreover, little vertical displacement as the ridges glided toward Knik Arm. Detailed studies along the disrupted seaward projection of Turnagain Parkway (pl. 2) indicated that clay ridges which were displaced as much as 300 feet laterally were reduced only about 12 feet vertically. Some of this reduction may have been due to attrition at the slip surface or to flowage in the sensitized clay below, but most of it probably was due to a seaward slope on the slip surface itself. This slope, then, must have had an inclination of about 4 percent.

The complete disruption of the ground surface within the Turnagain Heights slide may have been due to several factors in combination—including the shallow depth to the zone of failure—but the unhindered movement of the slide down the wet mudflat toward Knik Arm certainly was paramount. In every other translatory slide at Anchorage, slippage was resisted by dry or frozen ground at the toe and by an abrupt flattening of the ground slope at the foot of the slide below the bluff line. At Turnagain Heights, however, the slide broke away from the bluff within the intertidal zone and slid out directly onto the sloping tidal muds, which themselves were wet and sensitive. At the flanks of the slide, the tidal flat slopes about 3°, or 5 to 6 percent; before failure it probably had a comparable slope in front of the slide also.

Shear resistance of dry or frozen ground in front of a given slide must have had a natural buttressing effect that in turn must have been a factor in the resistance of the slide to slippage. All the translatory slides at Anchorage possessed such natural buttresses except the Turnagain Heights slide. The ground at the top of the intertidal zone, where the Turnagain Heights slide sheared to the surface, had previously had little opportunity to desiccate, and the thickness of the frozen layer must have been minimal—if indeed, the ground there was frozen at all. Reconnaissance along the foot of the bluff several years prior to the earthquake, moreover, disclosed a prevalence of saturation, slumpage, and flowage in the clay at the foot of the bluff. More or less continuous slumpage, abetted by wave and tidal action which constantly removed the accumulated debris, prevented drying of the naturally wet clay and kept the bluff line in a precarious state of repose.

Part of the slip surface of the slide was left as a window near the west end of the slide, where the overlying debris slid free out to tidewater (fig. 43). Sliding, however, was not confined to this one surface. On the contrary, there is every indication that blocks slid out one on another at higher levels in the same vicinity. In other places, failure may have occurred mainly at some other level. But this surface was plainly an important locus of shearing in this particular part of the landslide.

Altogether, about a quarter of an acre of the slip surface was uncovered. The surface was mantled here and there by small pyramidal mounds and blobs of clay. Long furrows and welts, oriented in the direction of slippage (N. 3° W.) and extending the length of the exposure, were well preserved 6 weeks after the earthquake, despite desiccation and cracking.

The surface itself passed below highest tide at its seaward margin. It sloped about 15 feet per 100 feet, on the average, but it was convex; the outer edge was steeper than the inner edge. The exposed surface was precisely on the projection of the old prequake shoreline, so there is no doubt as to where the slide sheared off with reference to the prequake topography; it must have sheared to the ground surface at tidewater at the very foot of the bluff.

Mounds of clay left on the slip surface, showing evidence of having been overridden themselves, were grooved and slickensided on their tops and flanks. Their sides sloped steeply down to the slip surface on which they rested, angles of 40° to 60°.

In places the slip surface was overlain by blocks of peat and forest duff. Frozen at the time of the quake, these blocks must have been lowered onto the slip surface as the intervening incompetent clay slid and flowed out from under them. The conclusion seems inescapable that the clay overlying the slip surface glided seaward primarily under the influence of gravity; a window to the slip surface could have been exposed in no other way.

At the extreme east end of the slide the slip surface did not break through to the ground surface. Instead, it died out laterally, as the thrust of the sliding mass was taken up by a large pressure ridge in the tidal silts just below the foot of the bluff (fig. 44). This ridge cut dramatically at an oblique angle across an old riprap embankment placed along the beach at the foot of the bluff—years before the earthquake—to retard marine erosion. The embankment was arched up into an anticline where the pressure ridge passed beneath it. The ridge was about 700 feet long, as much as 50 feet wide, and 10 to 15 feet high on the landward side. On the seaward side it was higher and was modified by subsidiary slumping on its oversteepened flank. Just west of the pressure ridge the translatory movement of the landslide was greater, and the slide itself was correspondingly more complex. The tidal silts failed, and the overriding slide blocks glided far out into the Knik Arm.

Subsurface explorations of the Turnagain Heights slide were made by the Corps of Engineers after an initial investigation was started by the Engineering Geology Evaluation Group. Undisturbed samples for field and labo-

43.—Slip surface of Turnagain Heights slide exposed near west margin of slide. Furrows and welts still well preserved despite desiccation and weathering. Point Woronzof and Knik Arm in distance.

ratory tests were collected from borings at 42 localities within the slide and adjacent to it. In some places the low strength of the clay made it difficult to collect undisturbed samples; some of this clay may have lost strength by remolding.

Surface and subsurface observations showed that the Bootlegger Cove Clay is mantled by outwash sand of somewhat varied thickness; the sand is about 20 feet thick at the east end of the slide and tapers nearly to zero at the west end. The clay is as much as 100 feet thick. Shear-strength profiles showed that the clay diminished gradually in strength from about 1 tsf at the top of the clay to about 0.25 tsf at an altitude of 15 to 20 feet above sea level; it then increased gradually in strength to about 0.6 tsf at some depth (generally about 30 feet) below sea level. Values of minimum strength in some borings were much less than 0.25 tsf, again perhaps because of partial remolding of disturbed sensitive clay.

The depth of lowest strength commonly coincided with the depth of highest sensitivity and highest water-plasticity ratios. Throughout much of the Turnagain Heights area the clay at lowest shear strength contained water in excess of its liquid limit and had sensitivities greater than 40. Remolded, this clay had a shear strength of about 0.02 tsf (Shannon and Wilson, Inc., 1964, p. 65).

Thus, the depth of failure probably was 15 to 20 feet above sea level, at or near the top of the zone of lowest shear strength. This altitude also was the height of the knick point or slope break at the base of the bluff before the earthquake—the position at which the slide evidently broke to the surface and glided seaward down the mudflat.

The mode of failure of the Turnagain Heights slide differs in degree rather than in kind from the other translatory slides of the Anchorage area. Some idea of the sequence of failure can be had by examining different parts of the slide where movements had gone to different stages of completion by the time the earthquake stopped. At the extreme east end of the slide near the mouth of Fish Creek, for example, the failure compared morphologically with failure at the Native Hospital slide and near the east flank of the Fourth Avenue slide. A little farther west where the deterioration was somewhat greater, the failure was comparable to that in the Government Hill slide.

As stated previously, the slide was composite in that it was formed by the lateral merger of two main lobes. Within each of the main lobes there were subsidiary lobes that apparently formed independently at the bluff line, then merged laterally as they retrogressed, and finally slid in unison toward Knik Arm. They moved different distances, and the leading edge of their combined mass at tidewater therefore had a scalloped outline after motion ceased. (See map, pl. 1.) In the East Turnagain lobe, they also retrogressed unevenly into the bluff, making the new bluff line very irregular.

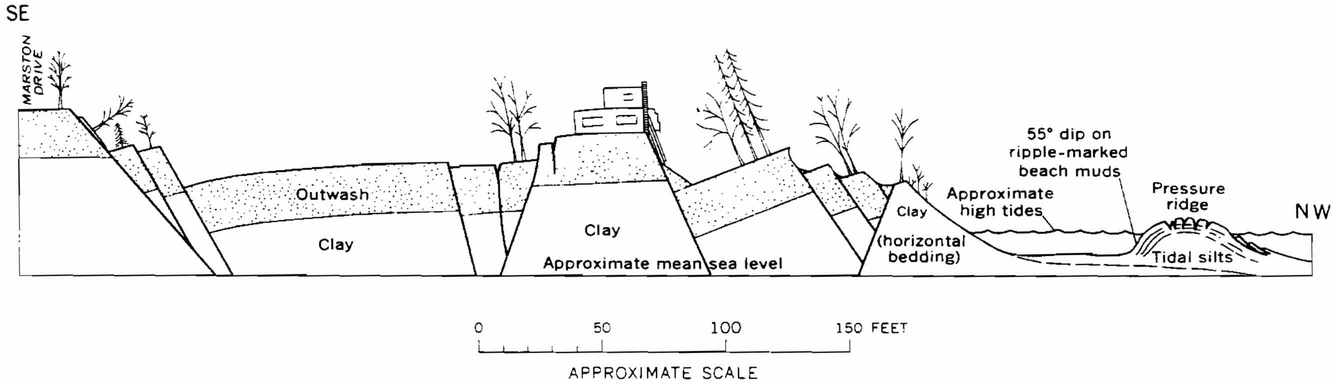
Interference or crowding between subsidiary lobes was accompanied by complex shifting and wrenching of blocks. Such movements may have caused lateral components of displacement of as much as 150 feet in a direction parallel to the old bluff line—as indicated by the postquake positions of objects such as houses transported on the slide toward Knik Arm (Engineering Geology Evaluation Group, 1964, pl. 1). As the blocks pushed and shoved in response to the overall motion of the landslide, they also were lurched sharply from side to side by the vibratory motion of the earthquake. This lurching was mirrored by abrupt deflections in furrows and slickensides preserved on the sides of clay ridges.

Some time after the onset of the earthquake, the bluff line is visualized as having begun to fail along a broad front, but mainly from one or more centers in each main lobe of the slide. Blocks then began to break off headward and to the flanks of each center. Eyewitness accounts indicated that 2 minutes or more elapsed from the start of the quake before the bluff began to give way (Shannon and Wilson, Inc., 1964, p. 64) and that movement continued for some time after the earthquake had subsided. A particularly lucid and informative report was given by Mr. Brooke Marston to Mr. E. R. Bush (Grantz, Plafker, and Kachadoorian, 1964, p. 14):

> I was driving my automobile westward on McCollie Avenue when the earthquake occurred. I immediately stopped my automobile and waited until the quake subsided. It appeared to me that the car was rocking from north to south. It rocked so violently that I nearly became seasick. From my car I could observe an earth crack alined north-south and opening and closing from east to west. As soon as the quake subsided I proceeded to drive westward to the corner of McCollie Avenue and Turnagain Parkway. After turning right on Turnagain Parkway and driving approximately 180 feet north, I realized the bluff was gone north of my driveway, which paralleled the bluff in an east-west direction. I got out of the car, ran northward toward my driveway, and then saw that the bluff had broken back approximately 300 feet southward from its original edge. Additional slumping of the bluff caused me to return to my car and back southward approximately 180 feet to the corner of McCollie and Turnagain Parkway. After I stopped at this point, the bluff continued to slowly slide northward as I continued to back my auto southward on Turnagain Parkway. The bluff slowly broke away until the corner of Turnagain Parkway and McCollie had slumped northward.
>
> It is my impression that the Turnagain Bluff area slumped northward in segments and that much of the southward receding of the bluff occurred after the major earthquake had subsided.

From Mr. Marston's account, it is clear that large-scale ground displacements were still in progress after the earthquake had stopped. After the quake, during the time that Mr. Marston drove down McCollie Avenue, turned north down Turnagain Parkway, got out of his car and returned, then backed south along Turnagain Parkway, stopped, and resumed backing—all this after the shaking had stopped—the bluff continued to regress southward. It regressed at least 200 feet, and the elapsed time must have been a full minute or more. If Mr. Marston had been driving 5 miles per hour, for example, including backing but not counting stops, he would have needed about 55 seconds to go 400 feet. He probably drove appreciably farther than 400 feet, however, and his stops must have taken several seconds each.

Small-scale slumping continued for several days after the quake as oversteepened slopes along the new bluff line continued to sluff away and frozen blocks of sand and forest duff began to thaw. One house, left precariously balanced, but not severely damaged on the rim of the bluff at Chilligan Drive, toppled and collapsed several days later. So far as is



44.—Sketch section through eastern part of Turnagain Heights slide. Note pressure ridge, upright blocks at center and to left of pressure ridge, and tilted collapsed blocks between.

known, aftershocks did not cause any additional sliding.

Failure at Turnagain Heights probably occurred near the top of the lowest-strength zone of the Bootlegger Cove Clay, at an altitude of 15 to 20 feet above sea level. Inasmuch as the top of this zone was a subhorizontal surface, the failure itself must have been along a subhorizontal surface also, and the initial movements of the slide must have been translatory. Because the clay contained water in excess of its liquid limit, because the strength of the clay under pulsating or vibratory stress was appreciably less than under static stress, and because the zone of lowest strength coincided with the zone of highest sensitivity, the face and crown of the bluff were—in effect—afloat on a liquefied layer of viscous clay. The bluff then began to glide slowly seaward under the influence of gravity, leaving a gaping crack in its wake. The presence of open cracks—denoting release of tension—is significant; nearly all eyewitness reports allude to them.

Earth materials strained under horizontally directed tension commonly have two sets of tension fractures: one set antithetical to the other such that alternate fractures dip toward and away from one another and both sets are oriented normal to the direction of maximum tension. As the slide began to spread, countless tension fractures disrupted its surface and the surface of the ground behind the slide; after movement stopped, incipient grabens were preserved in the subdivision behind the slide.

Along antithetical fractures dipping south, the unsupported hanging wall generally collapsed toward the north; along fractures dipping north, the hanging wall collapsed toward the south. At a given clay ridge bounded by opposed fractures dipping away from the ridge, adjacent collapsed blocks commonly were tilted toward the ridge on both flanks (figs. 41, 42). Farther from the ridge in the bottom of the adjoining trough or graben, collapsed blocks were tilted into helter-skelter attitudes.

As each new block pulled away from the bluff, the new bluff line was left unsupported on the seaward side. Tension fractures reduced support on the landward side. Blocks were pulled away successively by gravity as long as the shear resistance at the slip surface was exceeded by the force of gravity plus the accelerations of the earthquake. When the earthquake stopped, the force of gravity alone was sufficient to cause large-scale failure for some time afterward. And in the hanging wall of each tension fracture, subordinate blocks collapsed and toppled under their own weight—the entire mass at the same time gliding slowly toward Knik Arm.

After the earthquake, the outermost slide block of the original bluff line was preserved to view only at the extreme east end of the slide (fig. 44) where the thrust of the toe was countered by a pressure ridge. Elsewhere along the front of the slide, where movement was much greater, the toe of the slide and the old bluff line passed beneath the waves of Knik Arm. Only at the east end of the slide, therefore, where the movement did not go to completion, is it possible to visualize what happened at the outset when the bluff first began to fail. Rotational movement at the east end of the slide is ruled out; stratification in the Bootlegger Cove Clay was undisturbed despite lateral shifting of several feet, and trees on the slide block remained perfectly upright. Rotational effects came into play behind the block, however, where the overhanging trailing edge of the block, lacking support, tilted backward and collapsed.

Some idea of the complexity of the disruption is presented in plates 1 and 2 and in a restored section through the slide prepared for the Corps of Engineers (*in* Shannon and Wilson, Inc., 1964, pl. 10.8). This restoration is in general agreement with figure 43 and with the steps previously outlined, although it emphasizes the part played by rotation in the process and minimizes the effects of translatory motion. This restoration also contains minor geometric inconsistencies. Contrary to this view, translatory motion under gravity is here envisaged as the primary operative mechanism of landsliding, not only at Turnagain Heights but at the other major slides in Anchorage as well. Rotation, as well as flowage, must have occurred in some degree in all the slides, but rotation is viewed as a subordinate process not germane to the failure of the ground, even though it certainly was significant in the disruption of the ground surface.

LANDSLIDING PRIOR TO THE MARCH 27 EARTHQUAKE

Geologic evidence indicates that landslides similar to those set off by the March 27 earthquake have occurred previously in the Anchorage area. Most of these slides predated the settlement of Anchorage, and it is not known, therefore, whether or not most of them were triggered by earthquakes; the morphology of some, by analogy with the March 27 slides, suggests that earthquakes have had a part. Some inferences as to timing suggest the same. The earthquake of October 3, 1954, clearly triggered slides along The Alaska Railroad near Potter Hill.

Evidence of old landslides in certain areas of Anchorage has been cited elsewhere in this report (p. A31, A49). The evidence is here reiterated, together with evidence of sliding elsewhere. Miller and Dobrovolny (1959) recognized many areas of past landsliding, showing them on their map (pl. 1) as "areas covered by landslides, slumps, or flows," with the prognosis that "shocks, such as those associated with earthquakes, will start moving material that under most conditions is stable" (p. 104).

Old landslides, like recent ones, were concentrated along bluff lines where topographic relief caused high static-shearing stress on the underlying soil. Most old slides have the form of recessed alcoves with uneven terracelike floors. Small slides may have the form of steplike breaks on the sides of the bluffs. Some slumps have left lobate accumulations of earth at the foot of the bluffs. The fact that an area has slid previously apparently is no basis for predicting that it will or will not slide again. Many former landslides were stable during the March 27 earthquake, but in some places new slides were superimposed directly on old ones.

Compelling evidence of previous translatory sliding was found at the site of the Native Hospital slide, where a well-preserved graben was truncated by the slide of March 27. The old slide block and its graben are shown well in figure 32. Birch trees growing on the scarp of the old graben average about 7 inches in trunk diameter at breast height. Trees of that size on well-drained ground in the Anchorage area average about 60 years in age (Reed and Harms, 1956, p. 241). Five years or more may have been required after sliding to establish forest growth on the raw gravel. About 65 years, therefore, is the probable minimum age of the slide, an age which dates to about the turn of the century. Many earthquakes, some of them severe, occurred at about that time in southern Alaska (Davis and Echols, 1962), and one of them might have triggered landslides at Anchorage. The great Yakutat earthquake of 1899 was felt in much of southern Alaska, including points more distant that Anchorage (Tarr and Martin, 1912, p. 68 and pl. 33), but whether the quake's intensity was sufficient to trigger landslides as far away as Anchorage is uncertain.

Remnants of old slides are abundant elsewhere along the bluffs of Ship Creek. Between the Native Hospital slide area and the Fourth Avenue slide area, the Alaska State Highway Department facilities are located in an old slide area that was partly reactivated on March 27. The alcovelike area centered at the city parking lot north of Fourth Avenue, between C and E Streets, probably also is the scar of an old landslide (Shannon and Wilson, Inc., 1964, p. 39; R. M. Waller, written commun., 1965). The area coincides almost exactly with the Fourth Avenue slide of March 27, except that the Fourth Avenue slide retrogressed farther into the bluff. The salient west of this slide partly bounded by Christensen Drive and Second Avenue appears to be a lowered slump block.

On the north side of Ship Creek, the recess just west of the destroyed Government Hill School may be an old slide scar. Points farther west along the bluff and north around the bend above the Port of Anchorage have many of the topographic markings of landslides, including alcoves, step terraces, and grabenlike depressions. The bluff at Cairn Point below the Elmendorf Moraine has a long history of minor slumping. Evidence of prior slumping along Bluff Road east of Government Hill School has been noted previously.

Along the bluff above Knik Arm between L Street and O Street, scallop-shaped recesses and terracelike benches below the crown of the bluff probably are old slide blocks. A remarkable channellike trench that trends diagonally southwest from the vicinity of R Street and Tenth Avenue toward S Street and Eleventh Avenue probably is a remnant of an old graben. Trees in this trench are comparable in size to those in the old graben near the Alaska Native Hospital.

Farther southwest near the mouth of Chester Creek, the bench along Bootlegger Cove Drive probably is an old slide block, and due east from there the broad recess centered near Inlet View School probably is an old slide area also (fig. 27). This feature is very subdued and probably is very old.

The bluff line along the south side of Chester Creek from the mouth of the creek discontinuously as far east as Rogers Park shows evidence of past slumping. Terracelike features lacking continuity along the valley margin, especially those deeply recessed into alcovelike niches, or those having irregular floors—particularly back-sloping floors, probably are slump blocks. On March 27 a small slump occurred in the bluff just north of West High School. Forest Park Golf Course between the mouths of Chester and Fish Creeks appears to be situated in an area of old landslides.

West from Fish Creek to Point Woronzof the bluff line has had a long history of slumping. There is no indication of previous sliding comparable to the disastrous Turnagain Heights slide, but small-scale sluffing away and localized slumping were continuing problems; the long-range effect was a slow but relentless sapping away of the bluff line.

The long bluff line south of Anchorage facing Turnagain Arm showed little evidence of landsliding prior to the March 27 earthquake except at Potter Hill and Point Campbell. Slumping at Potter Hill has been a recurrent problem, only in part caused by earthquakes. At Point Campbell, intermittent debris slides and sand runs have been caused by ordinary wave and current erosion at the foot of the bluff—processes unrelated to earthquakes.

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CLARIFICATIONS

Geology Vol. Page	USGS Page	Column	Line	Remarks
347	A 59	1	19	Plate 1 is Plate E in Part B, the accompanying map case
349	A 61	3	20	Plate 2 is Plate F in Part B, the accompanying map case

Appendix D

Holcombe, T.L.; Taylor, L.A.; Reid, D.F.; Warren, J.S.; Vincent, P.A.; and Herdendorf, C.E. (2003). "Revised Lake Erie Postglacial Lake Level History Based on New Detailed Bathymetry," *Journal of Great Lakes Research* **29**, 681-704.

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Revised Lake Erie Postglacial Lake Level History Based on New Detailed Bathymetry

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ABSTRACT. *Holocene lake level history and paleogeography of Lake Erie are re-interpreted with the aid of new bathymetry, existing water budget data, and published information. Morphology and elevation of present and former shoreline features (sand ridges, forelands, spits, bars, and fans) record the water level at which they were formed. Of eighteen such features observed in Lake Erie, six occur nearshore and were formed at or near present lake level, and twelve features apparently formed at lower lake levels. It seems likely that lake level fell below the level of the outlet sill during the 9–6 ka climate optimum, when warmer and drier conditions prevailed. During such times lake level likely rose and fell as controlled by the water budget, within a window of constraint imposed by increases and decreases in evaporation, which would have varied directly with lake surface area. Near Buffalo, possible shoreline features occurring 3–6 km offshore at depths of 9–12 m could have formed at lower lake levels. Annual water volumes in each term of the water budget, (runoff, precipitation, and evaporation) are large relative to the volumetric capacity of Lake Erie itself. Such events as introduction of even a modest amount of upper Great Lakes water, or the onset of cooler and less dry climate conditions, could cause significant, rapid, lake level rise. Schematic reconstructions illustrate changing paleogeography and a Holocene lake level history which has varied with: blocking/ unblocking of outlet sills; erosion of outlet sills; distance from outlet sills; differential isostatic rebound; upper Great Lakes drainage flowing into or bypassing the lake; and climate-driven water budget of the Lake Erie drainage basin.*

INDEX WORDS: *Bathymetry, Holocene, Lake Erie, lake floor features, lake level, lake surface area, lake water volume, paleogeography, water budget.*

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INTRODUCTION

Lake Erie, shallowest and southernmost of the Great Lakes, has had the longest and in some respects the most complex postglacial lake history of any of the Great Lakes. Relatively modest rises or falls in lake level caused major changes in the paleogeography of the lake basin. Holocene changes in lake level at a given location were controlled by a combination of the following factors: amount of isostatic rebound, distance from the outlet sill, level of the outlet sill, lowering/ erosion of the outlet sill, volume of water passing through the lake, lake surface area, and paleoclimate. New evidence pertaining to lake levels comes from recently compiled, detailed bathymetry (National Geophysical Data Center 1998) which reveals lake-floor features indicative of former, but now inundated, shorelines. Previous work has addressed the history of postglacial lake levels in broad outline. Now, with the new evidence provided by the detailed bathymetry, it is appropriate to consider again the Holocene lake-level history of Lake Erie.

In this paper we identify and briefly describe lake floor features which are indicative of present and former lake levels. Computations of lake surface area and water volume are presented, which are more precise than previous computations relying on less detailed shorelines and bathymetry. Aspects of the present lake water budget are examined, with implications for water budgets during an early Holocene low water phase. Aspects of control of lake level by the water budget during times of lowered lake levels are discussed. Finally, a model for the Holocene lake level history of Lake Erie is presented and discussed.

Coakley and Lewis (1985) assembled information from: 50 radiocarbon dates on samples recovered from the lake floor; shoreline elevations of deglacial Lakes Whittlesey, Warren, Algonquin, and others; lake floor geomorphology documenting now drowned shorelines and channels; and miscellaneous other information; and they reconstructed a Holocene history of Lake Erie water levels. Calkin and Feenstra (1985), from a variety of evidence (elevation and tilt of former strandlines, level of downcutting of river channels, ice advances and retreats, chronology of high-standing proglacial lakes), and Barnett (1985), from studies of the geomorphology of southern Ontario, reconstructed lakes and lake levels within the Lake Erie basin during the past 14,000 years. Earlier reconstructions of lake level history were accomplished by Lewis (1969), and Sly and Lewis (1972). Pengelly *et al.* (1997) modeled the history of high-water lake levels of the past 12,500 years based on field information gathered from the Niagara region.

Reconstructions vary as to details and water levels. Some salient features of the models: 1) high lake level stages of proglacial lakes and Lake Algonquin followed by very low lake levels in early Holocene time, 2) isostatic rebound and concomitant gradual rise in lake levels in the early Holocene, 3) stable or slowly rising lake levels in Middle Lake Erie time, 4) rise in lake levels accompanying the Nipissing Rise to slightly higher than at present, and 5) lowering of lake levels to just below present level following the Nipissing Rise event. Calkin and Feenstra (1985) cited evidence for Lake Ypsilanti, a Lake Erie lowstand (low-standing stream gravels near Ann Arbor and Ypsilanti; Kunkle, 1963; depth to channels cutting till beneath the central basin of Lake Erie; Wall, 1968) prior to Lake Whittlesey time during the Mackinaw Interstade. The reconstructions recognize that lake level has been controlled since deglaciation by isostatically rebounding outlet sills of the Niagara River; and that the Nipissing rise accompanied a shift in upper Great Lakes drainage from the North Bay outlet to the Port Huron outlet. Pengelly and associates (1997) cited evidence for a post-Nipissing lowering of lake level 3,600–2,900 ya due to erosion of a former sill in the Niagara River referred to as the Lyell-Johnson Sill.

The Nipissing Rise is generally associated with diversion of upper Great Lakes outflow from the North Bay outlet to the Port Huron outlet in mid-Holocene time. Lewis (1969) and Pengelly *et al.* (1997) discuss evidence for this event and the times of its occurrence and cite the work of previous investigators. Two rising water events are recognized by these and other investigators, Nipissing I, which peaked between 5,500 and 5,000 ya, and Nipissing II, which peaked between 4,500 and 3,800 ya. There was a significant period (6,000 to 4,000 ya) when outflow of upper Great Lakes water was shared between three outlets—Chicago, North Bay, and Port Huron—in varying amounts, before final transfer of a majority of upper Great Lakes water to the Port Huron outlet. Recently, evidence from carbon and oxygen isotope ratios from molluscs and ostracods, contained in Lake Erie sediment cores, suggested a single and permanent water-change event (Nipissing II?) occurring about 3,600–4,600 ya (Kempthorne *et al.* 2000).

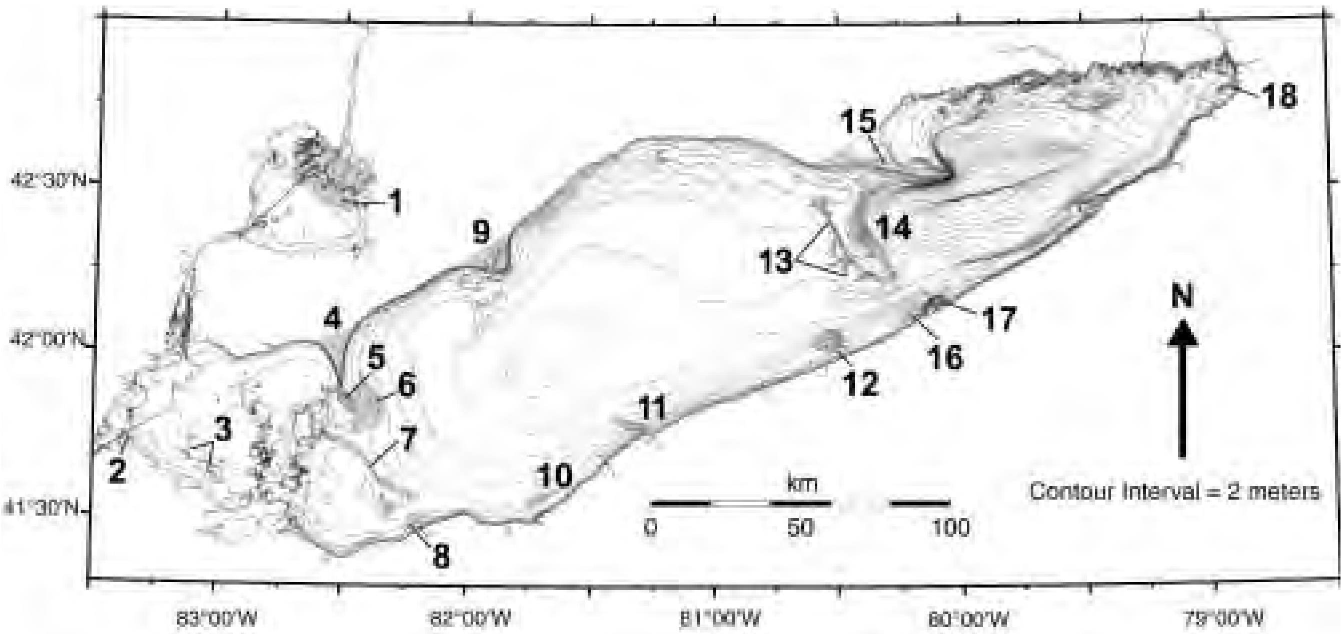


FIG. 1. Lake Erie and Lake St. Clair bathymetric features indicative of present and former lake levels. Numbered features, including bars, spits, banks, sand ridges, forelands, and fans, are discussed in the text.

LAKE FLOOR FEATURES INDICATIVE OF PRESENT AND FORMER LAKE LEVELS

Sand ridges, forelands, spits, bars, and fans are Holocene depositional features which formed in the shore zone and record lake level at the time of their formation. Therefore their occurrence at elevations other than present records the levels of former shorelines. Pre-existing topographic features such as moraines and drumlins, which date from the last period of glaciation, if later eroded in the shore zone, may also record lake level.

Of the eighteen such features recording present or former shorelines identified in Lake Erie and Lake St. Clair, based on the detailed bathymetry, six are sites of sediment deposition at present lake level. The remaining twelve features are interpreted as having formed beneath the modern lake surface during the Holocene at lower lake levels, and later abandoned. Two of these features are thought to have formed at lower lake levels, but alternative interpretations are possible, as noted.

Each of the eighteen features thought to record present and former shorelines are numbered 1–18 and are shown in Figure 1. These features are also listed below in the same numerical order. Accompanying notes briefly describe each feature, summarize its historical interpretation, and give a rough estimate of the lake level at which it was formed. For a more detailed description of each feature, refer to the detailed bathymetric map (National Geophysical Data Center 1998). For a more detailed discussion of the geomorphology of western Lake Erie features, refer to Holcombe *et al.* 1997.

St. Clair Delta. The dominant feature of Lake St. Clair, this platform of deltaic deposits and multiple distributary channels is bounded by foreset slopes extending downward from platform depths of less than 1 m to more than 3 m depth. The delta is areally large but the lake basin is only 5–6 m deep. Delta sediments prograde, but aggrade only minimally. A smaller delta of the Clinton River, which extends into Lake St. Clair from the northwest shore, partially coalesces with the St. Clair Delta and is included with this feature. Lake St. Clair is too small to be significantly affected by differential isostatic rebound, therefore the lake has remained near its present level since opening of the Port Huron outlet about 4,000–5,000 ya. A similar large delta did not form off the Detroit River. Lake level at 4,000 ya in Lake Erie was several meters lower than at present in the larger, slightly deeper western basin of Lake Erie.

2) Maumee Spits. An outer spit extends northward and northwestward from the south shore of Maumee Bay, and an inner spit extends southeastward from the north shore of Maumee Bay. These features formed during late Holocene time in 2–3 m of water and are thought to be adjusted to present lake level. The spits are formed of sand and gravel deposits, and the intervening areas contain silt and clay muds (Hartley 1960). The Maumee Spits seem to have formed at a rapid rate. Lake history here is short (about 2,000 years) and water depth is shallow.

3) Reefs Area Spits. Drowned spits extend to the NW and SE away from the bedrock reefs in the western basin of Lake Erie. These features lie at a depth of 4 to 7 m. The relict spits and the area surrounding the bedrock reefs are underlain by sand and gravel at present (Herdendorf and Braidech 1972), in contrast to deeper mud-filled areas of the western basin. Bathymetry shows that the reefs area in the western basin was formerly an eroding headland and site of net longshore divergence at water depths of 4 to 7 m (Holcombe *et al.* 1997). Judging from the water level history discussed below, these features formed around 2,000–3,400 ya, before being inundated and abandoned.

4) Point Pelee Spit. Point Pelee is a symmetrical horn-shaped spit which extends southward into Lake Erie from the Ontario shore and is adjusted to present lake level. Point Pelee Spit is formed of sand brought in from both the west and east by longshore drift. The spit formed in post-Nipissing time (0–4,000 ya) on a morainic foundation (Coakley 1978), at a site of longshore convergence activated when the western basin of Lake Erie was flooded by rising water. Bathymetry and shoreline erosion patterns demonstrate that the Point Pelee spit retreated northward as water level rose.

5) Point Pelee Ridge. Point Pelee Ridge is about 6–7 km long, 3–4 km wide, extends southward from Point Pelee, is capped by complex small-scale topography, and lies at a depth of about 5 m. Point Pelee Ridge is interpreted as a former extension of Point Pelee Spit, abandoned as a shoreline feature perhaps about 2,000 ya when rising water level exceeded a few meters lower than present. The feature was apparently part of a longer larger spit along which sand from the Ontario shore was swept southward and deposited.

6) Point Pelee Fan. Extending to the east of the Point Pelee Ridge is a small fan which crests at 11–12 m below present lake level, is recognizable downslope to a depth of 18 m, and extends southwestward as far as the Pelee-Lorain Ridge. This feature seems likely to have formed when lake levels were 11–12 m below present lake level, earlier than Point Pelee Ridge, and probably during the events of the Nipissing Rise around 4,000–5,400 ya. Morphology of the Point Pelee Fan suggests a point-source of sediment influx, possibly a river draining the western basin, although a more recent origin, with Point Pelee Ridge as a source of sediments, is not ruled out.

7) Pelee-Lorain Ridge. The Pelee-Lorain Ridge rises 1–3 m above the surrounding lake floor and extends 35 km southeastward from Pelee Island almost to the Ohio shore. It lies at a crestal depth of 10–12 m and separates the Sandusky basin from the central basin of Lake Erie. Its crest is broad and hummocky in the southeast, where relief increases to about 2 m, and a NW-SE valley separates the ridge into two segments. It was determined that the southeastern part of the ridge (Hartley 1960) is capped with sand and gravel deposits overlying a foundation of till deposits. Commercial quantities of sand accumulated at the southeastern extremity of the ridge (Hartley 1961). This feature is interpreted as a drowned spit, possibly sited over a low morainic ridge, which formed by southeastward longshore movement of sand from the former shoreline located southeast of Pelee Island and 11–12 m below present lake level. The Pelee-Lorain Ridge would have been an actively forming feature in early Nipissing time (4,000–5,400 ya) when lake levels were 11–12 m lower than at present, and the Sandusky basin was very shallow and separated from the central basin by the Pelee-Lorain Ridge, which formed a peninsula.

8) Lorain Bank. A submerged delta-like feature lies opposite the Pelee-Lorain Ridge at 10–12 m water depth off the mouth of the Black River. Mapping (Fuller and Foster 1998) and sediment sampling (Hartley 1961) revealed the presence of glacial till, capped by sand over part of its extent. The sand cover continues upslope to the shore and for 5 to 10 km along the shore in either direction (Pincus 1960). This feature is interpreted as a site of active formation contemporaneous with the Pelee-Lorain Ridge and the Point Pelee Fan. Its origin is uncertain; it may be an old, submerged delta of the Black River, possibly situated on a morainic foundation, or it may owe its formation in part to longshore convergence opposite the terminus of the Pelee-Lorain Ridge.

9) Pointe Aux Pins. Pointe aux Pins has the shape and morphology of a cuspate foreland. It has been interpreted as a sediment constructional feature formed in a zone of net longshore convergence (Carter *et al.* 1987, Coakley 1992). A succession of beach ridges extends southward parallel to the eastern face of this foreland, and these same beach ridges intersect with the southwest face of the feature (see Carter *et al.* 1987, p. 156). Patterns of beach ridges suggest erosion on its SW face and incremental addition of successive beach ridges to its SE face (Carter *et al.* 1987). Pointe aux Pins likely formed in post-Nipissing time, is active today, and is adjusted to present lake level. Pointe aux Pins may have initially formed opposite the moraines of Erieau Ridge, located just inland from the point, and slowly migrated northeastward along the shore.

10) Cleveland Ridge. Two approximately 4 × 8 km terraces, one at about 15 m depth and one at about 17 m depth, occur on the Cleveland Ridge about 5 km and 10 km from the south shore, respectively. The Cleveland Ridge itself has been interpreted as morainic in origin (Sly and Lewis 1972), and sand and mud deposits, together with some glacial drift, have been mapped in the area (Thomas *et al.* 1976, Fuller and Foster 1998). Location of the terraces associates them with the Cuyahoga River, and their depth clearly points to their formation in pre-Nipissing time. A deltaic origin is suggested.

11) Fairport Ridge. A small (3 × 6 km) terrace at 12–15 m depth and about 5 km from shore coincides with the Fairport Ridge, a NW-SE ridge which extends from near shore out to a depth of 18 m about 10 km from shore. Postglacial sands and muds, and glacial sediments, have been mapped in the area (Thomas *et al.* 1976, Fuller and Foster 1998). This terrace of deltaic aspect lies offshore of the Grand River, and its depth associates it with pre-Nipissing, mid-Holocene time.

12) Conneaut Bank. Conneaut Bank is a delta-shaped feature about 10 km × 15 km in size, slightly elongated in the longshore direction. Most of its terrace-top resides at a depth of 13–15 m. The feature is capped by small ridges of 1–2 m relief which are linear in approximately the E-W direction, diagonal to the shore. The top of the bank seems to follow a NW-SE broader ridge which is asymmetrical, being steeper toward the northeast. Surficial deposits of sand and/or gravel have been mapped on the top of the Bank (Hartley 1961). Conneaut Bank is interpreted as a delta which formed in middle Holocene, pre-Nipissing time. It is larger than other features of its kind along the south shore of Lake Erie. On the other hand, there are no large rivers entering the lake at this point. This leads to speculation that: 1) early in the Holocene Conneaut Creek, or another nearby stream, may have drained a larger area; 2) sediments on Conneaut Bank could have been augmented by sands transported from the Ontario shore via Clear Creek and Long Point Ridges; and 3) Conneaut Bank may have been an active delta for a longer period of time than the terraces farther west, because of greater stability of lake levels in this area in the early and middle Holocene (see lake level model and paleogeography discussed below).

13) Clear Creek Ridge. Clear Creek Ridge extends for a distance of about 50 km, from the north shore across the lake in a southeastward direction as far as the northern edge of the channel leading from the central to the eastern basin of Lake Erie. First discovered during preparation of new bathymetry (National Geophysical Data Center 1998), Clear Creek Ridge is a narrow linear ridge segmented into longitudinal hills in its northern half. Farther south the morphology of the ridge is broader, more continuous, and more complex, with numerous subsidiary ridges. The ridge has a fairly uniform crestal depth in the range of 16 m, and it lies 5–15 km west of the Long Point-Erie Ridge. Clear Creek Ridge, Long Point, and the Long Point-Erie Ridge are located near prolific sources of easily eroded sand on the Ontario shore (Barnett 1998, Rukavina and Zeman 1987). Clear Creek Ridge probably developed as a spit or bar in the early Holocene (pre-Nipissing time), when water first flooded the central basin. Large quantities of sand were swept southward, creating a peninsula, forming complex topography, and contributing to the sand deposits which occur along the southern reaches of the ridge.

14) Long Point-Erie Ridge. Long Point-Erie Ridge, a broad (14–22 km) arcuate ridge of 5–10 m overall relief, extends across Lake Erie from Long Point to the banks of the 5 km wide, 23 m deep channel connecting the eastern basin with the central basin, and it forms the boundary between the two basins. Morphology of this ridge is broader, more complex, less linear, more segmented, and shallower than the Clear Creek Ridge. The ridge is interpreted as a previously existing end moraine which developed about 13,400 ya (Barnett 1998), and which has been referred to as the Norfolk Moraine (Sly and Lewis 1972, Calkin and Feenstra 1985). The ridge formed a peninsula for a lengthy period of time in the early and middle Holocene, when lake water rose into the area, and as a peninsula, it became the site of longshore movement of sands from the Ontario shore southward across the lake floor. The ridge probably continued as an active site for longshore sand movement after sand transport ceased on the slightly deeper Clear Creek Ridge. Crestal depth of this feature is consistently about 14 m, which was probably its depth prior to being flooded by the Nipissing Rise.

15) Long Point Spit. Long Point Spit is actively building and its crestal elevation is adjusted to present lake level. This very large spit extends about 35 km east-southeastward from the Ontario shore out into the eastern Lake Erie basin. It exhibits complex depositional forms including a succession of en-echelon beach ridges diagonal to the main trend of the spit, and smaller, partially submerged small spits extending outward from the north side of the spit. Steep slopes and 55 m of relief, highest lakefloor relief in Lake Erie, separate the spit from the floor of the eastern basin of Lake Erie. It is a 5,000 ya-to-present feature (Coakley 1992) that apparently began to form after flooding associated with the Nipissing Rise inundated the Long Point–Erie Ridge. After flooding, the Long Point–Erie Ridge no longer formed a peninsula which impeded water circulation. Patterns of lake water circulation and longshore drift were apparently altered such that sands from the sand-rich Ontario shore which would have been formerly carried southward along the Clear Creek and Long Point-Erie Ridges were now employed in building and lengthening Long Point Spit.

16) Presque Isle Bank. This bank is 5 × 10 km in areal extent. It joins Presque Isle Spit and extends westward from it. Its crestal depth of 8 m is likely an index of water level at the time it was active. Atop this bank is a 1–2 m relief arc-shaped bar which resembles in size, shape, and orientation the main recurved portion of the Presque Isle Spit. Its position suggests eastward shift of the zone of longshore convergence to Presque Isle Spit, from its former position about 7–8 km to the west. Such an eastward shift in the convergence zone of net longshore drift may have been gradual, but more likely it was episodic. This feature likely mirrored the early development of Long Point Spit, only to be later abandoned when the nodal point of longshore convergence shifted eastward as Long Point Spit lengthened. This would place the period of its active development during the Nipissing Rise about 5,000 to 3,500 ya.

17) Presque Isle Spit. This feature is presently active and is adjusted to present water level. In morphology Presque Isle has the form of a recurved spit or hook, and has been described as a sand spit. It consists of an offshore bar which follows the arcuate outer shore of the spit facing the lake, and a succession of en-echelon sand ridges, which project shoreward from the main spit. From the pattern of sand ridges, it is apparent that the spit is eroding from the west and incrementally adding new sand ridges at its eastern extremity (Bolsenga and Herdendorf 1993). The spit is presently migrating eastward, apparently in response to a strong northeastward component of longshore drift (Carter *et al.* 1987).

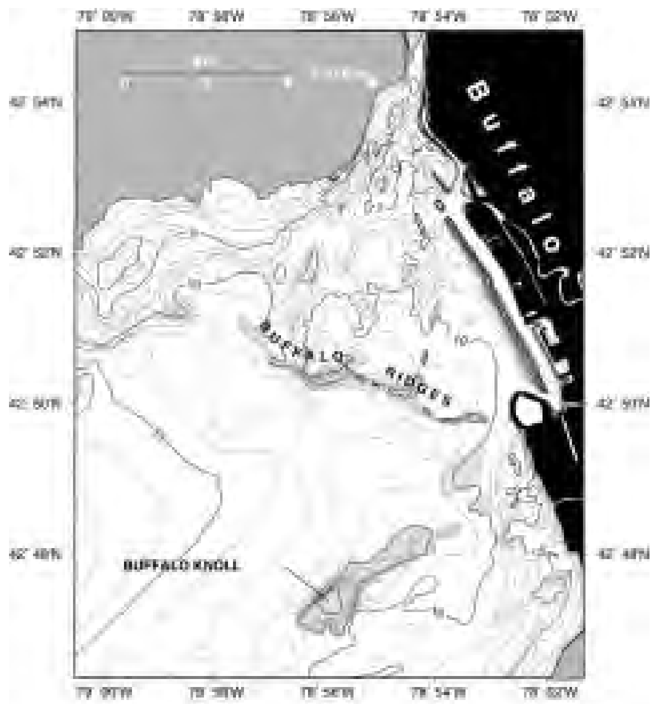


FIG. 1a. Bathymetry of the eastern end of Lake Erie showing the Buffalo Ridges and Buffalo Knoll. The two features are highlighted in gray. Contour interval 1 m.

18) Buffalo Ridges. In the northeastern corner of Lake Erie just off Buffalo, features resembling offshore bars and spits occur at depths of 10–12 m (Fig. 1a). The features were first outlined in detail in the new bathymetry (National Geophysical Data Center 1998) and given the name Buffalo Ridges. The ridges connect the offshore ends of three ridges extending southwestward from the Niagara River outlet at Fort Erie, and they form a barrier across the valleys between the headlands formed by the ridges. A similar bar or spit, also at 10–12 m depth, is on trend with the Buffalo Knoll, and extends northeastward away from the knoll. Six shallow basins behind the ridges resemble embayments. The Buffalo Ridges are interpreted as bay bars, extending away from eroded headlands and coalescing, which formed in pre-Nipissing time (5,000-9,000 ya) when water level fell below the level of the outlet sill. Two other possible interpretations are noted.

One, the feature could be an end moraine, and the small enclosed shallow basins on the shore side of the feature could be kettles. An inferred connection between the Crystal Beach Moraine in Ontario and the Alden Moraine in New York State (through the southern edge of Buffalo) extends through the area (Calkin and Feenstra 1985). Two, the feature could have formed during repeated bottom scouring and redeposition associated with heavy storms, whose effects at depth may be magnified at the eastern extremity of the lake.

LAKE AREA AND LAKE VOLUME AS A FUNCTION OF WATER LEVEL

Rate of increase in lake surface area and lake water volume as a function of rising lake level is simulated by hypsographic curves and cumulative lake water volume curves. These simulations are derived from the overall shape of the Lake Erie basin as it exists today, and they do not take into account 1) amount of infilling by sediments, and 2) amount of differential isostatic rebound, which has occurred in the Holocene. Nevertheless, the hypsometric and water volume computations are instructive in helping to explain Holocene lake level history as a function of overall shape of the lake basin, water budget, and paleogeography.

Surface area of Lake Erie was computed by cumulating the surface areas enclosed by each adjacent pair of 1 m contours projected to the zero datum, beginning with the 63–64 m interval, and ending with the 0–1 m interval. The results of this computation are shown as a cumulative hypsographic curve in Figure 2. The hypsographic curve simulates starting with an empty lake basin and filling it with 1-meter increments of water until the water level reaches the present lake level. In this manner the increase in lake surface area which would be achieved by each 1-meter increase in lake level is measured. The surface area at 0 m, 26,427,000,000 m^2, or 26,427 km^2, is the computed surface area of present Lake Erie.

From the 64 m to 24 m depth levels, surface area does not increase at a high rate. At these levels water is confined to the relatively small eastern basin. At the breakpoint observed at 24–23 m depth, surface area increases dramatically with each 1-meter rise in lake level. This accelerated increase in surface area results from the simulated flooding of the large central basin of Lake Erie. Shallower than the 20–19 m interval, the rate of increase in

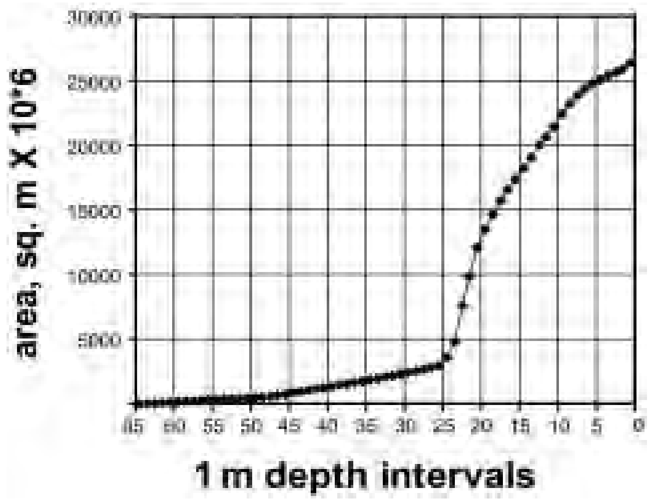


FIG. 2. Surface Area of Lake Erie as a function of lake level. Depicted are the surface areas enclosed by each successive pair of 1-m bathymetric contours, starting with 63–64 m, 62–63 m, 61–62 m, and continuing to 0–1 m depth; and cumulated from 64 m to 0 m.

lake surface area is less dramatic, but the rate of increase still exceeds that of the 64–24 m interval.

Lake volume as a function of lake level was likewise simulated by cumulating successive volumes of 1-meter layers of water stacked one upon the other, beginning with the 63–64 m layer and contin-

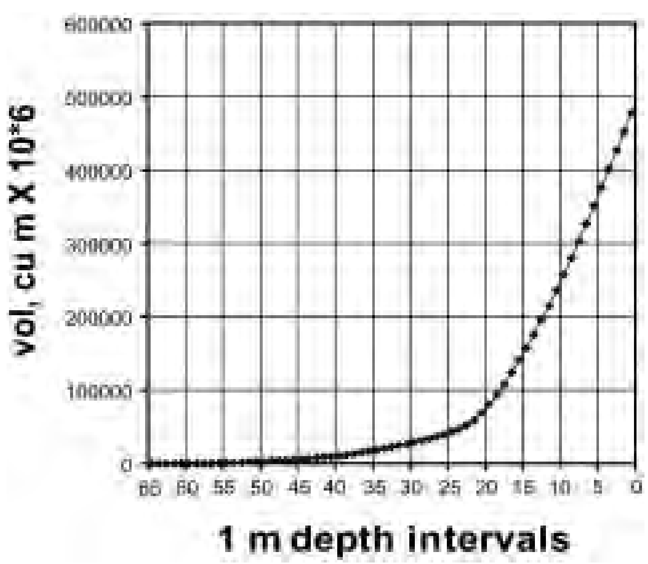


FIG. 3. Water Volume in Lake Erie as a function of lake level. Depicted are water volumes of successive stacked 1-meter-thick layers of lake water, cumulated from 63–64 m depth to 1–0 m depth.

uing upward to the present lake surface (Fig. 3). Volume of each 1-meter layer is adjusted for bottom slope along its edges. The cumulative total is the computed water volume of present Lake Erie, 478,074,000,000 m^3, or 478.074 km^3. Increase in water volume accelerates at 25–20 m as does surface area, and for the same reasons. Unlike lake surface area, the incremental increase in cumulative water volume continues to accelerate as water level rises from 15 m to 0 m.

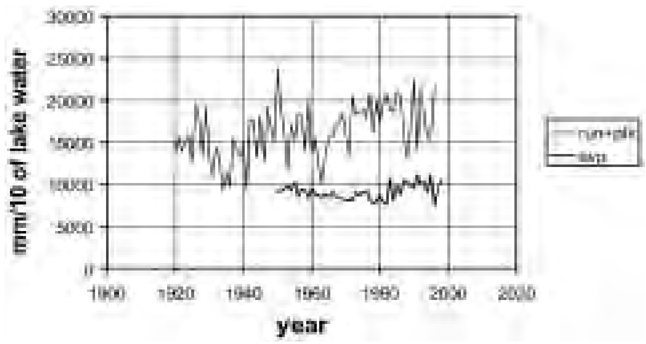


FIG. 4. Annual totals of Lake Erie drainage basin runoff + lake surface precipitation (run+plk), 1920–1996, compared with lake surface evaporation (evp), 1950–1998. Values are expressed in tenths of mm of lake water at present lake levels. Thin line is run+plk, heavy line is evp. Runoff from the Detroit River is not included in these annual totals.

LAKE WATER BUDGET

Quantitative measurements and estimates of the present water budget are of profound importance to understanding the water-level history of Lake Erie in the Holocene. Present-day water budget measurements and estimates for Lake Erie appear in spreadsheet form on the NOAA Great Lakes Environmental Research Laboratory website: http://www.glerl.noaa.gov/data/arc/hydro/mnth-hydro.html

Highlights of the present Lake Erie water budget have been presented and discussed (Quinn and Guerra 1986).

Annual totals of runoff and over-lake precipitation (run+plk) from 1920–1996 are compared with annual total of evaporation (evp) from 1950–1998, as shown in Figure 4. These annual totals are from the Lake Erie drainage basin exclusive of Detroit River inflow. Evaporation exceeded the lowest historic (run+plk) during nine of the past 48 years. If historically high evaporation should occur in the same year as historically low precipitation, the condition would exist today for Lake Erie water level to fall below the outlet sill, if Detroit River runoff were not included in the water budget. The annual total of net basin supply (nbs=run+plk-evp) from 1950–1996, has remained positive though highly variable, as shown in Figure 5. It is apparent from Figure 5 that net basin supply could go negative

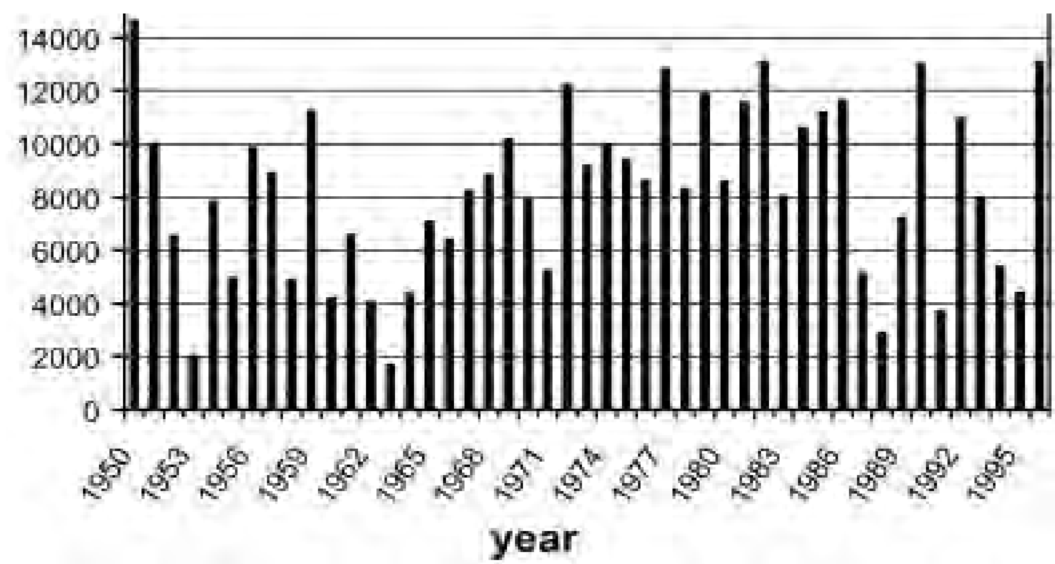


FIG. 5. Lake Erie annual total net basin supply (nbs = run+plk-evp) 1950–1996. Values are expressed in tenths of mm of lake water at present lake levels. Detroit River runoff is not included in these annual totals of nbs.

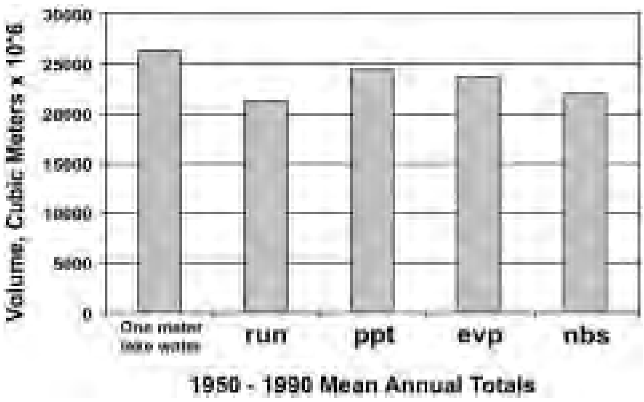


FIG. 6. Volume of 1 m of surface lake water, compared with 1950–1990 mean annual totals: runoff, over-lake precipitation, evaporation, and net basin supply. Values are expressed in millions of cubic meters. Detroit River runoff is not included in the annual totals. The volume of 1 m of lake water is computed for present lake level at the Lake Erie low water datum.

even today during an extraordinarily hot, dry, windy year, again assuming no Detroit River flow.

The 1950–1990 means of annual total water volume of runoff, over-lake precipitation, evaporation, and net basin supply, for the Lake Erie drainage basin exclusive of Detroit River inflow, shown in Figure 6, are all about the same magnitude. There is considerable variability from year to year, but even without Detroit River water being included, present net basin supply contributes enough water most years to raise water level of Lake Erie by almost 1 m in a year, enough to completely fill the lake basin in less than 20 years. Net basin supply is especially variable from year to year. Annual variations in runoff and/or precipitation and/or evaporation over the lake basin combine to have a large effect on the variability of net basin supply. Mean annual runoff carried by the Detroit River for 1940–1979 was about 170,000 × 10*6 cubic meters (Quinn and Guerra 1986), or about 8 × runoff originating from within the Lake Erie drainage basin.

Considering the main features of today's Lake Erie water budget, as outlined above and illustrated in Figures 4–6, the following observations relate to the understanding of Holocene lake levels:

1) The amount of runoff entering present Lake Erie from the upper Great Lakes is almost an order of magnitude greater than that entering Lake Erie from the local drainage basin.
2) An amount of water equal to the entire present water capacity of the lake is supplied by the Detroit River in about 2 years.
3) An amount of water equal to the entire present water capacity of the lake is supplied by the local net basin supply in about 20 years.
4) The uppermost 10 m of Lake Erie water could be supplied by the Detroit River in little over a year, or by the local drainage basin in less that 10 years.
5) Lake surface evaporation is a significantly large term which during today's driest years would be large enough to equal over-lake precipitation plus drainage basin runoff, if the water contribution from the upper Great Lakes is not included.
6) Lake surface evaporation is directly proportional to lake surface area for a given climatic condition.
7) Lake Erie evaporation may have exceeded drainage basin runoff plus over-lake precipitation for long periods of time during the middle Holocene prior to the Nipissing Rise. During the middle Holocene (about 9–6 ka), climate models and paleoclimate data (pollen, lake levels) reveal warmer temperatures and greater aridity, compared to present, in the continental interior of North America (Webb *et al.* 1993). Summer insolation in the northern hemisphere exceeded that of today by 4 to 7 % (Kutzbach and Webb 1993).
8) Lake level rises resulting from greater humidity and lower temperatures, or from significant amounts of water entering Lake Erie from the upper Great Lakes, should appear as relatively rapidly occurring events in the Holocene record.

DID LAKE ERIE BECOME A CLOSED LAKE BASIN IN EARLY AND MIDDLE HOLOCENE?

The likelihood that Lake Erie resided within a closed lake basin in the early Holocene and possibly middle Holocene was recognized by Lewis (1999), who cited deeply-buried shoreline morphology and an unconformity in the eastern basin of Lake Erie as evidence. Lewis (1999) also computed sustainable lake-areas for Lake Erie, using the Bengtsson and Malm (1997) model and estimates of early Holocene paleoclimate, and found that the computed sustainable areas were less than projected open-water lake areas for the early Holocene (9,500–7,000 ya).

Bengtsson and Malm (1997) discussed water budget within closed lakes, and the (run+plk-evp) conditions necessary for closure, and they presented equations for computing sustainable lake area in a closed lake if drainage basin area, drainage basin runoff, over-lake precipitation, and evaporation are known or estimated.

Another large lake in the continental interior of North America that probably existed within a closed basin during early and middle Holocene is Lake Winnipeg. Lewis *et al.* (2001) found evidence for closed lake basins and desiccation in seismic reflection data and cores obtained from this lake. From drainage basin areas and climate estimates they derived a sustainable lake-area model for the north and south basins of Lake Winnipeg. Their model (Lewis *et al.* 2001) projects a closed lake basin at approximately 8,300–5,300 ya.

Our exercise in reconstructing lake level history and paleogeography of Lake Erie, based on new bathymetry which reveals details of former shoreline features now submerged in Lake Erie, also leads us to seriously consider that Lake Erie was a closed lake basin in the early and middle Holocene.

For the early and middle Holocene we propose the following scenario for consideration. At first the central basin was the site of a small isolated lake separated from the lake in eastern Lake Erie by the sill in the Pennsylvania Channel (Pennsylvania Sill); later, as isostatic rebound progressed, rising water flooded the Pennsylvania Sill and formed a single lake. As the central basin was flooded, surface area of the lake increased. A very large increase in lake surface area was accomplished with only a few meters rise in lake level, as illustrated in Figure 2. Increase in the surface area of the lake proportionately increased evaporation (evp) relative to precipitation (plk) plus runoff (run). Increased evaporation as well as reduced precipitation during the 9,000–6,000 ya climate optimum (Webb *et al.* 1993) favored reduction of lake level below the level of the Lake Erie outlet sill. It seems likely that Lake Erie remained a closed lake basin for an extended period of time.

During the early and middle Holocene (10,300–5,400 ya), except for pulses of upper Great Lakes water (such as the Mattawan high-water event at about 8,500 ya), water supply for Lake Erie was limited to the Lake Erie drainage basin. Lake level was controlled by the Lyell-Johnson Sill, which was at a higher elevation than the Fort Erie Sill (Tinkler *et al.* 1994).

During the lengthy period Lake Erie is projected to be within a closed basin, water level undoubtedly fluctuated, but remained below (12–18 m?) today's level. Lake floor topographic features in the vicinity of the Long Point-Erie Ridge (Fig. 1) require water levels in this range for their formation. A number of radiocarbon dates show a long period of lake levels about 15 m lower than at present (Coakley and Lewis 1985). Control of water level by the water budget in a closed low-water basin, is compatible with a smooth decay in isostatic rebound as favored by isostatic rebound models of the Great Lakes region (Clark *et al.* 1994), and does not require a start-stop model. Throughout this period, the western basin probably remained mostly as dry land but with separate small marshes and lakes in several small basins (Coakley and Lewis 1985).

In the closed-lake scenario for Lake Erie, water level and lake area were held between upper and lower limits, in equilibrium with the water budget (run+plk-evp). Rising water level would have increased lake surface area and correspondingly increased the evaporation term, whereas runoff + over-lake precipitation would have remained constant. Water level would have been prevented from rising above the level at which the water budget (run+plk-evp) became negative. Lowering water level would have reduced lake surface area and correspondingly reduced the evaporation term, preventing water level from falling below the level at which the water budget becomes positive.

If the Buffalo Ridges (Fig. 1a) are in fact drowned shoreline features, this would constitute evidence that lake level did indeed fall below the level of the outlet sill. Morphologically the features resemble bay bars enclosing valleys between the NE-SW ridges which extend away from the area of the outlet sill. The bar complex is continuous across the ridges, which appear to be truncated. This suggests that the lake shore remained at about this level for a sufficient length of time to erode and straighten the headlands, but as recounted earlier, other interpretations are possible. The geomorphology of the Buffalo Ridges is one of the intriguing questions of Lake Erie geology remaining to be answered.

Twelve of the eighteen depositional features described in this paper are interpreted as having formed in the shore zone at lower lake levels. Once their period of active formation ceased, it seems unlikely that these features could long survive in a

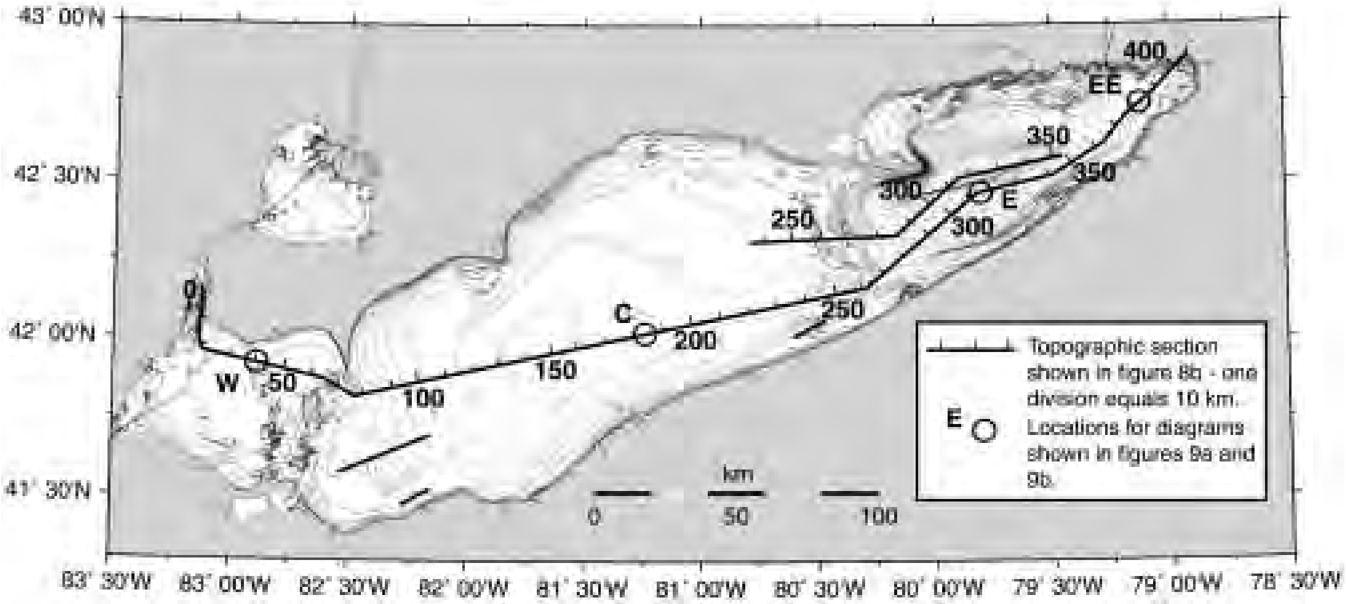


FIG. 7. Longitudinal Lake Erie section lines shown in Figures 8a and 8b, and sites of postulated lake level histories illustrated in Figure 9. Distance along section lines is shown in kilometers. Sites of postulated lake level histories are labeled W, C, E, and EE. Geographic locations are W (41° 54′N, 82° 45′W), C (42° 02′N, 81° 06′W), E (42° 33′N, 79° 43′W, and EE (42° 48′N, 79° 00′W).

high-energy shore zone characterized by slowly rising water levels under the control of differential isostatic rebound. Rapid rise in lake level was needed to rapidly inundate these features before they could be reduced by shore zone erosion. This strengthens the argument in favor of lake levels about 10 m below the level of the outlet sill for considerable periods of time, followed by rapid rises in lake level in response to alterations in the water budget which favor such rise.

It has been commonly assumed that the Nipissing Rise in Lake Erie is associated with diversion of upper Great Lakes outflow from the North Bay and Chicago outlets to the Port Huron outlet. However, if water level fell below the level of the outlet sill in early Holocene time, a climate-driven return of water level to the level of the outlet sill could have been a significant rising water event. On the other hand, if water level remained at the level of the outlet sill through the mid-Holocene, then the introduction of upper Great Lakes water would be expected to raise water level by only an estimated 3 to 3.5 m (estimate based on today's elevation difference between the Lake Erie low water datum and the sill level at Fort Erie).

Figure 7 is an index map which shows the location of sections shown in Figures 8a and 8b, and sites of lake level history simulations shown in Figure 9. Figure 8a illustrates inferred synchronous lake levels for three conditions which may have existed before and during the Nipissing Rise: 1) water budget negative, lake level 10 m below the level of the (Lyell-Johnson) outlet sill; 2) water budget weakly positive, no upper Great Lakes water, lake level adjusted to the level of the outlet sill; and 3) water budget strongly positive, includes upper Great Lakes water, lake level adjusted to the level of the outlet sill. In Figure 8a, the solid green line is a longitudinal end-to-end topographic profile of Lake Erie along the section line shown in Figure 7. Profiles of nearby topographic features are projected onto the plane of this section, including the Buffalo Ridges, eastern Lake Erie basin, Long Point-Erie Ridge, Clear Creek Ridge, Conneaut Bank, Pelee-Lorain Ridge, and Lorain Bank. The solid orange line approximates the buried interface between postglacial Holocene deposits above and glacial/ glacio-lacustrine deposits below, as previously mapped by Lewis (Lewis *et al.* 1966, Sly and Lewis 1972). Amount of sedimentation vs. time in the Holocene was interpolated linearly between the buried interface, interpreted to represent lake level at the beginning of Holocene, and the present lake floor. The interpolated sedimentation vs. time was used 1) to

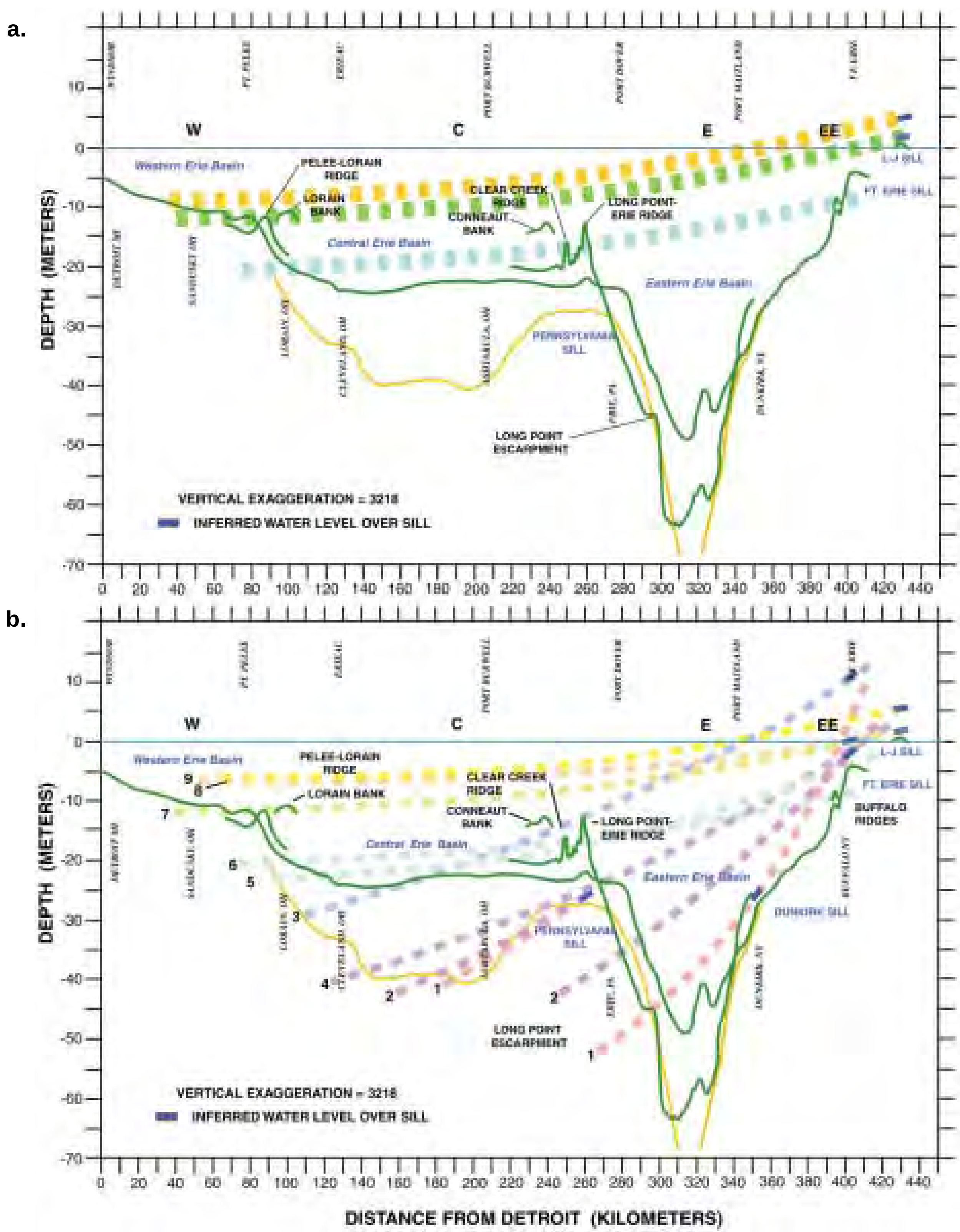


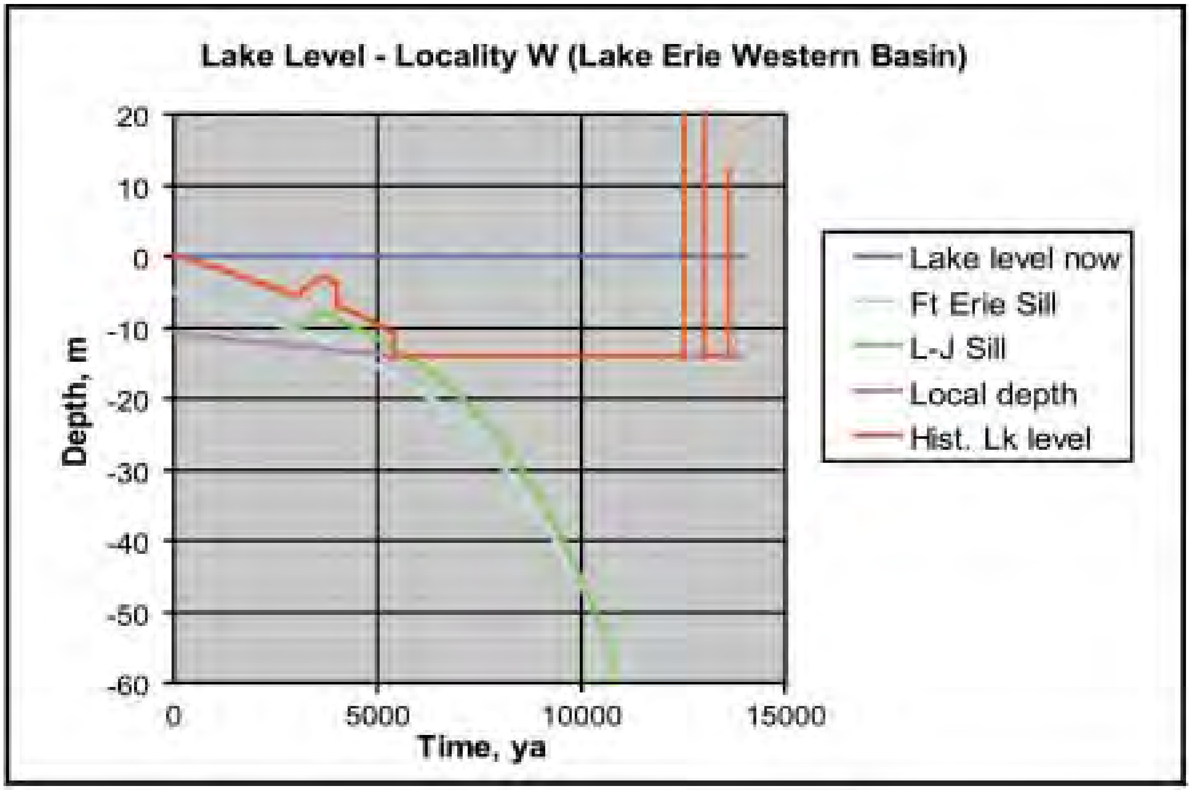
FIG. 8. a. Water planes for three possible lake levels in mid-Holocene time. Refer to text for additional explanation. b. Schematic model of lake levels at successive times during the Holocene. Refer to text for additional explanation.

estimate local depth as a function of time in Figure 9, and 2) to refine estimates of paleogeography shown in Figure 10. The three dashed lines show inferred water planes simulating three different lake levels at about Nipissing time. The blue dashed line shows inferred water plane for a pre-Nipissing lake, which has fallen 10 m below the level of the outlet sill. The green dashed line shows inferred water plane adjusted to the level of the outlet sill, but with the water supply limited to that coming from the Lake Erie drainage basin. The orange dashed line shows inferred water plane adjusted to the level of the outlet sill, but with the water supply including water from the upper Great Lakes.

In Figure 8b, the postulated former water levels (labeled 1 through 9) are sketched on a longitudinal topographic section through the deep axis of Lake Erie, to approximately agree with topographic features which are thought to be a record of the previous levels of shorelines. Positioning of each level also takes into account isostatic rebound models (Clark *et al.* 1994); the present position of the Lake Whittlesey and other shorelines as shown in Coakley and Lewis (1985); and lake level histories outlined in Barnett (1985), Calkin and Feenstra (1985), Pengelly *et al.* (1997), and Lewis (1969). Lake level 1, two lakes, one in the central Lake Erie basin and one in the eastern Lake Erie basin, were controlled by the Pennsylvania Sill and the Dunkirk Sill respectively. They are assumed to have about the same west-to-east elevation gain as the Lake Whittlesey shoreline, but are at a low elevation associating them with the Ypsilanti lowstand (Kunkle 1963), and are therefore assigned a time of 13,400 ya. Lake level 2, two lakes, are adjusted to the Pennsylvania Sill and the Fort Erie Sill, and are assigned an age of around 12,000 ya. Lake level 3, now a single lake, coincides with the highstand of Main Lake Algonquin and is assigned an age of 10,400 ya. Lake level 4 coincides with the beginning of the pre-Nipissing lowstand, before water occupied a significant area of the central Lake Erie basin, and before the shift to the Lyell-Johnson Sill; it is assigned an age of 10,300 ya. Lake level 5 illustrates a time about midway into the pre-Nipissing lowstand, after the central basin was flooded, and after lake level had fallen below the sill and consequently was controlled by the water budget; it is assigned an age of 7,500 ya. Lake level 6, adjusted to lakefloor features which occur all around the lake, and influenced by the Clark *et al.* (1994) and Larsen (1985) pre-Nipissing isostatic rebound models for Lake Michigan, is believed to represent the period before the Nipissing Rise, and is assigned an age of 5,400 ya. Lake level 7 represents an early Nipissing Rise event prior to the main introduction of upper Great Lakes water, but after water level has risen to the level of the Lyell-Johnson outlet sill. It is assigned an age of 5,300 ya. Lake level 8 represents the Nipissing Rise, after introduction of upper Great Lakes water but before significant erosion of the Lyell-Johnson Sill. It is assigned an age of 3,600 ya. Lake level 9 represents a lower lake level resulting from erosion of the Lyell-Johnson Sill, and return of control of lake level to the Fort Erie Sill. It is assigned an age of 3,000 ya.

The locations for the sites depicted in Figure 9 are shown in Figure 7, and are labeled W (western Lake Erie) and C (central Lake Erie) (Fig. 9a), as well as E (eastern Lake Erie), and EE (near the outlet sill) (Fig. 9b). For Figure 9, low lake levels are inferred from location and depth of diagnostic bathymetric features, from the Clarke *et al.* (1994) isostatic rebound model, and from previous reconstructions of lake level history (Lewis 1969, Sly and Lewis 1972, Barnett 1985, Coakley and Lewis 1985). High lake levels and sill evolution are adapted from Pengelly *et al.* (1997) and Barnett (1998). Highstands in chronological order are 1) the series of short-lived glacial Lakes Maumee and Arkona (levels for each lake not shown), pre-13,600 ya; 2) the series of short-lived glacial Lakes Whittlesey, Warren, Wayne, Grassmere, and Lundy (levels for each lake not shown), 13,000–12,500 ya; 3) early Lake Algonquin, 12,000–11,600 ya; main Lake Algonquin, 11,000–10,400 ya; 4) main Mattawan, 8,500 ya; and 5) Nipissing, 5,400–3,600 ya. Lowstands in chronological order are 1) Ypsilanti low, 13,600–13,000 ya; 2) unnamed low, 12,500–12,000 ya; 3) Kirkfield low, 11,600–11,000 ya; and 4) pre-Nipissing Lake Erie 10,400–5,400 ya. Beginning with the Ypsilanti lowstand, lake level was controlled by the Dunkirk Sill, but by the time of the Kirkfield lowstand, lake level was controlled by the Fort Erie Sill. At about the time lake level fell following the main Algonquin highstand, the rising Lyell-Johnson Sill began to control lake level. With further isostatic rebound and lake level rise during 10,400–5,400 ya, lake water occupied a large area of the central basin, and lake level fell below the sill, controlled instead by the water budget during this period, when climate was warmer and drier than at present (Webb *et al.* 1993). This regime continued, with one or two pulses of high

a.



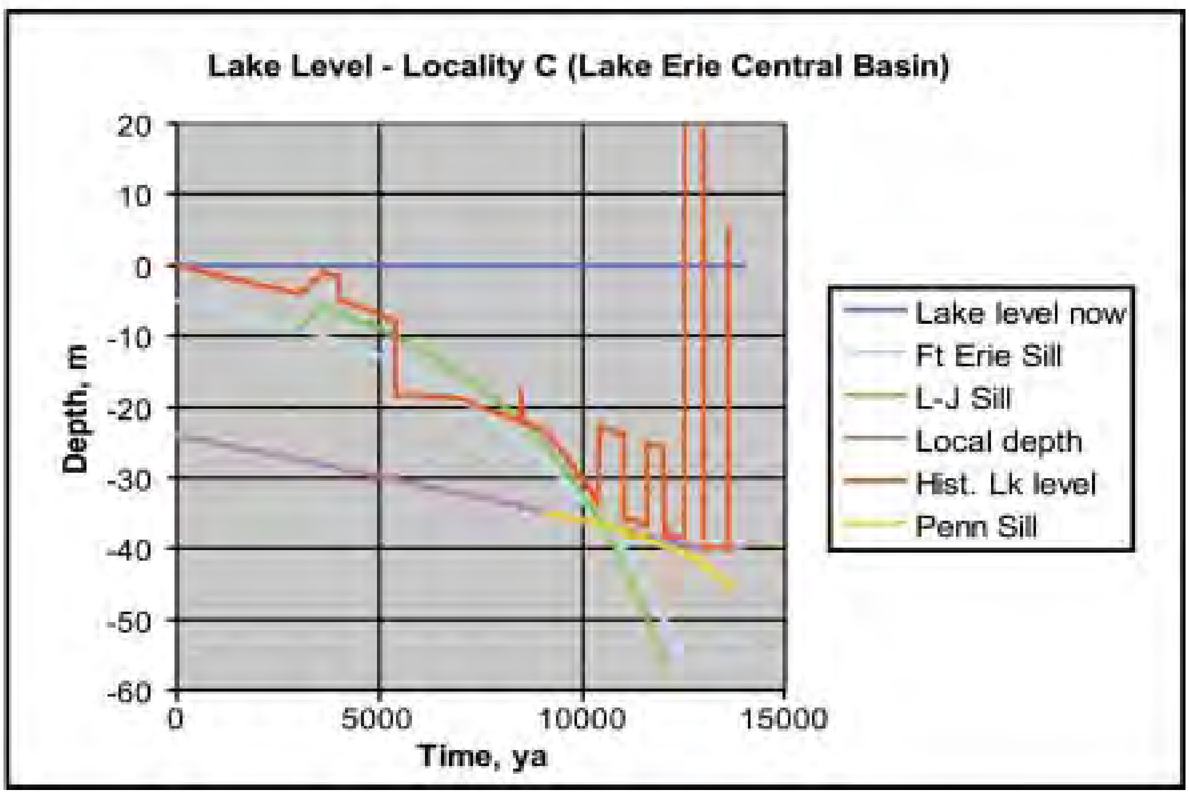
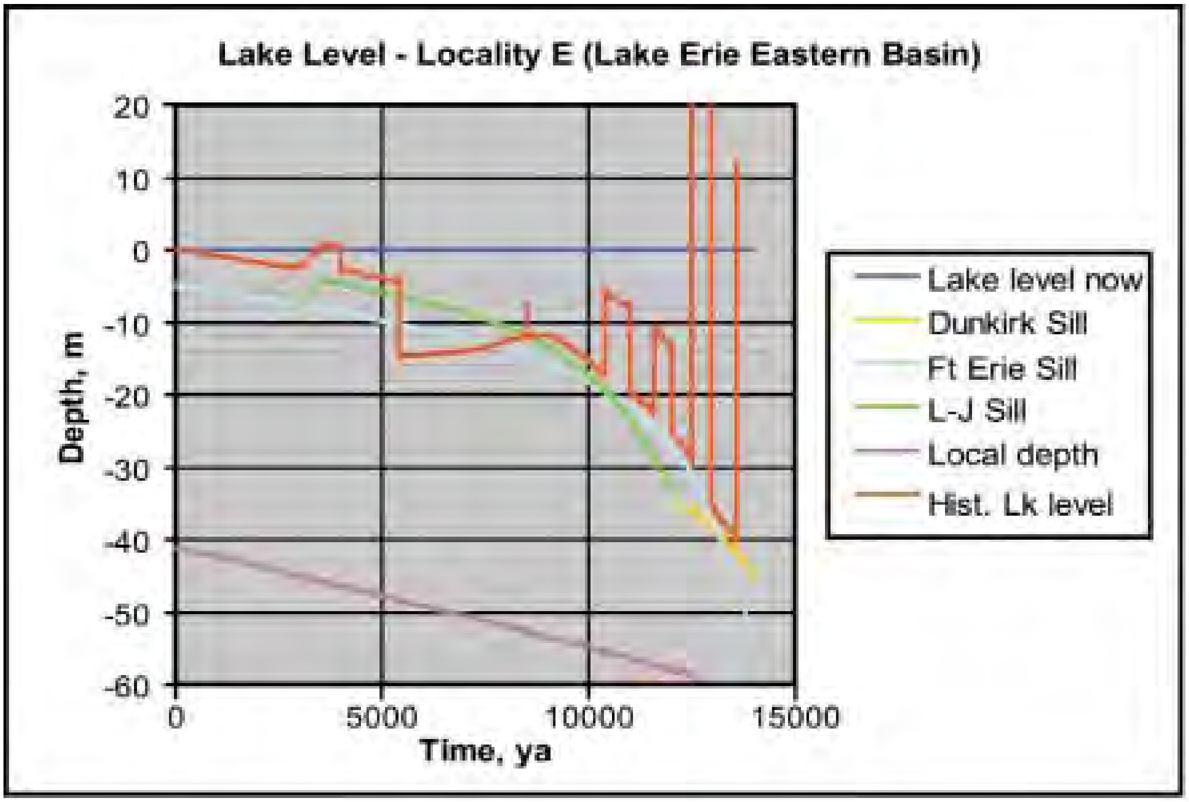
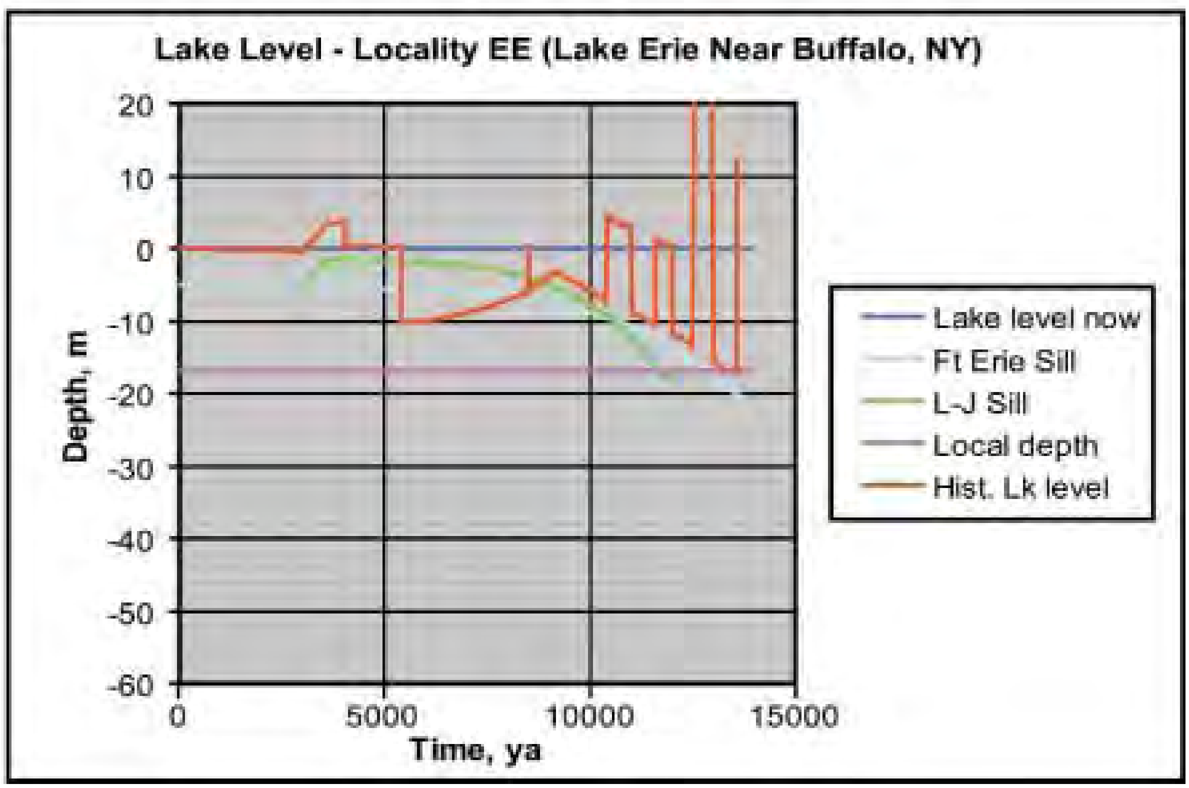


FIG. 9. Schematic model of Holocene lake level history at four locations: in the western, central, and eastern Lake Erie basins; and near the outlet sill in the eastern end of the lake. The locations for the sites depicted in Figure 9 are shown in Figure 7, and are labeled W (western Lake Erie) and C (central Lake Erie) (Fig. 9a), as well as E (eastern Lake Erie), and EE (near the outlet sill) (Fig. 9b). Refer to text for additional explanation.

b.


Lake Level - Locality E (Lake Erie Eastern Basin)
20
10
0
-10
-20
-30
-40
-50
-60
Depth, m
0
5000
10000
15000
Time, ya
Lake level now
Dunkirk Sill
Ft Erie Sill
L-J Sill
Local depth
Hist. Lk level


Lake Level - Locality EE (Lake Erie Near Buffalo, NY)
20
10
0
-10
-20
-30
-40
-50
-60
Depth, m
0
5000
10000
15000
Time, ya
Lake level now
Ft Erie Sill
L-J Sill
Local depth
Hist. Lk level

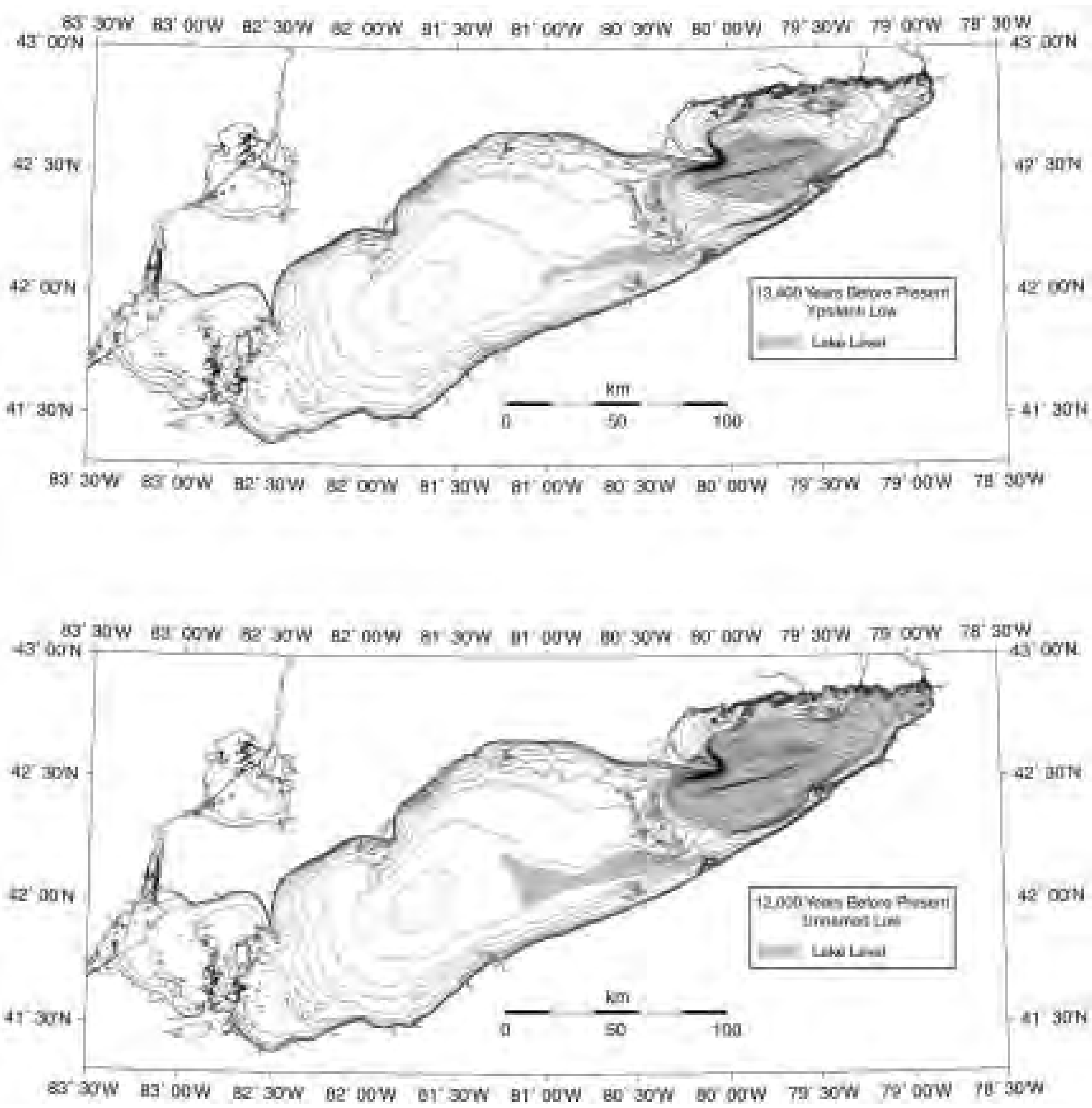


FIG. 10. Postulated paleogeography of Lake Erie at nine selected times which are representative of the entire postglacial period from 13,600 ya to present. Refer to text for additional explanation. Figure 10 continued on next four pages.

water over the Port Huron Sill (Mattawa highstands), until changing climate and/or introduction of upper Great Lakes water initiated the Nipissing Rise, which returned water level to the level of the Lyell-Johnson Sill. Water level was elevated again by return of large-scale upper Great Lakes drainage about 4,000 ya. Water level then began to fall, as the Lyell-Johnson Sill was eroded, lowering water level onto the Fort Erie Sill, which controlled lake level from then until present. Lake-level changes are represented as rapidly occurring events.

RECONSTRUCTION OF HOLOCENE LAKE LEVEL HISTORY

Locations and depths of bathymetric features in Lake Erie which are indicators of present and historic lake levels (Fig. 1), together with relevant area (Fig. 2), volume (Fig. 3), and water budget data (Figs. 4–6), further constrain the postglacial lake

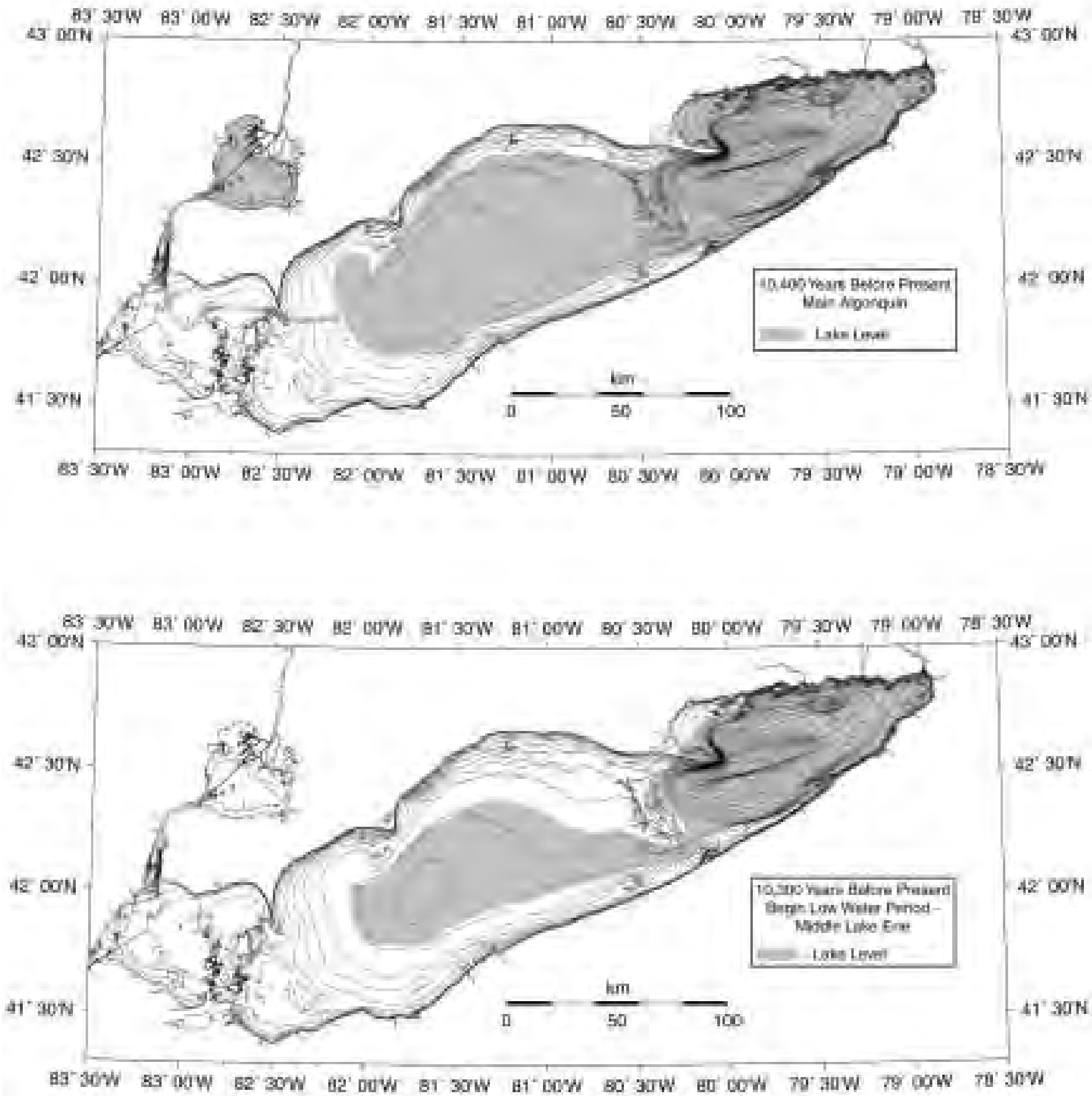


FIG. 10. Continued.

level history of Lake Erie. These additional data make reconstruction of Holocene lake levels a useful exercise, when previous reconstructions and existing lake level information are also taken into consideration. Toward this end we have developed a schematic model. This model, though elaborate, leaves questions unanswered, contains elements of speculation, and is undoubtedly simplified in comparison to the actual unfolding of events. Even so, we believe that it is a reasonable model which contributes to further understanding of Lake Erie's Holocene geology and lake level history.

The model incorporates the following features:

1. During earliest postglacial time (13,600–12,000 ya), lake level oscillated between very low levels, when the Niagara outlet was open but isostatically depressed, and very high levels, when the Niagara outlet was blocked by ice and outflow was toward the west.
2. Lake level rose and fell rapidly in response to blocking, unblocking, or erosion of the outlet sill; whether upper Great Lakes water passed through Lake Erie or bypassed it; and changes in the water budget. Large water budget terms, compared to the volume of 1 m of surface lake water (Fig. 6), support this statement.

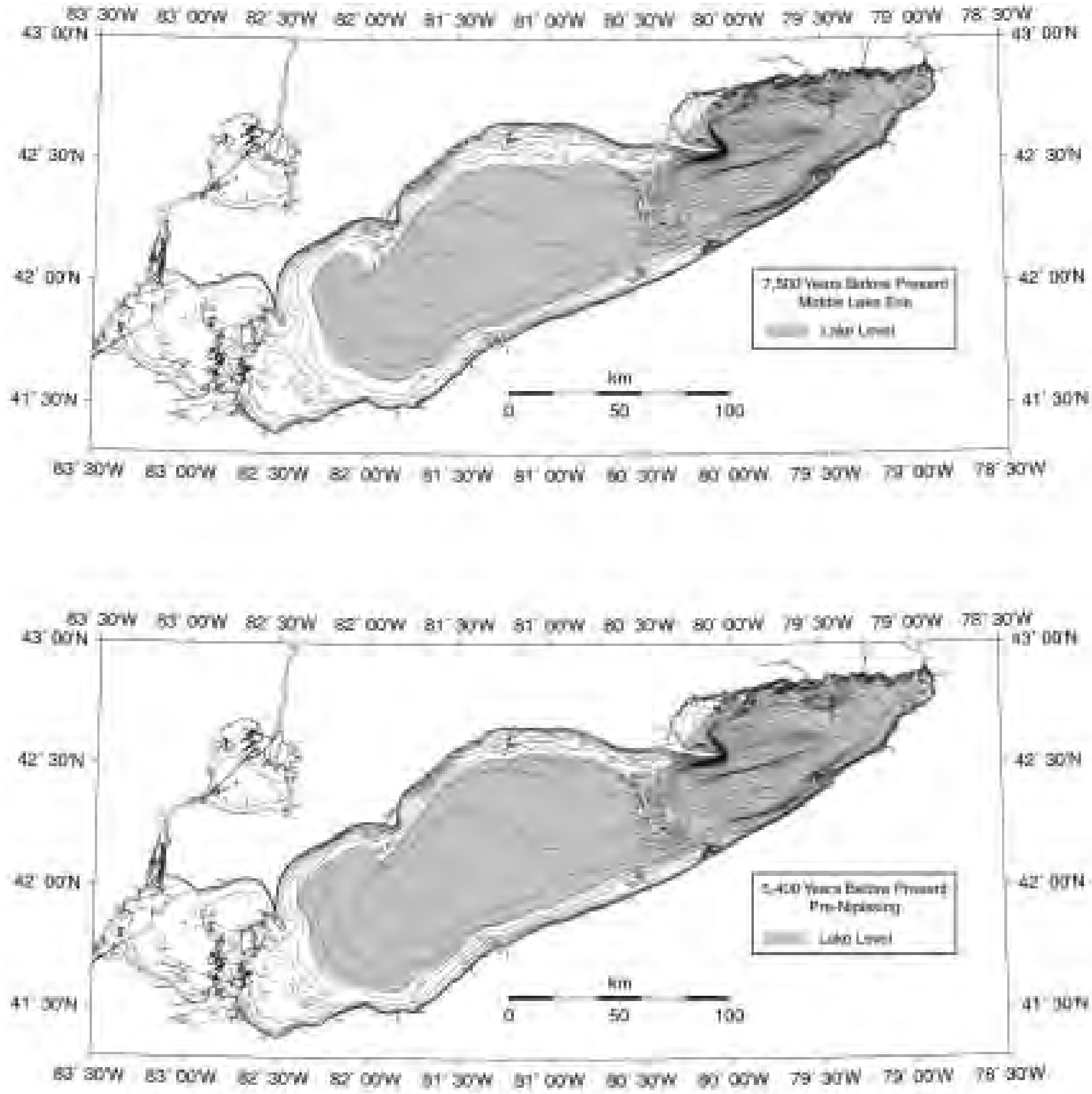


FIG. 10. Continued.

3. Lake level fell below the level of the outlet sill during substantial intervals of the first half of Holocene time, 10,300–5,400 ya, most likely during the period 9,000–6,000 ya, when climate was warmer and drier than at present.
4. Also contributing to a low water level during about 9,000–6,000 ya was flooding of the central basin of Lake Erie as isostatic rebound continued, greatly increasing lake surface area (Fig. 2) and concomitantly increasing lake surface evaporation (the negative term in the water budget).
5. The early Nipissing Rise (Nipissing I), about 5,400 ya (?), resulted when lake level, which formerly had fallen below the level of the outlet sill, was restored to the level of the Lyell-Johnson outlet sill, as climate became less arid and the water budget became positive. Small and variable influx of upper Great Lakes water may have entered the Lake Erie basin at times over the next thousand years. Low-water depositional features previously formed in Lake Erie (Bay bars off Buffalo, Clear Creek Ridge, Long Point–Erie Ridge, Conneaut Bank, Fairport Ridge, Cleveland Ridge) were flooded at this time. Water circulation patterns changed, and Long Point Spit began to form.
6. Later, about 4,000 ya (?), an additional rise in

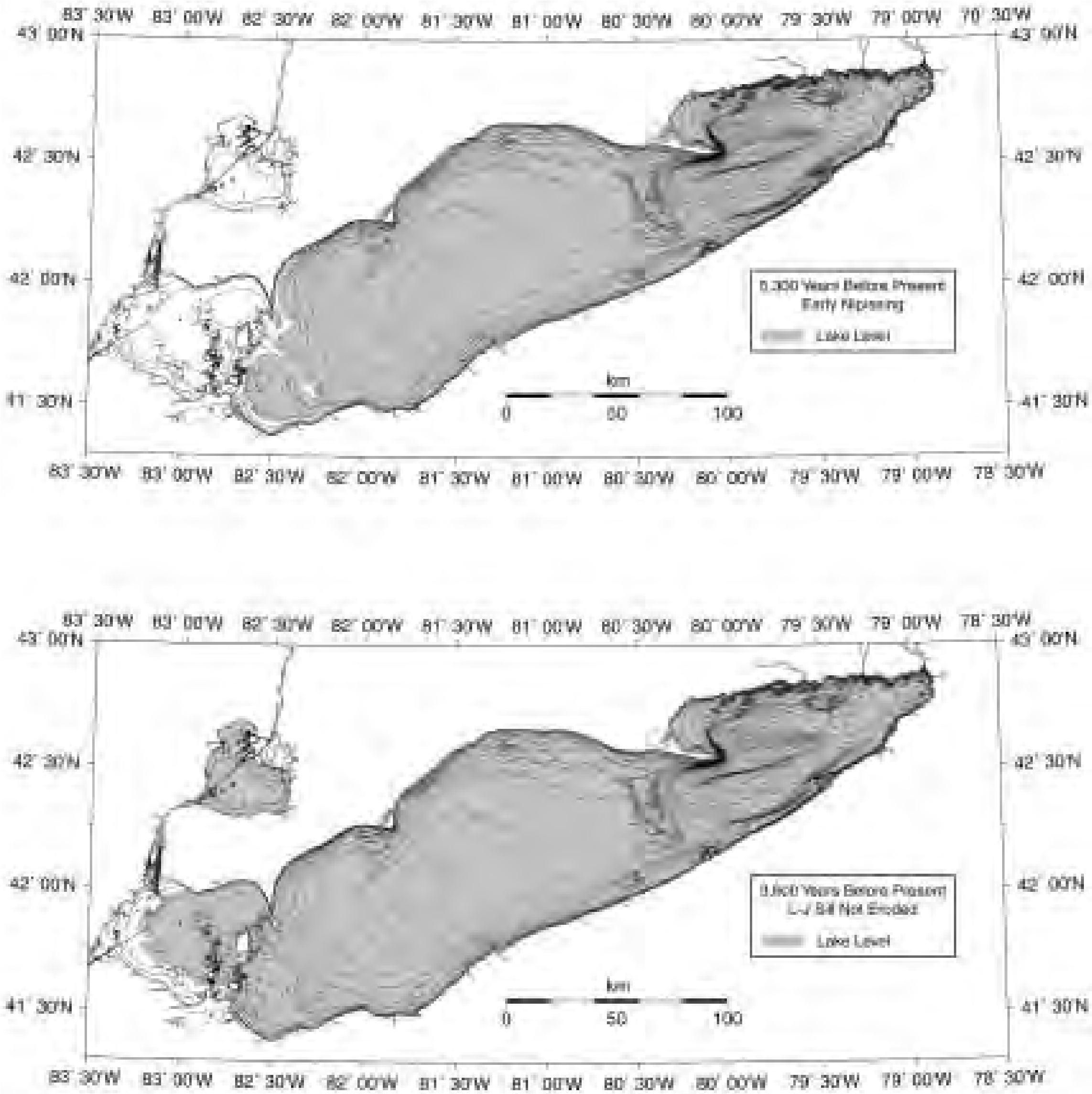


FIG. 10. Continued.

lake level (Nipissing II) resulted from the final introduction of large volumes of upper Great Lakes water into Lake Erie. At this time the western basin of Lake Erie was flooded. Also flooded were the younger and shallower depositional features which had formed in central and western Lake Erie (Pelee-Lorain Ridge, Point Pelee Fan, Lorain Bank, Presque Isle Bank). The youngest, and presently active, depositional features of Lake Erie began to form at this time (Point Pelee, Pointe aux Pins, St. Clair Delta, Maumee Spit).

7. Water levels in eastern Lake Erie were slightly higher than at present. Lake level was lowered several meters during the interval 3,600–3,000 ya by headward erosion of the Niagara gorge. Erosion of the Lyell-Johnson outlet sill returned Lake Erie to the level of the Niagara River at Fort Erie. In the western basin, water level did not fall significantly because lower water level at the sill was partially compensated by differential isostatic rebound in the west.

This Holocene lake level model for Lake Erie is illustrated in Figure 8b, which like Figure 8a is also a topographic section through the deep axis of Lake Erie as shown in Figure 7. In Figure 8b, the inferred

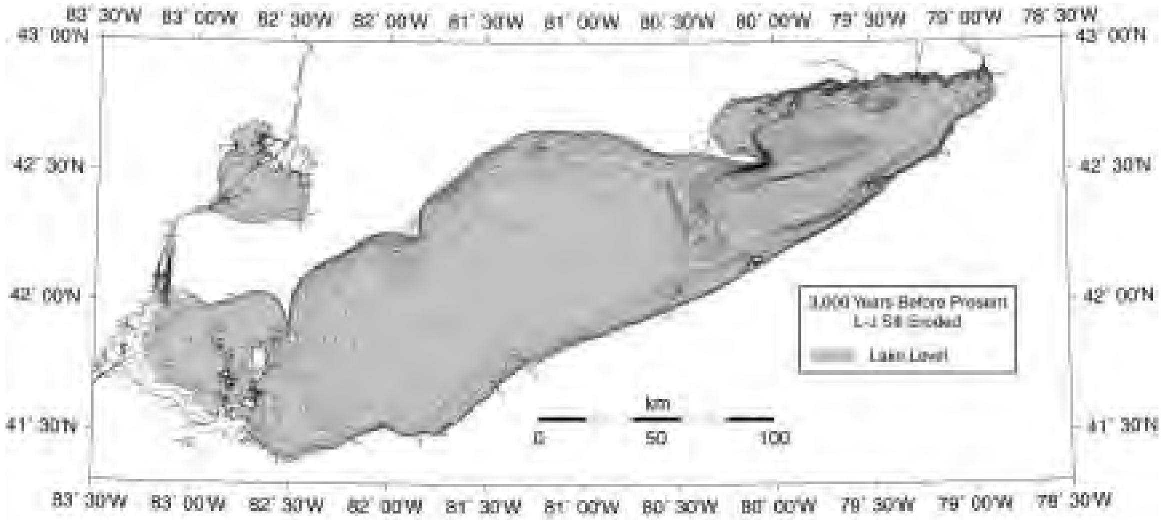


FIG. 10. Continued.

present position of water planes relative to the topography are sketched in for nine representative time intervals of the Holocene. Figure 9 shows schematic diagrams of inferred lake level history at four locations shown in Figure 7. These schematic diagrams illustrate rather large differences in lake level history between each of the four locations.

Differential isostatic rebound (Clark *et al.* 1994, thin-ice model) between each Figure 9 locality and the outlet sill varies in proportion to distance from the sill. According to the Clark model, the rate and amount of differential isostatic rebound decrease non-linearly with increasing distance from the sill. Applying the Clark model, the outlet sill rose less than 10 m in the last 10,000 years relative to the location near the sill (diagram EE). During the same period the outlet sill rose 10–20 m relative to the eastern basin locality (diagram E), 20–30 m relative to the central basin locality (diagram C), and 40 to 50 m relative to the western basin locality (diagram W).

According to the model the outlet sill moved eastward twice, first from the Dunkirk Sill to the Fort Erie Sill, and later from the Fort Erie Sill to the Lyell-Johnson Sill, as isostatic rebound elevated downstream sills at a higher rate than those upstream. Still later the Lyell-Johnson Sill was eroded (Pengelly *et al.* 1997) and the Sill shifted back upstream to the Fort Erie Sill. During the earliest low-water period (Ypsilanti Low), when the Dunkirk Sill was presumed to be active, location EE, near the Fort Erie Sill, was probably dry land. According to the model, location E, in the eastern basin, was water-covered during the entirety of the Holocene. Location C, in the central basin, was the site of a small lake impounded by the Pennsylvania Sill until the central basin was flooded about 10,500 ya. Location W, in the western basin, remained the site of dry land or small marsh/ lake basins until it was finally flooded about 5,400 ya.

According to the model, during the highest level reached by the Nipissing Rise about 3,600 ya, lake level was 4 m above present lake level at locality EE; 1 m above present lake level at locality E; 1 m below present lake level at locality C; and 3 m below present lake level at locality W.

The model does not predict shorter term fluctuations in lake level, such as annual and decadal changes which have occurred in response to the changing water budget. Such changes have undoubtedly occurred throughout the Holocene and continue at present. Shorter term variations in lake level would have been greater during times when lake level dropped below the level of the outlet sill. However large variations, when lake level is lower than the outlet sill, may be the exception rather than the norm, since most of the bathymetric features which record low lake levels are indicative of a specific depth range. Variability in lake level may have been larger during the early Nipissing Rise, during a time of delicate balance between the water budget and the introduction of water from the upper Great Lakes.

LAKE ERIE PALEOGEOGRAPHY

Paleogeographies of the nine episodes of Lake Erie lake level history depicted in the model (Fig. 10) are shown as nine separate overlays to the bathymetry. Details of Holocene shoreline configurations, to the extent that the model is accurate, are revealed, thanks to the detailed bathymetric mapping. Figure 10 shows reconstruction for times that correlate with inferred water levels 1 through 9 shown in Figure 8b; these times illustrate the postulated Lake basin(s) and outlet sill(s) during: 1) the Ypsilanti Low (13,400 ya); 2) the unnamed low episode (12,000 ya); 3) main Lake Algonquin (10,400 ya); 4) at the beginning of the pre-Nipissing lowstand (10,300 ya); 5) about the midpoint of pre-Nipissing Lake Erie time (7,500 ya); 6) just prior to the early Nipissing Rise (Nipissing I) (5,400 ya); 7) just after the early Nipissing Rise (5,300 ya); 8) following introduction of a significant volume of upper Great Lakes water via the Detroit River (3,600 ya); and 9) following erosion of the Lyell-Johnson Sill (3,000 ya).

During the first postglacial lowstand (13,500–13,000 ya) two small lakes existed, one behind the Pennsylvania Sill, the other behind the Dunkirk Sill. During the second postglacial lowstand (12,500–12,000 ya), the two small lakes persisted, one still behind the Pennsylvania Sill, the other now impounded behind the Fort Erie Sill. The lakes were larger during the second lowstand. During the main Algonquin Highstand (11,000–10,300 ya), when upper Great Lakes water was coming through Lake Erie, the central basin sill and the Long Point-Erie Ridge were inundated, and a large lake formed in the central basin. After initiation of the lengthy early/ middle Lake Erie lowstand (10,300–5,400 ya), when upper Great Lakes water again bypassed Lake Erie, the central basin sill was barely flooded, and a smaller lake in the central basin was very shallow.

Snapshots of the paleogeography at 7,500 and 5,400 ya show a large shallow lake now largely filling the central basin, but with the Clear Creek Ridge and Long Point-Erie Ridge exposed as peninsulas. The lake in the eastern basin is almost as large as it is at present, and the Long Point Spit had not yet begun to form. Following the early event (s) of the Nipissing Rise (5,300 ya), configuration of the lake in the eastern basin was approximately the same as it is at present but at a slightly higher level near the Lyell-Johnson Sill, and beyond in the Niagara River. The Clear Creek Ridge and Long Point-Erie Ridge were flooded, the central basin was about as extensive as it is at present, and water had not yet entered the western basin. At this time the Pelee-Lorain Ridge became a spit, and the Point Pelee Fan began to form. At the height of the Nipissing Rise in the time interval 4,000–3,600 ya, the Detroit River had received the full and sustained influx of upper Great Lakes water, and erosion of the Lyell-Johnson Sill had begun. The eastern end of the lake was 4–5 m above present level, and the eastern and central basins were about the same extent as at present. The western basin was flooded though water level was several meters lower than at present. Following erosion of the Lyell-Johnson Sill (3,000 ya), lake configuration was only slightly less extensive than at present. The largest difference is that the western basin was less extensive at its extremely shallow western end.

CONCLUSIONS

Eighteen Lake Erie and Lake St. Clair lakefloor bathymetric features are interpreted as indicators of present and former lake levels (Fig. 1). Of these six are presently active features adjusted to present lake level, and twelve were apparently formed at lower lake levels and later abandoned. These features include six deltas/fans, eight spits, one cuspate foreland, two bay bars, and one morainic ridge.

Around the eastern extremity of the lakes, the Buffalo Ridges occur at a depth of 10–12 m (Fig. 1a). These are provisionally interpreted as shoreline features consisting of a line of several spits and bay bars. This line is continuous across several eroded headlands truncating NE-SW ridges. Other interpretations for these features include end moraine, or formation at present depth during strong storms, effects of which are magnified in the eastern extremity of the lake. If the first interpretation is correct, the Buffalo Ridges document the existence of lake levels which had fallen below the level of the outlet sill in middle Holocene time.

Lake surface area and lake water volume as a function of lake level in the present Lake Erie, have been computed using present bathymetry, and are illustrated graphically (Figs. 2 and 3). Notwithstanding constraints imposed by not considering differential isostatic rebound and sediment infilling of lake basins, two episodes of lake level history are apparent. At lower water levels Lake Erie was small and largely confined to the eastern lake basin. Lake area and water volume did not increase at a high rate with respect to increasing lake level. At higher water levels, lake area and water volume increased at a very high rate with respect to increasing lake level, as the much larger central basin of Lake Erie was flooded.

Annual totals of the water budget terms evaporation (evp), runoff (run), and over-lake precipitation (plk) were compared with the amount of water necessary to raise present Lake Erie lake level by 1 m, based on measurements and estimates from 1950–1990 (Figs. 4 and 5). Detroit River throughput is ignored. The three water budget terms (evp, run, and plk), the net basin supply (nbs), and the volume of water contained in 1 m of lake water, all lie within the narrow range of 20–25 billion cubic meters (Fig. 6).

Features of the present water budget are instructive regarding the water budget of early Holocene Lake Erie. Two conclusions: 1) without Detroit River input, Lake Erie's present water budget could go negative during extremely dry and windy years; and 2) changes in lake levels in response to changes in water budget would occur very rapidly. At present, the entire volume of Lake Erie is supplied by the Detroit River in about 3 years, or would be supplied by the local drainage basin in about 20 years.

In the scenario where lake level falls below the level of the outlet sill, lake level ceases to be controlled by sill depth and passes primarily to the climate-driven water budget. Large increases in lake surface area as a function of lake level would have the effect of increasing evaporation, the negative term in the water budget, without significantly affecting the positive terms, over-lake precipitation plus runoff. Increased evaporation concomitant with rising water levels would tend to prevent lake level from rising beyond a certain point. Conversely, falling lake level would decrease the evaporation term, which would prevent lake level from falling beyond a certain point at which the water budget becomes positive. An equilibrium lake level would be maintained as governed by the water budget. Seasonal, annual, and longer-term changes in the water budget would raise or lower lake level, introducing a "noise level" which is not easily quantified.

During the middle Holocene (9–6 ka) climatic optimum, warmer temperatures and greater aridity characterized the climate of the region. Considering the present water budget, Lake Erie may have fallen below sill depth for a long period of time between 10 and 5 ka. Following the climatic optimum, lower temperatures and higher rainfall may have returned Lake Erie to the level of the outlet sill, possibly before introduction of significant amounts of upper Great Lakes water. An early episode of the Nipissing Rise could have been the result.

A provisional model of lake level history is proposed, notwithstanding all its uncertainties (Figs. 7 and 8). Taken into consideration are: 1) lake floor features, described in detail by recent bathymetry, which are indicators of present and past lake levels, 2) differential isostatic rebound, (3) water budget data, (4) past climate information, and (5) previous investigators' interpretations of lake level history derived from radiocarbon dates on materials indicative of lake level. Also considered are geological evidences of higher and lower lake levels from the surrounding land areas, including downcutting of stream valleys below present lake level, and beach ridges above present lake level. The model is illustrated by water-plane sketches (Fig. 8b) which depict nine episodes of lake level history, all plotted against a longitudinal section of lake floor topography.

In an isostatically rebounding lake, lake level history varies with location, especially as a function of distance from the outlet sill. A new feature of the model are lake level histories which are location-specific, as illustrated by schematic diagrams (Fig. 9) of lake level history for four locations: near the outlet sill; and in the eastern, central, and western basins. Another new feature of the model is the proposed extended period of low lake level in the middle Holocene, when lake level was controlled not by the level of the outlet sill, but by climate and the water budget. Still another new feature of the model are water-level rises associated with the Nipissing Rise which occur rapidly (1–20 yrs) and may amount to 10 m or more. What the model cannot show are short-term, cyclic variations in lake level brought about by seasonal/ annual/ decadal variations in climate and the water budget.

Based on the model of lake level history, Lake Erie paleogeography varied spectacularly during the Holocene, as illustrated (Fig. 10) for the nine episodes of lake level history shown in Figure 8b. Shoreline configurations for each episode are depicted as overlays on the bathymetry.

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Appendix E

Newman, W.S.; Marcus, L.F.; and Pardi, R.R. (1981). "Palaeogeodesy: Late Quaternary geoidal configurations as determined by ancient sea levels," in *Sea Level, Ice, and Climatic Change*, IAHS Publication No. 131, Wallingford, UK: IAHS Press.

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Palaeogeodesy: late Quaternary geoidal configurations as determined by ancient sea levels

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ABSTRACT Contemporary sea level approximates a gravitational equipotential surface - the geoid. The present global marine strandline is the location of organisms possessing a firm relationship to the local datum. These strandline indicators exhibited similar relationships throughout the Holocene and can be measured relative to the local datum in time and elevation. Since past isochronous marine surfaces must also have been gravitational equipotential surfaces, they describe the deformation of that palaeogeoid at a point in time with respect to the contemporary geoid. The *idee fixe* of the Quaternary hydrologic budget is that ice volume must be related to sea level. We have catalogued some 3000 radiocarbon-dated sea level indicators back to 12 000 years BP. An elevation-time plot of the data indicates those points which might be expected to be eustatic in nature have a spread of nearly 20 m at 5000 years BP and more than 50 m at the base of the Holocene. It follows that "Z" axis eustasy is an elusive and improbable model. Dividing the data into millennial cohorts, we constructed worldwide contour surfaces. We analysed the anomalies and found they have three major sources: area of post-glacial isostatic rebound, plate boundaries, and apparently erroneous data. The contoured surfaces demonstrate a number of transient anomalies such as the progress of glacial isostasy through time, a decaying "equatorial bulge", and a "hole" centred on the northern Hatteras and Sohm Abyssal Plains. The relation of glacial ice volume to sea level remains a riddle. The data demonstrate that the Holocene geoid has apparently suffered chronic instability. Thus mass distribution within the Earth must have changed significantly during the past 12 000 years. Finally, since many aspects of rheological studies depend upon sea level data, both fields will remain naive until additional information is gathered from those areas which are still data deserts.

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INTRODUCTION

Since about 1950, the application of Libby's radiocarbon method of age determination has resulted in the generation of numerous elevation-time plots of the apparent changes in sea level during the past 12 000 years. There is general agreement that glacial eustasy is the dominant factor regulating sea level within this time span. Nevertheless, within a decade of the initiation of late Quaternary sea level dating, Fairbridge (1961) brilliantly summarized all those factors which would tend to diminish the possibility of determining the "ideal" eustatic sea level curve for this time interval. Indeed, it was Fairbridge who first quantitatively delimited the effects of waxing and waning glaciers on hydro-isostasy, the geoid (including asthenosphere channel flow), the gravitation anomalies resulting from both surficial and deep-seated mass transfer, and the isostatic effects of sediment loading. Perhaps the only fault of Fairbridge's (1961) benchmark paper was a reluctance to accept the concept that at least some of the fluctuations in his Holocene sea level curve were probably due to the chronic instability of the geoid.

Newman (1968) suggested that since a dynamic view of the Quaternary period recognizes a fluctuating sea level resulting from the summation of glacial isostasy and crustal warping, both operating at varying rates, it would be rather exceptional to find chronologically equivalent and accordant eustatic shorelines over widely scattered areas. Accordingly, he concluded that coastal mobility was ubiquitous during the Quaternary period. Newman *et al.* (1971) presented quantitative arguments supporting Daly's (1934) concept of an inward migrating and decaying peripheral bulge around former major glacial dispersal centres. However this bulge appears to be a special effect superimposed upon more general modes of geoidal deformation.

It was Walcott who, in a series of three papers (Walcott, 1972a, 1972b, 1973), redirected the course of sea level studies. He pointed out that the eustatic change itself would induce a global deformation that must be considered before proceeding with the inverse problem of determining the eustatic change from the observations of past sea levels. Walcott noted that the geoid itself was probably changing because of the redistribution of mass within the Earth (including glacial rebound). He then proceeded to map these displacements on a global scale based on the available data. Finally Walcott related these results to the nature and viscosity of flow channels within the asthenosphere. Cathles (1975) independently developed a global distortion model similar to that of Walcott but based it on a rather different earth rheology model involving most, if not all of, the Earth's mantle. Faure & Hebrard (1977), summarizing more than a decade of west African sea level investigations, concluded that apparent sea level movements during the Holocene were partly noneustatic and recognized that both tectonic and geoidal distortions were probable causative factors.

Predictions of global sea level changes resulting from deglaciation were calculated by Clark *et al.* (1978) through use of a numerical modelling procedure (Farrell & Clark, 1976). Here again, changes in sea level are caused by redistribution of ice and water loads upon the Earth's surface and of material within the Earth's mantle. Clark *et al.* argue that observed sea level changes are due to the changing distance between the sea level geoid and the floor of the ocean. They reported that vertical sea level changes varied with geographic location in a predictable sequence.

Mörner (1976) directly faced the issue and utterly destroyed the sea level curve "grail". He reasoned that since the ocean surface (the sea level geoid = the gravitational equipotential surface of the Earth) is demonstrably rough and uneven, characterized by humps and depressions of several tens of metres, the present geoid configuration cannot have remained stationary through time. He therefore proposed that diverging Holocene sea level curves are due to changes in the configuration of the geoid.

Newman *et al.* (1980b) used 768 radiocarbon dates, purporting to measure sea level during the past 6000 years, to construct synthetic sea level curves within discrete latitudinal and longitudinal bands. These curves rarely coincide and were often out of phase. This divergence indicates that the search for "the worldwide sea level curve" is futile. The problem then became how to display these data so that we might distinguish patterns unprejudiced by local or spurious data. Experimentation with sixth-order trend surface analysis using these data indicated that this approach would indeed be helpful in our search for global patterns. Quadratic trend surface analyses had previously been employed by McCann & Chorley (1967) for the post-glacial shorelines of western Scotland. Flemming (1978) used cubic trend surfaces to demonstrate both eustatic changes and coastal tectonics in the northeast Mediterranean.

We proposed (Newman *et al.*, 1980b) that the study of the former shapes of the Earth such as those derived from past sea levels, be termed "palaeogeodesy". We have since discovered that the term was originally introduced by Hospers & Van Andel (1970) who computed ancient Earth radii from palaeomagnetic data. "The authors (*op. cit.*, p. 411) suggest that this branch of palaeomagnetic research be referred to as palaeogeodesy". Although we recognize the priority of Hospers & Van Andel, we prefer and offer our more general definition.

MATERIALS AND METHODS

We have assembled nearly 3000 radiocarbon dates purporting to measure the level of the sea going back to 12 000 years BP. Data stored in our computer bank include latitude, longitude, laboratory and laboratory number, date and standard deviation, elevation with respect to the local datum and type of material dated. The major source of our data is the journal *Radiocarbon* (1959-1979). Data were also contributed by J. T. Andrews (University of Colorado), R. I. Walcott (formerly at the Department of Energy, Mines and Resources, Canada), P. Pirazzoli (Laboratoire de Géomorphologie, de l'Ecole Pratique des Hautes Etudes, Montrouge), and H. Faure (Laboratoire de Géologie du Quaternaire, Marseille). These data are accepted as received with no attempt to standardize as, e.g. using a single half-life for radiocarbon. We also accept the local datum reported by the contributing investigator(s). A normalizing procedure might only serve to mask geoidal anomalies. Note that the contemporary geoid is defined by sea level, i.e. the "sea level geoid". We have not corrected the data so as not to prejudice our results. Information for appropriate corrections is either unknown or is not provided along with the published data.

Figure 1 shows bar graphs displaying the distribution of our current data holdings through time, latitude and longitude respectively. The bulk of our data are mid-Holocene in age: we have little data prior to 12 000 years BP. Figure 2 displays the very uneven geographical distribution of our data. These data points roughly outline the world land masses. There are still vast reaches of continental land-mass shorelines unrepresented by data points. The same is true also for most oceanic islands of the world. Figure 3 is a time *vs.* elevation plot of all our data points. We believe that the centre of gravity for most time intervals should plot very near the contemporary datum thus suggesting that isostasy operates on a global scale. However, our data collection is heavily biased as most of the information comes from northern North America and northwest Europe, especially those areas which have undergone postglacial rebound. Examination of Fig. 3 demonstrates that choosing a truly eustatic sea level curve for the latest Quaternary is an improbability. For example, one would expect the "eustatic curve" should be at a negative elevation 10 000-12 000 years BP. However, inspection of Fig. 3 reveals a range of at least 50 m at any point in time below present sea level during this 2000 year interval.

Following our contention that a universal eustatic sea level curve is a manifest impossibility, we have produced a series of world maps of former sea level surfaces for specific short time intervals. Figure 1 indicates that our data bank contains from 100 to 300 data points for each millennial interval over the last 12 000 years.

We experimented with trend surface analysis to fit sea level with first to sixth order polynomial surfaces. This procedure involves fitting surface features of the order of thousands of kilometres in diameter. Such a method cannot sufficiently resolve those features one might expect on geological evidence. On the other hand, we should be able to see broad features such as the postglacial uplift of the Fennoscandian Shield. There is further appeal in that the resolution is nearly of the same order of geoidal deformation as that theorized by Clark *et al.* (1978). Therefore, we hypothesize that the trend surfaces we produce should have had some functional relationship to geoidal deformation, and that short wavelength features, such as those produced by regional tectonics, should show up as residuals from the higher-order surfaces.

In our original work, we developed a program using SAS76 (Barr *et al.*, 1976) to provide contour maps of the fitted surfaces. We are now using the computer packages SYMAP and SYMVU (Dougenik &

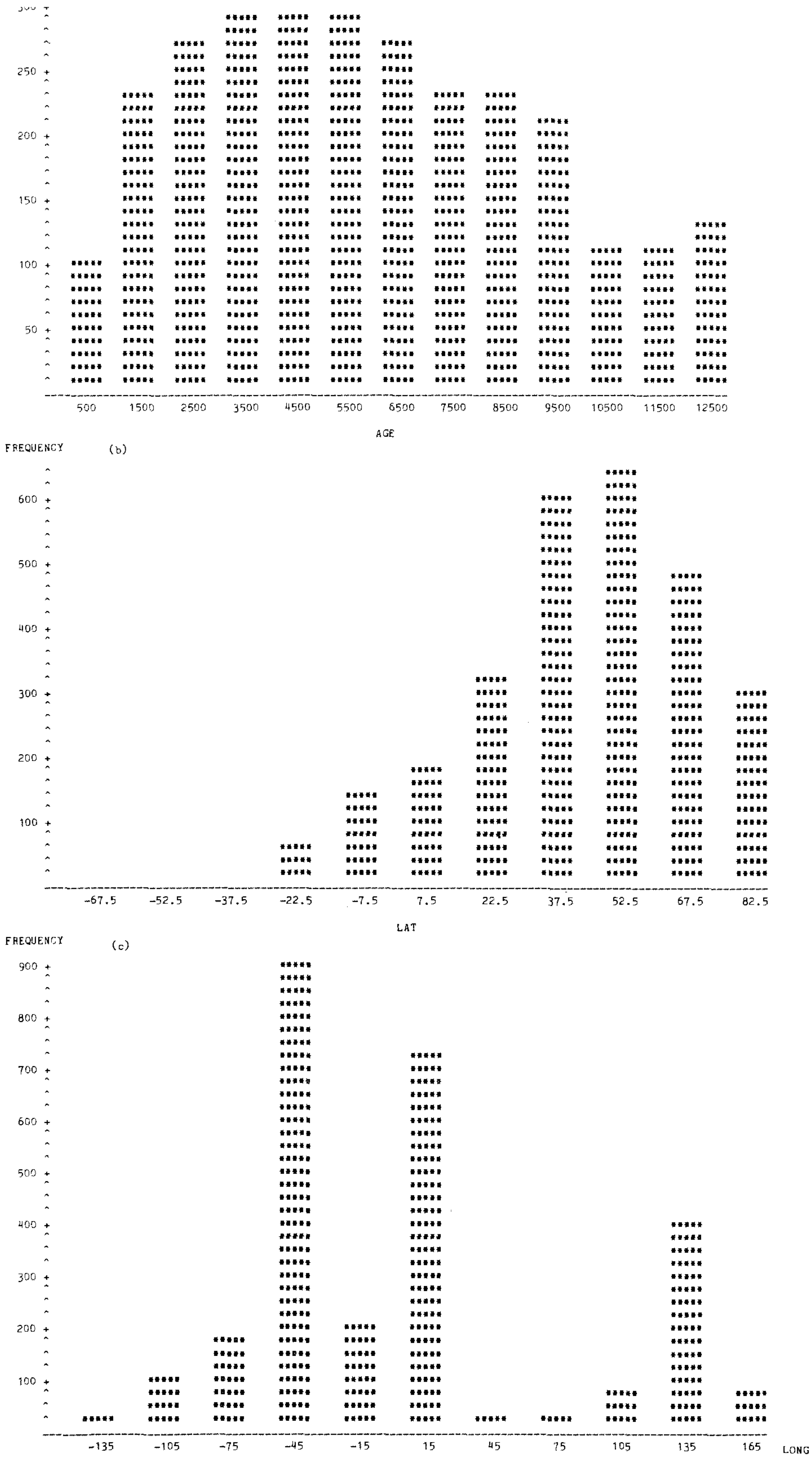
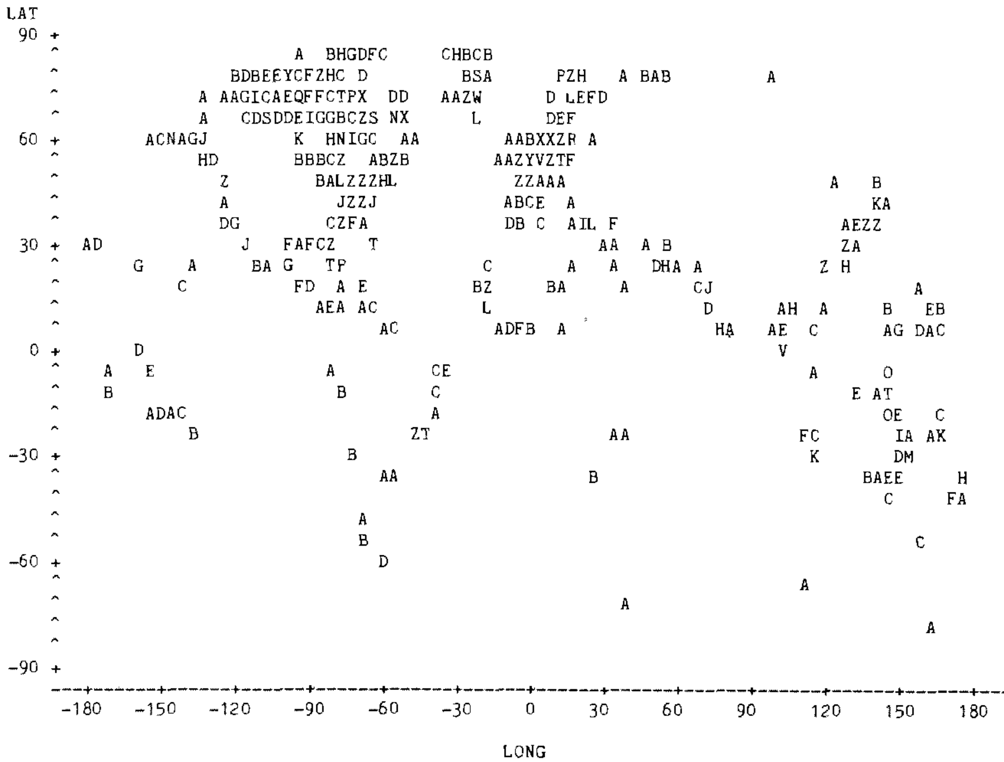
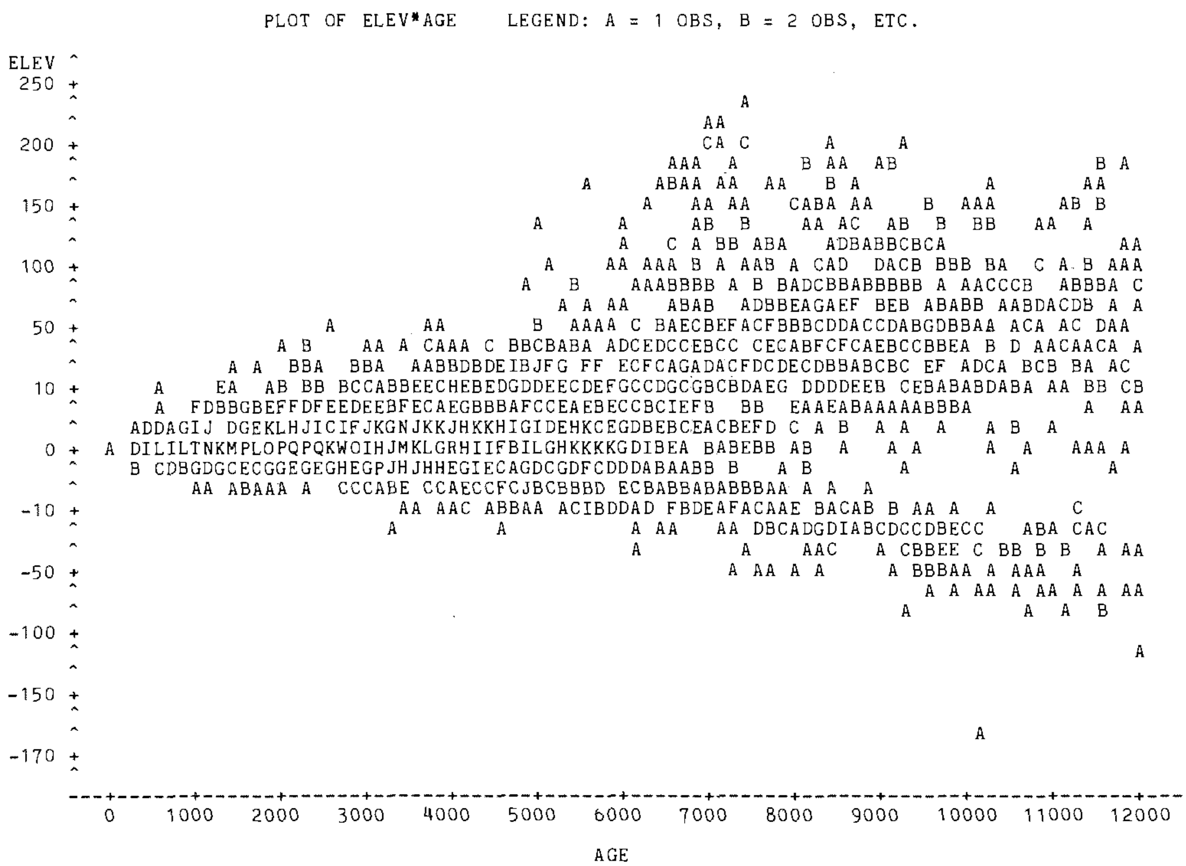


Fig. 1 Bar graphs of radiocarbon-dated sea level data through time and geographical coordinates. (a) Distribution through time in 1000-year intervals. (b) Distribution by latitude. (c) Distribution by longitude.



NOTE: 1 OBS HAD MISSING VALUES OR WERE OUT OF RANGE 718 OBS HIDDEN

Fig. 2 Geographic distribution of radiocarbon-dated sea level data. The data roughly outline the world's landmasses. A = one observation; B = two observations; etc. Note the many areas lacking data.



119 OBS HAD MISSING VALUES OR WERE OUT OF RANGE

Fig. 3 Time *vs.* elevation distribution of radiocarbon-dated sea level. The bulk of the early Holocene data orginate in those areas which have undergone postglacial isostatic rebound. The postglacial marine transgression is represented by the negative elevation data in early Holocene times. Note that the range of data at any point in time in this sector of the plot is at least 50 m.

Sheehan, 1975) which not only produce contour maps and trend surfaces but also can produce hidden line perspective plots of three-dimensional surfaces. In our earlier analyses we were faced with the severe limitations imposed by the data anisotropy - namely, that trend surface techniques tend to extreme values in those regions where the data are sparse or nonexistent. Taking advantage of SYMAP interpolation routines, we attempted to analyse data spread over an evenly spaced grid. Sea levels were interpolated at the centre of cells (representing 3° of latitude and 4° of longitude) by a weighted average of points near the centre of each cell. These weights are inversely proportional to the distance of a data point from the centre of a cell. We then used SYMAP to contour this interpolated data set and SYMVU for hidden line plots. The grid of 90 by 60 points was used as a new data set for trend surface analysis. This smoothing procedure had the advantage of keeping the trend surface in bounds over the entire globe, but still retained the limitations discussed in our earlier papers. The interpolated data overwhelmed and dominated the trend surfaces and were not useful for interpretation of past sea level surface reconstructions. Hence, we proceeded to generate sixth order trend surfaces based only on the actual data. This procedure also lead to problems.

Figure 4 shows graphic displays of our contour surfaces for the millenial intervals from 0 to 12 000 years BP. Superimposed on these diagrams are outlines of the world's land masses. The diagrams are contours drawn upon the actual data.

RESULTS

The most dramatic feature of our graphic display is the upward movement of those areas once occupied by or presently still covered by Northern Hemisphere ice sheets. This conspicuous doming extends from the western Canadian archipelago eastward to Novaya Zemlya. Presumably, most of this uplift is due to isostatic rebound. Another rising feature can be traced from Japan southwest to Taiwan, then south and southeast to New Guinea. We believe this ridge marks the boundary between several tectonic plates: specifically the Pacific, Eurasian and Philippines Plates. Deflections from the present geoid are also conspicuous along several segments of other plate boundaries.

On the other hand, a conspicuous depression has developed off the northeast coast of the United States in the western North Atlantic Ocean. We can think of three possible causes for this phenomenon. The sinking could be simply the downwarping of a passive continental margin. Or it could be caused by peripheral bulge subsidence in the sense of Daly (1934). The third possibility is sediment isostasy caused by the deposition of glacial outwash into the North Atlantic basin by Laurentide glacial meltwaters. The later possibility is discussed by Newman *et al.* (1980a). Two or three of these processes may operate simultaneously.

Our graphics also suggest that some regions of the Earth have changed direction of vertical motion. Rising areas were subsiding

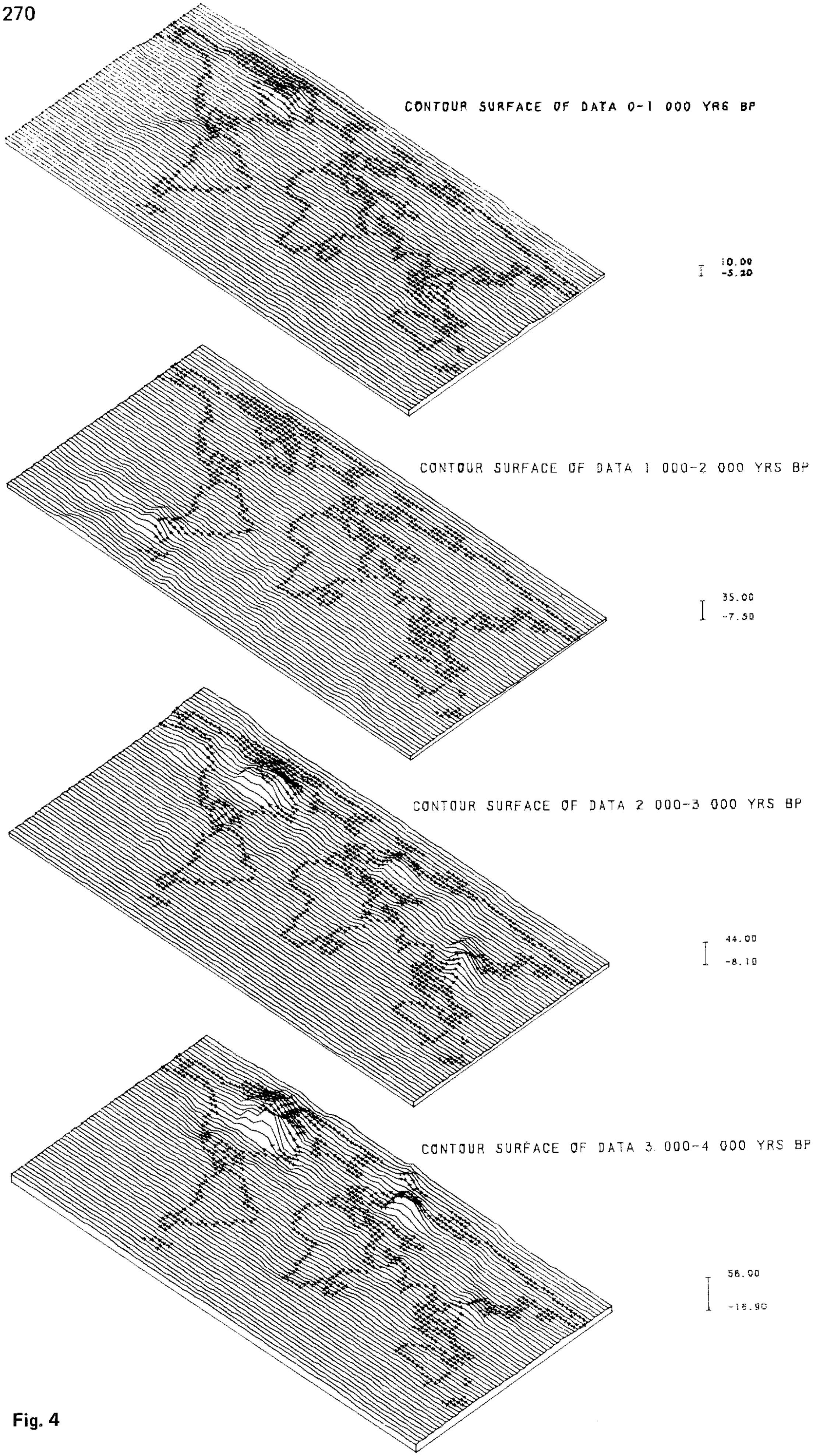

CONTOUR SURFACE OF DATA 0-1 000 YRS BP
10.00
-5.20
CONTOUR SURFACE OF DATA 1 000-2 000 YRS BP
35.00
-7.50
CONTOUR SURFACE OF DATA 2 000-3 000 YRS BP
44.00
-8.10
CONTOUR SURFACE OF DATA 3 000-4 000 YRS BP
58.00
-15.90

Fig. 4

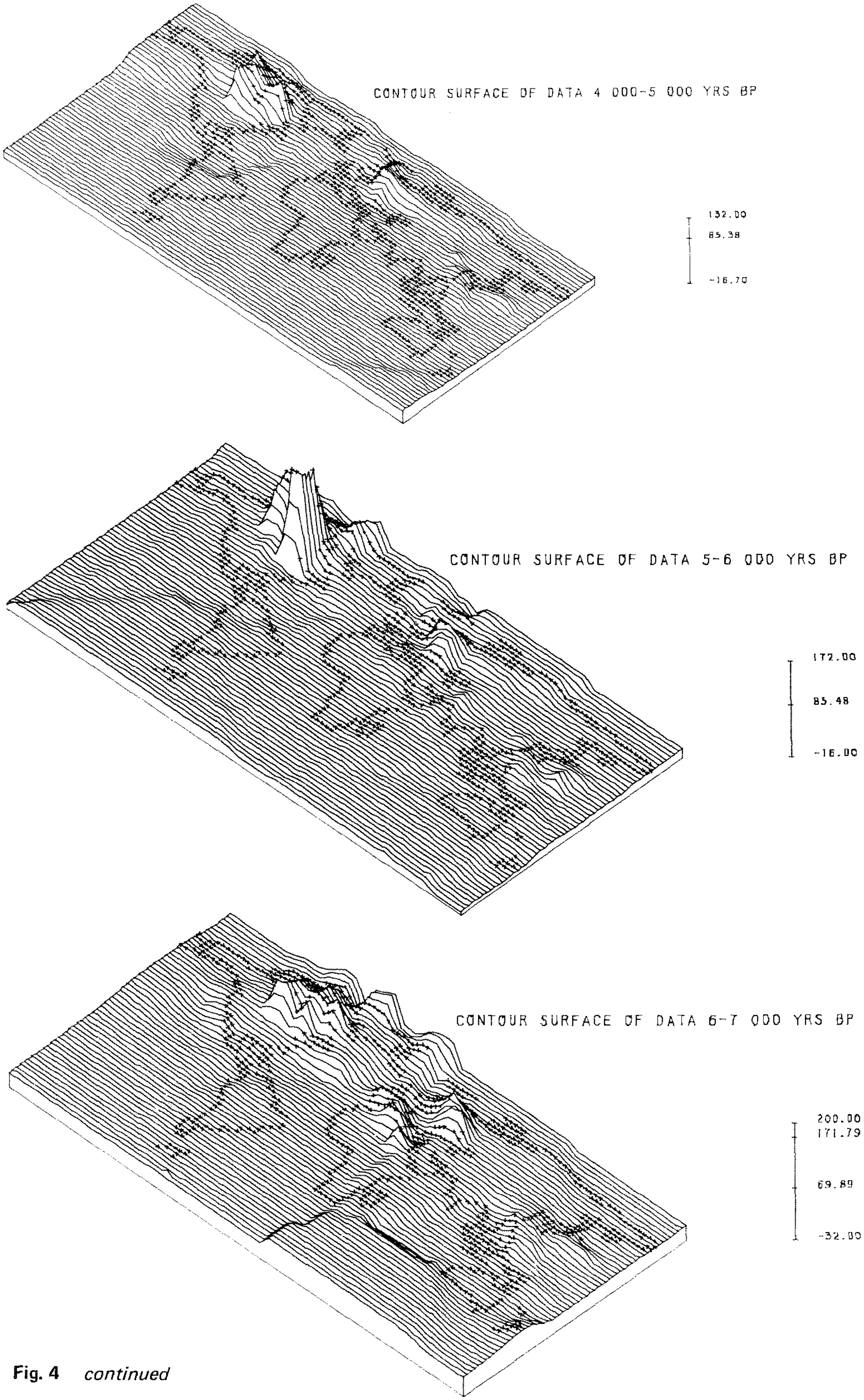


Fig. 4 *continued*

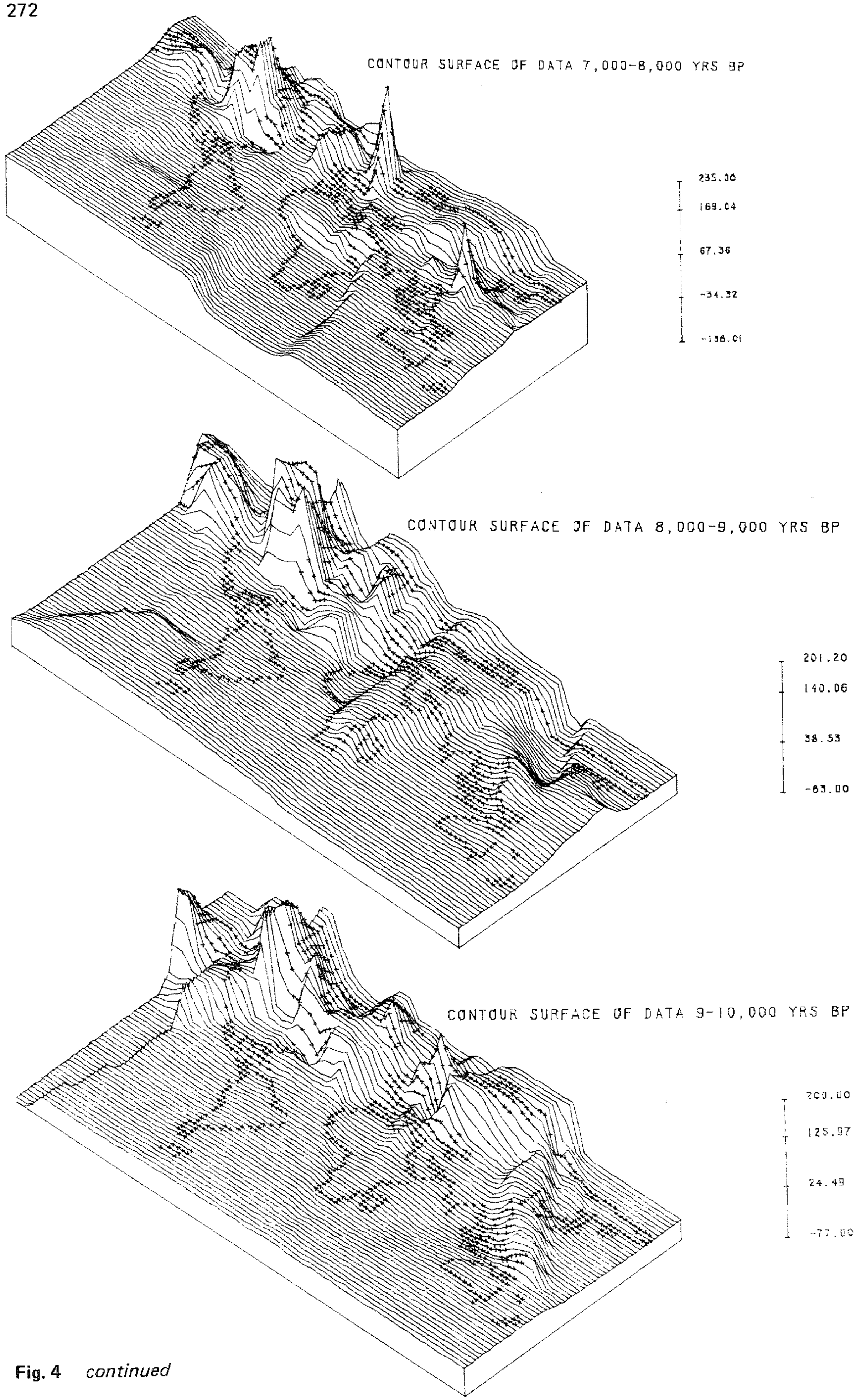


Fig. 4 *continued*

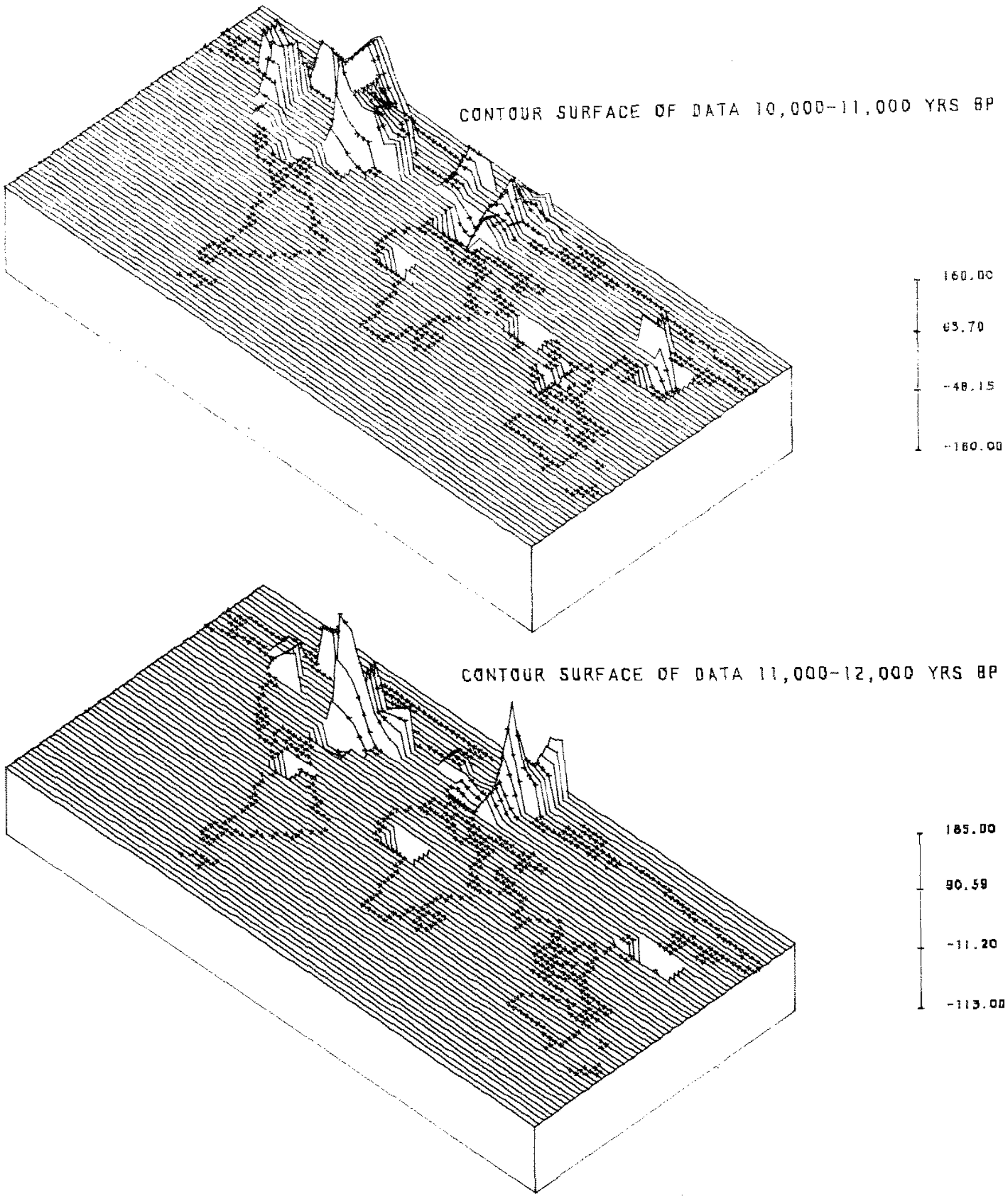


Fig. 4 Contours of actual data for millenial intervals 0–12 000 years BP. The vertical scales depict range of elevation in metres for each of the diagrams. The early Holocene disarray is due both to data maldistribution as well as the increasing departure of the contemporary geoid from that of early Holocene times. Plate boundary anomalies are especially evident.

in earlier Holocene times while once rising areas are now subsiding. However, at least some of these motions are spurious in that they result from either a maldistribution of data points within our 1000 year time frame or a regional absence of data.

Still another unexpected feature disclosed by our trend surfaces is a persistent Holocene equatorial bulge. Although this feature may be in part an artifact of our graphical development, its constancy through several millennia suggests to us that the feature is real and is still another facet of global mass redistribution, a result of the change from glacial to interglacial regime. This mass redistribution may be responsible for a change in the rotational velocity of the Earth which, in turn, may have caused a change in the shape of the sea level geoid.

Indeed, this change in the figure of the geoid is undoubtedly still underway.

CONCLUSIONS

Our statistical and graphic analysis of sea level data demonstrates that the geoid is chronically unstable. A single eustatic curve cannot be determined from the study of sea level changes through time because such a curve is, at best, regional in scope. We cannot, therefore, determine ice volumes of the past using sea level curves. However, sea level data can be used to determine the past configurations of the geoid, thus opening up a new field of investigation which we term "palaeogeodesy".

Finally, it is evident from our studies and cited analyses that the globe has been altering its shape over the past 12 000 years. However, the absence of a coherent data analysis technique and the lack of an adequate data base have prevented the testing of realistic models to explain these changes. We believe that our graphical and statistical data analysis will help in the development of better deformation models for the Holocene and also point out priority data needs for future refinement.

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Appendix F

Vaughan, R.C. (1994). "Geologic and Hydrologic Implications of the Buried Bedrock Valley that Extends from the Western New York Nuclear Service Center into Erie County, N.Y.", in *Geology Reports of the Coalition on West Valley Nuclear Wastes*, East Concord, NY: Coalition on West Valley Nuclear Wastes, 1994.

GEOLOGIC AND HYDROLOGIC IMPLICATIONS OF THE BURIED BEDROCK VALLEY THAT EXTENDS FROM THE WESTERN NEW YORK NUCLEAR SERVICE CENTER INTO ERIE COUNTY, N.Y.

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January 16, 1994

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Introduction

Geologists working for the West Valley Demonstration Project have recently begun to consider the hydrology and certain other aspects of the buried bedrock valley that lies beneath the Western New York Nuclear Service Center. To date, their treatment of the subject is a meager first step; it falls short of the level of understanding needed for a good characterization of the site. It neither answers, nor notes the need for answering, a variety of questions associated with the buried bedrock valley. This report will review and discuss some of the questions.

When the 3345-acre Western New York Nuclear Service Center was created in the early 1960s, the buried bedrock valley beneath the site was identified and profiled by a series of borings and seismic reflection tests. Unpublished work by D. S. Hodge and/or Parker Calkin has subsequently shown that the buried bedrock valley is connected to the buried bedrock valleys of two northward-flowing creeks in Erie County: the west branch of Cazenovia Creek and the main branch of Eighteenmile Creek. The hydrology of this subsurface connection will be one of two main topics covered here.

The other main topic will be the unstable lacustrine soil or sediment layer that is found north of the Western New York Nuclear Service Center in both of the above-mentioned valleys: the west branch of Cazenovia Creek and the main branch of Eighteenmile Creek. Discussion will focus on whether a similarly unstable sediment layer occurs in the buried bedrock valley beneath the Western New York Nuclear Service Center.

Overview of hydrology

There are two distinct hydrologic pathways that carry water away from the 3345-acre Western New York Nuclear Service Center. One is the surface pathway, fed by both surface runoff and certain groundwater sources, that goes westward—toward Zoar Valley, Gowanda, the Seneca Nation, and Lake Erie—by way of Buttermilk and Cattaraugus Creeks. The other is a subsurface or aquifer pathway, fed by groundwater from Dutch Hill and hillsides east of Buttermilk Creek, that carries water northward toward aquifers in

Erie County by way of the buried bedrock valley under the Western New York Nuclear Service Center.

The first of these pathways is relatively well known. Indeed, one of the advantages sometimes claimed for the nuclear site is that any water from the site follows a predictable path down Buttermilk Creek before leaving the site. The fact that Buttermilk Creek passes through a narrow channel affords an opportunity for monitoring any contaminants leaving the site, according to those who make this claim.

This claim is essentially true as far as it goes, but its limitations need to be recognized. One limitation is that any system for monitoring contaminated water in Buttermilk Creek would be a relatively "late warning" system; as such, it has little relevance under current state and federal regulations for radioactive waste management. A second, more serious limitation is that the claim applies only to the central part of the Western New York Nuclear Service Center. This is where the thick, relatively impermeable layers of Lavery and Kent Till are located. These layers seem to act as effective plugs across part of the Buttermilk Creek valley, diverting water horizontally and preventing all but a trickle of downward flow. But these thick "plugs" occupy only the central portion of both the valley and the nuclear site.

Hillsides on the eastern and western portions of the Western New York Nuclear Service Center are outside the thick "plugs" of till. Much of the water from these hillsides flows down around the "plugs" and enters the deeper groundwater system that flows northward into Erie County. The thick "plugs" of till, despite their effectiveness as barriers to downward flow, are not well sealed to the hillsides. Groundwater is thus able to flow around their edges into the buried bedrock valley that lies hundreds of feet below the nuclear site.

Details of hydrology

Figure 1 is a map of the Western New York Nuclear Service Center, taken from the local topographic map (Ashford Hollow quadrangle). Figures 2 and 3, taken from Teifke (1993), show cross-sections of the site along the lines **X-X'** and **Y-Y'** drawn on Figure 1. Note that the vertical scale is very exaggerated in both Figure 2 and Figure 3. Note also that the upper layer of "decomposed" bedrock is shown only intermittently but actually constitutes a continuous aquifer, or flow pathway, formed of weathered and highly fractured bedrock in a layer about 10 feet thick.

Thick till layers can be seen in the Buttermilk Creek valley in Figures 2 and 3. Thick till also occupies a small parallel valley in Figure 3. Relatively thin till layers can be seen on the hillsides. In general, drainage from the thick till follows the surface pathway down Buttermilk Creek, then westward along

Cattaraugus Creek to Lake Erie. Drainage from the thinly mantled hillsides goes partly to the surface pathway and partly to the deep aquifer at the bottom of the buried bedrock valley.

A rough analogy to these two flow regimes is provided by a group of familiar items from a household kitchen. Let a large wok or bowl and two plates—large and small—be placed as shown in Figure 4. The plates represent the thick layers of Lavery and Kent tills, while the bottom of the bowl represents the bottom of the buried bedrock valley. Water sprayed into the bowl will fall partly on the top plate (the Lavery till) and partly on the sloping inside surface of the bowl (the adjoining hillsides).

Water that falls on the sloping inside surface of the bowl will flow downward. When it encounters the rim of the larger plate, part will be diverted onto the plate and part will continue to run downward. Upon encountering the rim of the smaller plate, part will be diverted onto the smaller plate and part will continue to run downward toward the bottom of the bowl.

Water that collects on the two plates is diverted to Buttermilk Creek, then westward along Cattaraugus Creek. Water that reaches the bottom of the bowl flows northward along the buried bedrock valley toward aquifers in the buried bedrock valleys in Erie County.

Figure 5 shows contours of the buried bedrock valley under the Western New York Nuclear Service Center. Figure 6 shows groundwater flow patterns in the upper, decomposed layer of bedrock. It is evident from Figure 6 that the predominant flow carries groundwater from the upper, decomposed layer of bedrock into the buried bedrock valley under the Western New York Nuclear Service Center. This flow process is outlined in the <u>West Valley Demonstration Project Site Environmental Report for Calendar Year 1992</u>, prepared by West Valley Nuclear Services and Dames & Moore (1993), page xxxix:

> The uppermost weathered bedrock is a...water-bearing unit that consists of fractured and decomposed shale and rubble ranging in thickness up to 3 meters (10 ft) along the top of the solid, unweathered bedrock.... Groundwater from [this] aquifer tends to move east toward the lowest point of the valley, about 300 to 350 meters (980 to 1,148 ft) west of Buttermilk Creek, and may emerge to flow north-northwest as surface water.

The above description is somewhat misleading for two reasons. First, while it is true that the "lowest point" or thalweg of the buried bedrock valley lies west of Buttermilk Creek in the vicinity of the old nuclear facilities, it should be noted that in a down-valley or northward direction the thalweg crosses under Buttermilk Creek to its east side. The configuration is shown in Figure 5. Second, the above description is confusing, or simply mistaken, in its assertion that groundwater that has reached the "lowest point" or thalweg "may emerge to flow north-northwest as surface water." Groundwater that has reached the thalweg is more than 200 feet below local creeks and the local ground surface; we see no evidence that it can emerge, or has emerged, as surface water. On the contrary, it moves down the thalweg toward the aquifers lying to the north in Erie County.

Connection to the Springville aquifers

Additional description can be found in the Geology Environmental Information Document that was recently prepared by Teifke (1993). On page 18 of Part 1, for example, Teifke explains the recharge process: "The mantle of basal till immediately above bedrock throughout the region and the site area apparently functions as a leaky seal or aquitard through which the bedrock aquifer is recharged on the gentle slopes and summit levels..."

Despite the detail he provides, Teifke (1993) presents contradictory information on this topic in different parts of the same report, with no attempt at resolution. In particular, he presents a confusing or misleading overall picture of the pathway followed by water that has entered the bedrock aquifer through the hillslopes and hilltops of the Western New York Nuclear Service Center. Teifke correctly notes (pp. 18 and 20) that some of this water escapes to the surface, and he acknowledges in his figures and text (p. 18) that some water reaches the thalweg of the buried bedrock valley, yet he fails to state clearly where the water goes after reaching the thalweg. On pp. 13-14 of Part 2, and in Fig. 6-1 of Part 1 and Fig. 2-2 of Part 2, he shows unequivocally that the buried bedrock valley under the Western New York Nuclear Service Center is physically connected to the buried valley in which the Springville aquifers lie, yet he fails to note or discuss a hydraulic connection. (See Figure 7 of this report, taken from Fig. 2-2 of Part 2 of Teifke, for a diagram showing the physical connection of the buried valleys. Evidence for this connection is based on unpublished work by D. S. Hodge and/or Parker Calkin and on published work that Calkin has done in a parallel valley. See Teifke 1993; Calkin and Muller 1980.)

Although he does not discuss a hydraulic connection, Teifke does discuss the fact that the Springville and Sardinia aquifers are similar in structure to the buried bedrock valley that lies beneath the Western New York Nuclear Service Center.

Wilson (1990) provides a more detailed view of the Springville aquifers. Lying beneath the village of Springville, these aquifers are stacked up in multiple layers (alternating layers of till and more permeable materials) more or less like those under the Western New York Nuclear Service Center. Information provided by Teifke (1993) and Wilson (1990), especially in combination, implies that the deepest of the Springville aquifers is hydraulically connected to the bedrock valley that lies beneath the Western New York Nuclear Service Center. The previously described deep flow of groundwater from the nuclear site must therefore enter the deepest of the Springville aquifers, roughly 400 to 500 feet below the surface.

The village of Springville obtains its water from a shallower aquifer in the same group. (The municipal wells draw from a depth of about 145 feet below the surface.) Information is lacking on the degree of hydraulic connection, and the direction of flow, between the aquifer that supplies Springville and the underlying aquifer that is hydraulically connected to the buried bedrock valley beneath the nuclear site. It should be noted, however, that the calculated recharge to the village water supply is not fully accounted for (see Wilson 1990). Thus, the possibility exists that some of the recharge to the municipal wells may be coming from the deeper layers that receive water from the Western New York Nuclear Service Center. It should also be noted that one Springville industry, Robinson Knife Company, has its own well that is substantially deeper than the municipal wells: about 385 feet below the surface. The probability of recharge from the nuclear site is greater for the deep Robinson well than for the shallower municipal wells.

There is a slight possibility that the buried bedrock valley discharges most of its groundwater to Cattaraugus Creek before reaching Springville, in which case there would be little or no hydraulic connection between the nuclear site and the aquifers that lie beneath Springville. This idea cannot be excluded, yet both the lack of evidence and the physical configuration argue against it. Geologists associated with the West Valley Demonstration Project hint at the idea, but in the absence of evidence seem unable or unwilling to engage in substantive discussion.

Noticeably absent from the reports of the West Valley Demonstration Project geologists is any type of water budget for the buried bedrock valley. The general approach of these geologists is to avoid any focused discussion of a destination for groundwater that has reached the thalweg of the buried bedrock valley under the nuclear site. For example, a hydrology report by Aloysius et al. (1993, p. 64 and Fig. 5-24) suggests in a general way that groundwater in bedrock valleys "tends to flow...upward near major streams," apparently meaning that some or all of the groundwater from the buried bedrock valley is discharged to Cattaraugus Creek. Teifke (1993, Part 1, p. 18) goes one step further and suggests, without any explanation or discussion, that groundwater from the buried bedrock valley is discharged to the surface at the confluence of Cattaraugus and Buttermilk Creeks.

From the evidence presented above, it is clear that the thalweg of the buried bedrock valley passes beneath Cattaraugus Creek. Elevation of the northward-trending thalweg is about 800 or 900 feet above sea level; elevation of the westward-flowing creek at this location is about 1100 feet. Thus, the thalweg lies 200 to 300 feet beneath the creek. The point at issue is whether a significant quantity of the groundwater flowing along the thalweg rises 200 to 300 feet to be discharged into Cattaraugus Creek, or, alternatively, whether most of this water continues flowing northward along the buried thalweg toward the Springville aquifers and other aquifers in Erie County. Either possibility is physically plausible, but no coherent evidence has been presented in favor of the first possiblity by any of the above-quoted authors. The facts as presented favor the second possibility: that the major flow carries groundwater along a continuous bedrock valley that extends northward from the Western New York Nuclear Service Center, passing under Cattaraugus Creek and through the deepest Springville aquifer.

From Springville the flow of groundwater apparently continues northward through Erie County, following one or both of the buried bedrock valleys indicated in Figure 7. The main aquifer evidently follows the west branch of Cazenovia Creek and flows beneath the hamlets of Footes, Glenwood, Colden, and West Falls, and perhaps as far as the village of East Aurora. Teifke (1993, Part 2, p. 13) indicates that the thalweg of the buried bedrock valley is at 700 to 800 feet elevation between Springville and Footes; this fits the generally accepted idea that the buried bedrock valley descends or dips toward the north. Presumably the drainage direction for the buried bedrock valley is also to the north, as one source cited by Teifke suggests. It should be noted, however, that the northward-dipping bedrock valley is cut into bedrock strata that dip toward the south or southwest. At some point, groundwater flowing northward through the buried valley may encounter strata that are sufficiently fractured to accommodate most or all of the flow, in which case the main flow will zigzag back toward the south or southwest at a deeper level.

In summary, the groundwater system that is fed by the hillslopes and hilltops of the Western New York Nuclear Service Center is complex and largely unknown. Figure 8 shows the areas of the Western New York Nuclear Service Center that, in our best estimate, serve as recharge areas for this groundwater system. These recharge areas are roughly delineated in Figure 8; further work is needed to define their boundaries more precisely.

More work is also needed to trace the path of the groundwater through Erie County and to determine the various connections, branches, and residence times within this complex aquifer system. Such work is needed as a preliminary step toward the main task of preparing an overall water budget for this aquifer system. The preparation of a defensible water budget is essential for characterization of the Western New York Nuclear Service Center, given the large portions of the site that are either underlain by, or serve as recharge areas for, the aquifer.

Unstable lacustrine sediments

The above-described system of interconnected valleys is of interest not only for its hydrology but also for the unstable lacustrine sediments that are found north of Springville in this valley system. Owens et al. (1977) describe the occurrence and properties of these unstable sediments and compare them to the sensitive or quick clays of Canada.

This report will examine several types of evidence for the occurrence of such unstable sediments beneath the Western New York Nuclear Service Center. If unstable lacustrine sediments of this type do indeed occur there, either under or in close proximity to the nuclear site, the long-term integrity of the site may be threatened.

Occurrence of unstable lacustrine sediments in Erie County

The unstable lacustrine sediments studied by Owens et al. (1977) are located a few miles north of Springville in the valleys of the west branch of Cazenovia Creek and the main branch of Eighteenmile Creek. Except where exposed by subsequent erosion, these sediments occur as subsurface layers. The overlying soils are typically stable and provide no clear indication of the problems of mass movement that may occur in the lacustrine sediments below. Among the observations mentioned by Owens et al. (1977) are "some nearly continuous areas of lacustrine sediments that are moving or have moved in the last few years." As an example of the resulting problems, Owens et al. mention "considerable difficulty in maintaining the roadbed of a railroad." The railroad in question is the same railroad that passes through Springville and the Western New York Nuclear Service Center.

In more detail, Owens et al. (1977) describe the properties of, and some of the problems associated with, these unstable lacustrine sediments:

> The lacustrine sediments have a thin skin of brown soil over a firm to soft, in places nearly fluid, body of dark gray silt and clay. Road cuts that expose these soft sediments are unstable, and the road ditches require periodic cleaning to remove the sediment that moves from the road cuts. The B&O Railroad has had several derailments along a short stretch of track built on these lacustrine sediments in the West Branch of Cazenovia Creek.

Problems with the lacustrine sediments tend to occur 1) when their relatively low load-bearing capacity is exceeded and 2) when they are either exposed or sufficiently near the surface that an outward path exists along which mass movement may occur. Owens et al. note that the "mica or illite in the clay and silt fraction controls the behavior of these sediments as engineering materials" and point out that "The clay and silt mineralogy of the lacustrine sediments is the same as that of the shale bedrock in the area." They find the lacustrine sediments to be similar to the sensitive or quick clays found in Canada in terms of texture, mineralogy, liquid limits, and plasticity index, though they acknowledge that the Canadian clays are capable of much more rapid movement.

Indications of unstable soils in or near the nuclear site

Several types of evidence will be presented here for the occurrence of similar types of unstable lacustrine sediments, either under or immediately adjacent to the Western New York Nuclear Service Center. Some of the evidence is anecdotal; it is not intended as proof but, in combination with the other evidence, as an urgent warning of the need for full investigation. It should be noted that the term "quicksand," when used anecdotally, may include various materials such as quick clays that exhibit flow properties similar to those of quicksand.

Various lacustrine sediments are already known to exist beneath the Western New York Nuclear Service Center. Some of these are indicated by the stippled areas in Figures 2 and 3. As of early 1993, the geomechanical properties of these known lacustrine units had not been fully investigated (Teifke 1993, Part 3, p. 3 ff.). Deeper lacustrine layers may also exist; these have been neither identified nor investigated. According to Teifke (1993, Part 2, p. 15), "the complete sequence of glacial deposits has not been penetrated by the drill. Comparison of borehole records and seismic reflection data indicate an unexplored thickness of at least 50 meters and as much as 100 meters of older glacial deposits along the axis of the bedrock valley." This unexplored thickness is designated "ALT" or "Pre-Kent Till" in Figures 2 and 3, but the actual sequence is unknown.

Problems of mass movement may occur if any of the lacustrine units beneath the Western New York Nuclear Service Center exhibit the properties observed by Owens et al. (1977) in Erie County. Owens et al. make the general observation that "Mass movement of the lacustrine sediments on the lower slopes can affect large areas upslope." They note that one of the figures in their paper "illustrates a large area of the Cazenovia Creek Valley that is actively moving or slumping because a small stream at the base of the slope started to undercut the soft lacustrine sediments. If the slumping extends upslope it will include the loams, sands, and gravels of the glaciofluvial cap that usually are considered stable materials." In general, mass movement of unstable sediments will disrupt overlying or upslope deposits; hence the concern about unstable sediments beneath the nuclear site.

The following logic suggests that unstable sediments do occur beneath the nuclear site:

--Effects similar to those described by Owens et al. are observed within or adjacent to the Western New York Nuclear Service Center.

--Soft clayey sediments which appear (to the untrained eye) to be generally similar to those described by Owens et al. are observed within or adjacent to the Western New York Nuclear Service Center.

--The same raw materials (i.e., the shale bedrock) and the same formative agents (glaciers) were available at the Western New York Nuclear Service Center as at the locations observed by Owens et al.

Possible evidence for the occurrence of unstable lacustrine sediments beneath the Western New York Nuclear Service Center includes:

1. Persistent problems in past years in maintaining the roadbed of the railroad at or near the northwest boundary of the nuclear site, at an elevation of about 1280'. The relevant section of track extends from Cascade Bridge (location "A" in Figure 9) to some point further south (perhaps location "B" in Figure 9). As described in a history of the railroad by Pietrak (1979),

> Another drawback the construction crews ran into on the Buffalo Branch was quicksand. While the track had been laid for several months on the section from Ashford Junction to Buffalo [ca. 1883], it was almost impossible to keep it up to grade for a good distance of the line just south of the Cascade Bridge. This area known as Buttermilk Swamp, kept on sinking due to the fact that there was quicksand under the clay soil. Because of the track settling, the R.& P. was obliged to employ two steam shovels with four construction trains for a period of several months hauling in gravel and building up the roadbed. This section of road was under constant watch and even up to the end of the B.R.& P. when the B.& O. took them over [ca. 1930], a watchman was kept on twenty-four hour duty making sure the roadbed was in good shape and did not sink.

2. Abandonment of the Schichtel #1 gas well that was being drilled in 1975, apparently because the drillers encountered quicksand and other unstable materials. (Location "D" in Figure 9.) According to reports filed with DEC by the driller, the well was intended to be 3200 feet deep but was abandoned at a depth of 322 feet "due to thick (unknown amount) glacial material, gravel and quicksand." (DEC 1975) Since the surface elevation at the drilling site was 1330', the quicksand was encountered at about 1008' elevation. It should be noted that one of the drillers claimed a few days later that drilling of this particular well was stopped when radioactivity was encountered. This story was publicized in December of 1975 by Channel 7 in Buffalo, investigated immediately by DEC, and denied as untrue. (DEC 1975) If it were true, it might relate to the hydrology discussion above. Note that quicksand and radioactivity are not mutually exclusive explanations. Some local residents who have heard both reasons for the abandonment of drilling are inclined to believe that quicksand was encountered, but they are not convinced that it would be a sufficiently serious or unusual problem to stop the drilling of a gas well.

3. Observation and photos, by the author and K. McGoldrick, of mass movement along the lower reach of Buttermilk Creek, especially at the sharp bend in creek designated "C" in Figure 9. The elevation here is about 1140'. The mass movement is occurring over a broad area on the east bank of the creek, in medium gray clayey material that appears soft, at the foot of a relatively gentle slope. (It is quite different in appearance from the slumping, landsliding, or mass wasting of banks along Buttermilk Creek that are steeper and apparently consist of stiffer material, as can be seen further upstream near the SDA burial grounds, for example.)

4. Observation and photos, by the author and others, of mass movement on the east bank of Buttermilk Creek, a few hundred feet downstream from the railroad bridge (location "E" in Figure 9). The elevation here is about 1200'. This mass movement is occurring over a broad area, in medium gray clayey material that appears soft, at the foot of a relatively gentle slope. (Again, this is quite different in appearance from the slumping, landsliding, or mass wasting of banks along Buttermilk Creek that are steeper and apparently consist of stiffer material.)

5. Observation and photos, by the author, C. Mongerson, and B. Cooke, of mass movement along Gooseneck Creek over most of the distance from Riceville to the Western New York Nuclear Service Center boundary, at elevations between about 1330' and 1400'. According to local resident E. Zimmerman (1993), there have been longstanding problems of slumping in this area, including slumps that affected both Buttermilk and Fox Valley Roads. Gooseneck Creek is identified by the letter "G" in Figure 9. The area described in this paragraph lies immediately east of the Western New York Nuclear Service Center boundary, just off the right edge of the map in Figure 9.

6. Reports of mass movement along Gooseneck Creek within the Western New York Nuclear Service Center, including a report (Zimmerman 1993) that slumping or landsliding blocked the railroad tracks several years ago at a location slightly south of Buttermilk Road (approximately at location "F" in Figure 9). The creek elevation here is about 1300' to 1330'; track elevation is about 1300'.

7. Reports of quicksand encountered in construction projects in and near the hamlet of West Valley, particularly in the construction of West Valley High School in 1935 or 1936. According to local resident L. Gerwitz (1993), quicksand was encountered in building the foundation of the school and was a serious enough problem that it almost caused the school construction project to be abandoned. According to Zimmerman (1993), whose father worked as a mason on the school construction project, work began on the foundation of the school at a location relatively close to the highway (NY 240), but, due to quicksand problems, the site for the school was moved further from the highway to the current location. Both locations are at about 1530' elevation.

Liquefaction potential of sediments beneath the nuclear site

Unstable sediments, especially those that lie beneath a more stable layer, are sometimes not recognized until an earthquake occurs. A problem that may happen then is called seismically induced liquefaction. Unstable (or quasi-stable) sediments which are reasonably firm and solid under ordinary circumstances may liquefy and flow under the seismic shock of an earthquake. The results can be dramatic or disastrous, as seen several years ago when groups of buildings built atop unstable sediments collapsed in a San Francisco earthquake. Even though the surface soils may be stable, they will shake and shift like the top of a waterbed as underlying sediments liquefy and flow. When the earthquake is over, the underlying sediments are likely to have shifted from their original positions and may have even spurted out to the surface. As a result, the ground surface is often left in a partially collapsed or heaved condition.

On sloping ground, the seismic shock of an earthquake can cause unstable sediments to undergo more rapid or more extensive mass movement, slumping, etc., than would occur under normal circumstances. At least one case is known in which an overlying layer of relatively stable soil "floated" a short distance down-slope on unstable sediments that liquefied under seismic shock.

Liquefaction of soil or sediment layers at the Western New York Service Center could seriously compromise the integrity of waste-management facilities at the site. For this reason, lique-faction was one of the issues raised five years ago in scoping comments for the EIS that is now being prepared for site closure. One of the scoping comments in Vaughan (1989), for example, indicated the need for testing the liquefaction potential of the lacustrine sediments that lie beneath the layer of Lavery Till at the site. (See Figures 2 and 3. These lacustrine or glaciola-custrine sediments are the uppermost of the two stippled or dotted bands in Figures 2 and 3.) According to Teifke (1993, Part 3, pp. 3 and 13), these lacustrine sediments have not yet been tested for liquefaction potential. This is a serious omis-sion that must be corrected before the Draft EIS is issued.

Although the lacustrine layer has not yet been tested, Teifke (1993, Part 3, p. 13) indicates that tests on a fluvial-alluvial sediment layer found on the North Plateau of the Western New York Nuclear Service Center show it to be somewhat suscepti-ble to liquefaction. Teifke indicates that the probability of liquefaction for this layer in a magnitude 5.25 earthquake ranges from about 1% to 30%, depending on the assessment procedure.

More information on liquefaction potential, while too long to cover here, is given in an appendix at the end of this paper. Points that should be noted here are that at least one sediment layer is somewhat susceptible to seismically-induced liquefac-tion, and that results are not yet available for the lacustrine sediments that lie beneath a large portion of the site.

Conclusions

Various types of evidence suggest that unstable sediments exist in and around the Western New York Nuclear Service Center. In Erie County, several miles north of the nuclear site along the same valley, a large area of unstable sediments has been identified and characterized (Owens et al. 1977). Closer to the site in Cattaraugus County, problems of quicksand have been reported in maintaining a railroad (Pietrak 1979), drilling a gas well (DEC 1975), and building a school (Gerwitz 1993; Zimmerman 1993). Within and immediately adjacent to the site, mass movement of soft clayey sediments can be observed in many places. Tests have shown that one soil layer on the site has up to 30% probability of liquefaction in an earthquake, and results are not yet available for a crucial layer of lacustrine sediments that lies beneath the site.

No single piece of this evidence is conclusive, but the quantity and variety of evidence suggests a real need for thorough examination. The individual pieces of this puzzle must be examined in a coordinated way. The evidence is too extensive to be ignored or dismissed in any competent characterization of the site.

In the area of groundwater hydrology, this report has reviewed the pieces of another puzzle that has never been put together by geologists working for the West Valley Demonstration Project. The puzzle in this case is a complex aquifer system. Almost the entire area of the Western New York Nuclear Service Center is either underlain by, or serves as a recharge area for, this aquifer system. The aquifer system almost certainly extends into Erie County and appears to be hydraulically connected to the Springville aquifers. As before, the evidence is not fully conclusive but requires immediate attention, particularly to reach the level of understanding needed to develop an overall water budget for the aquifer system. As with the problem of unstable sediments, the hydrology of this aquifer system cannot be ignored; it is a necessary part of site characterization.

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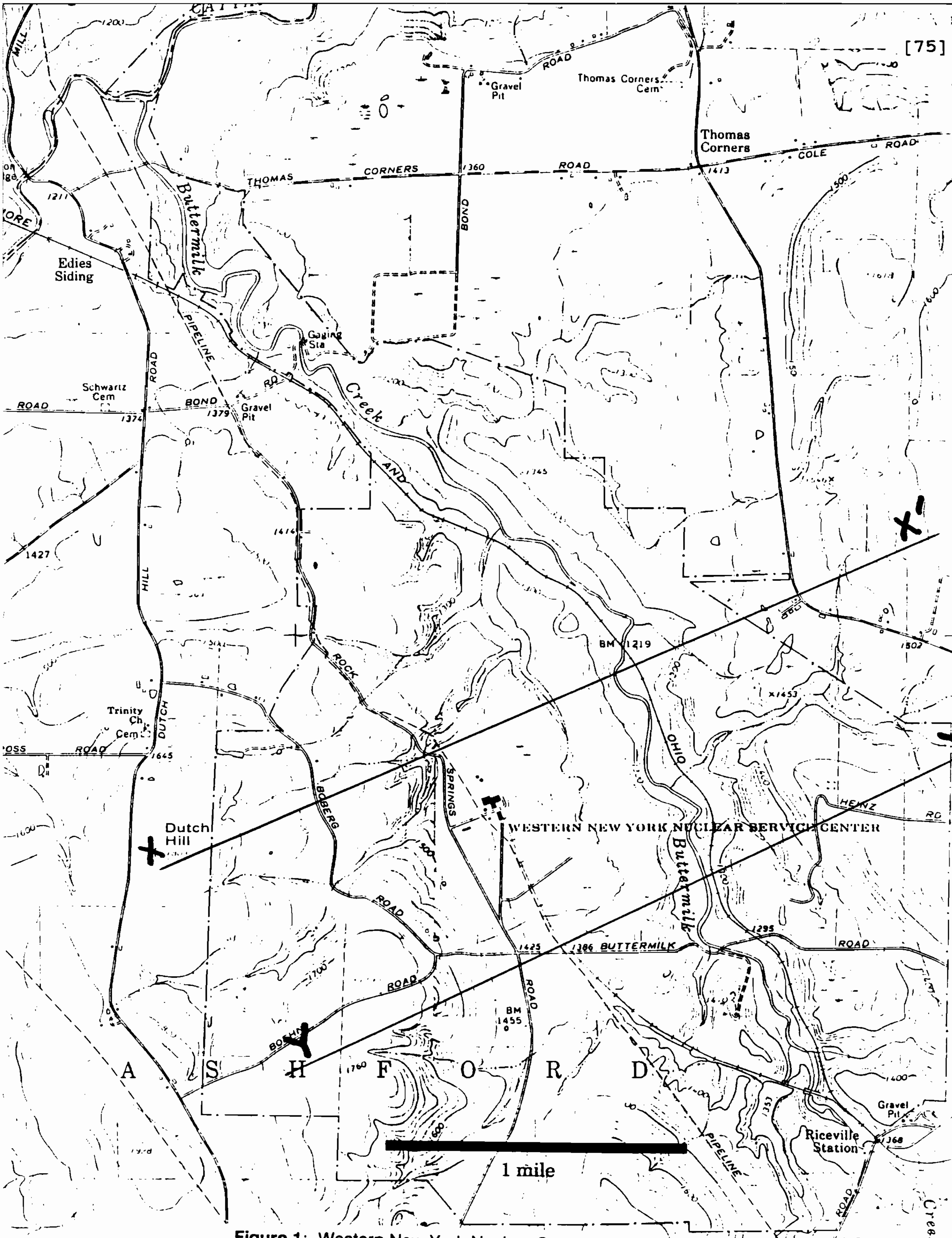

Figure 1: Western New York Nuclear Service Center

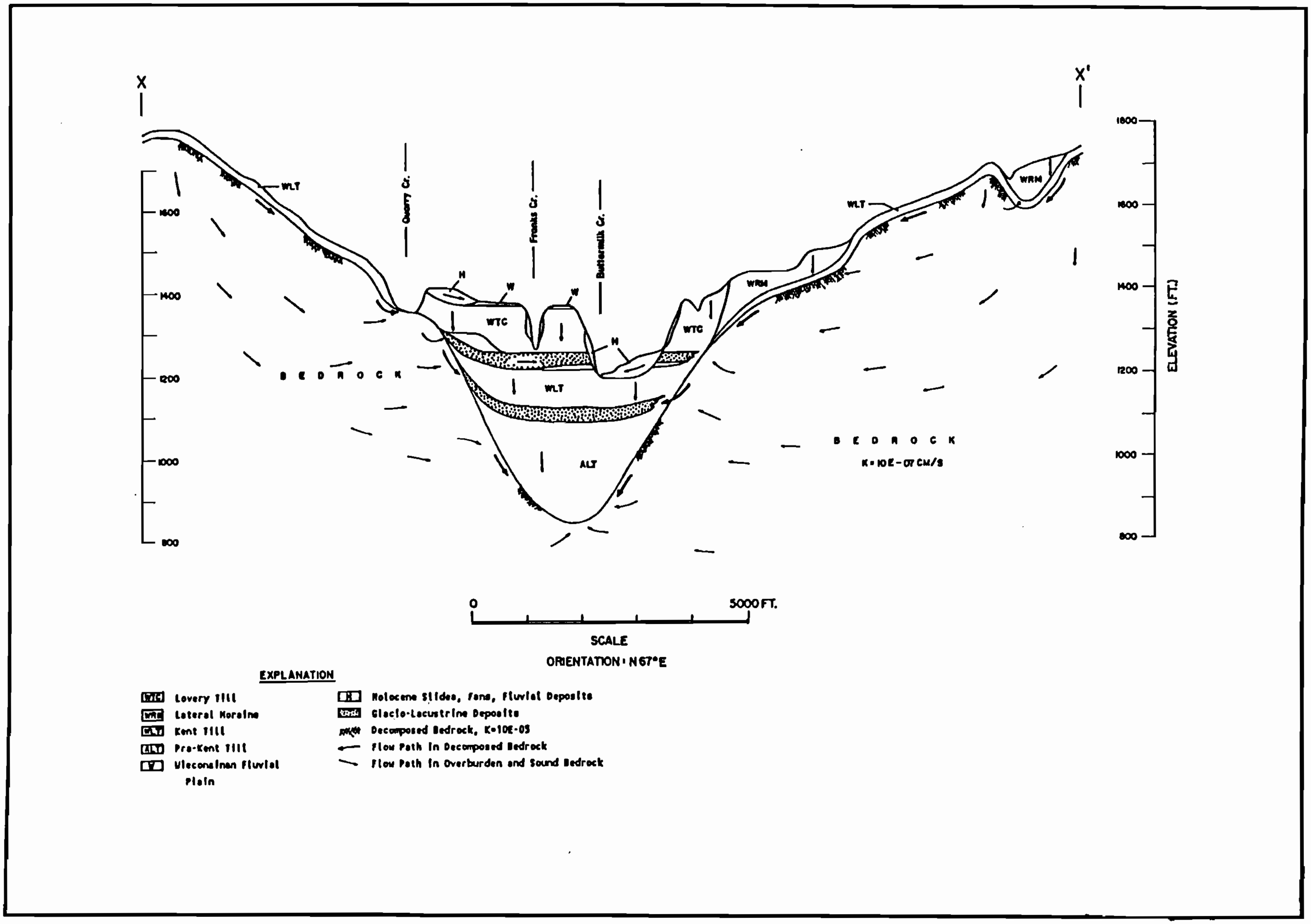

Figure 2: Cross-section **X-X'**, including groundwater flow paths, from Teifke (1993), Part 2, Fig. 2-15.

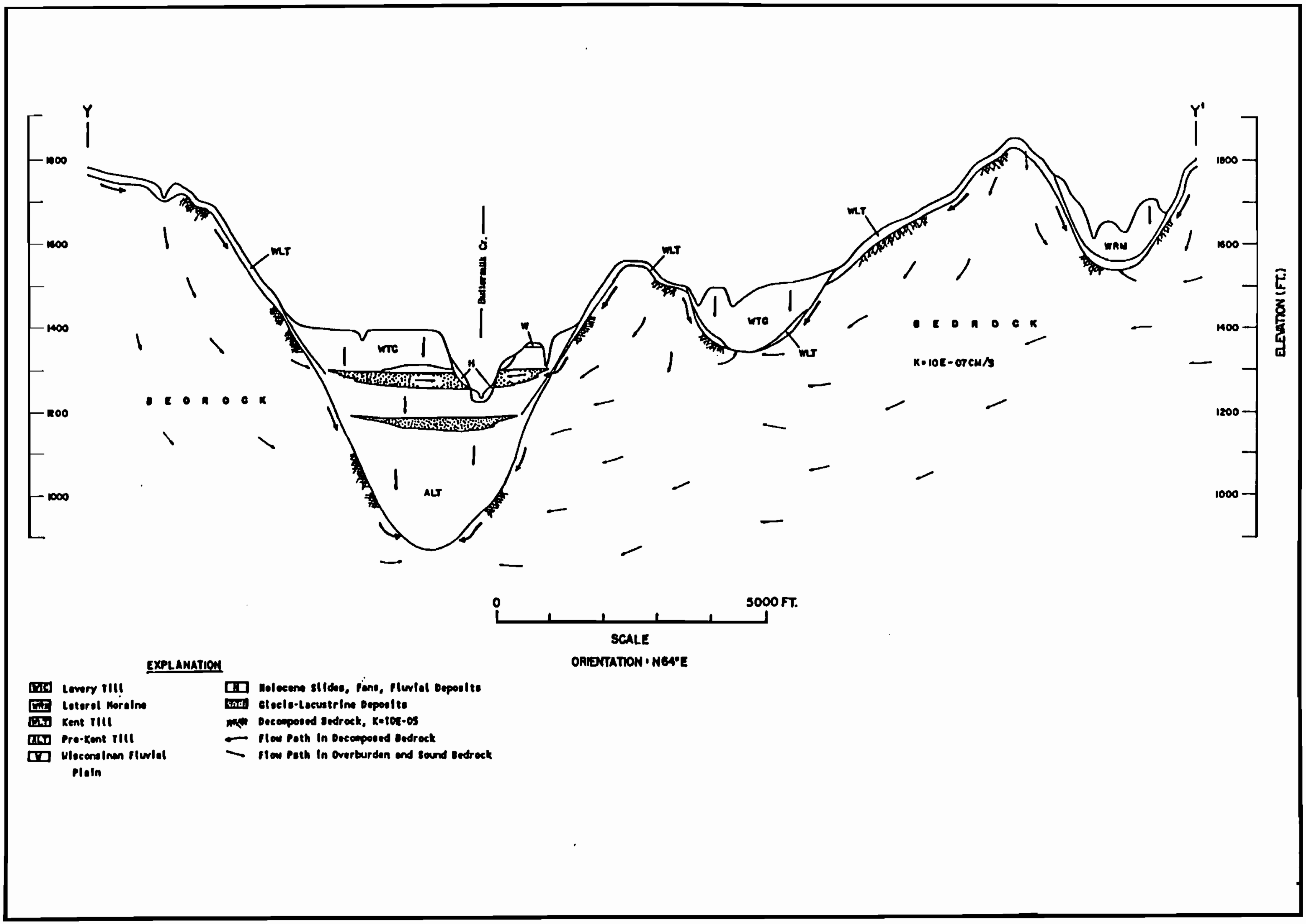

Figure 3: Cross-section **Y-Y'**, including groundwater flow paths, from Teifke (1993), Part 2, Fig. 2-16.

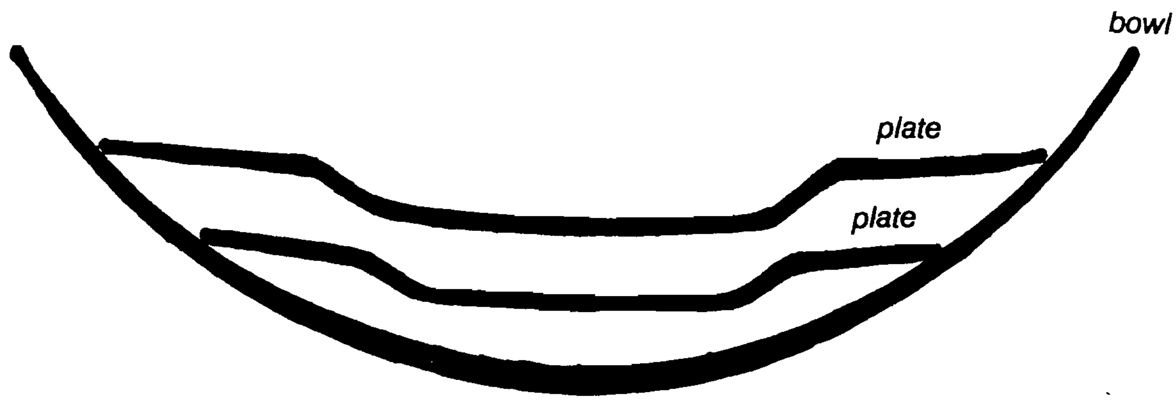

Figure 4

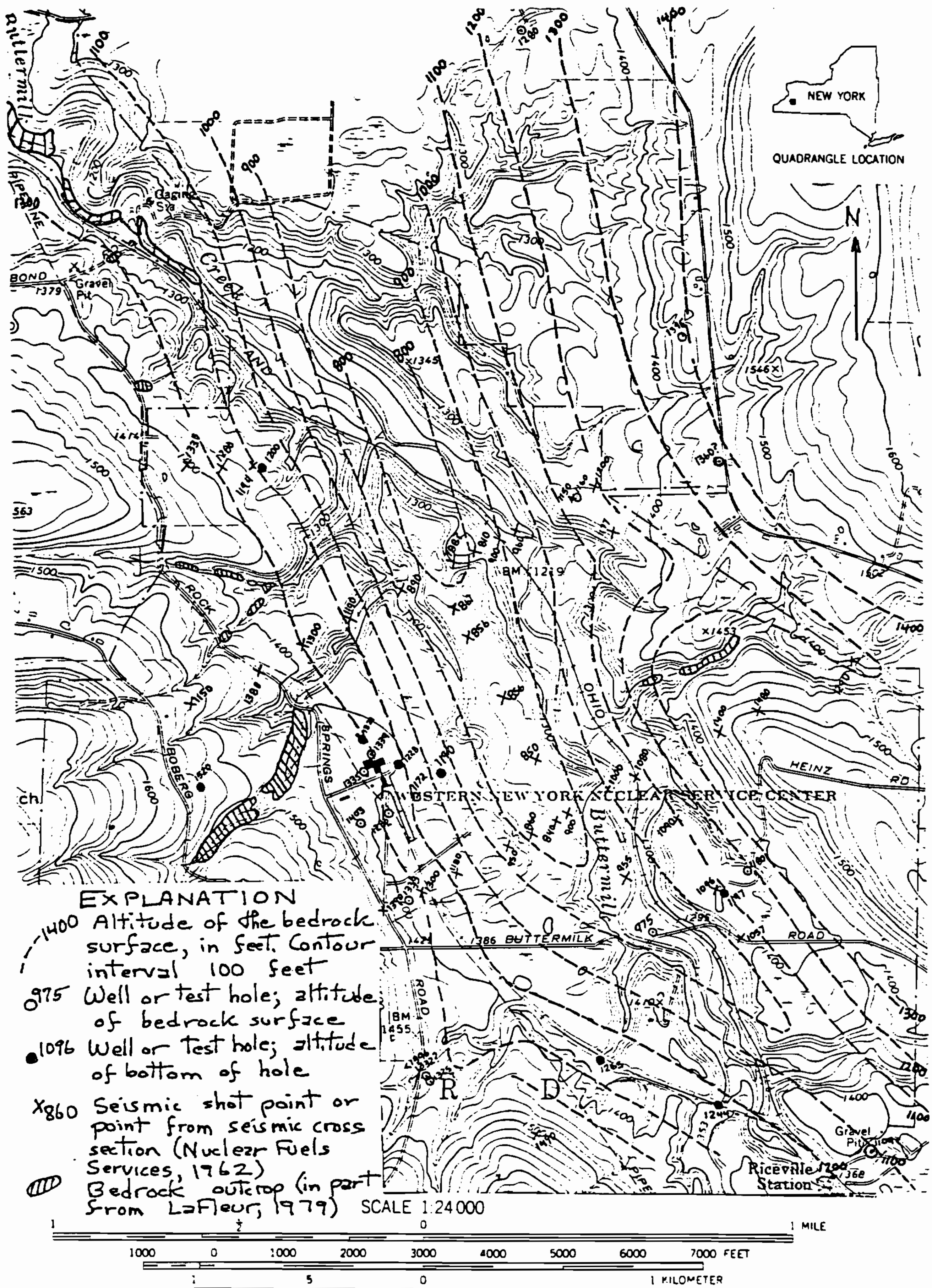

Figure 5: Contours on top of bedrock, Buttermilk Creek valley, from Randall (1980), Fig. 5.

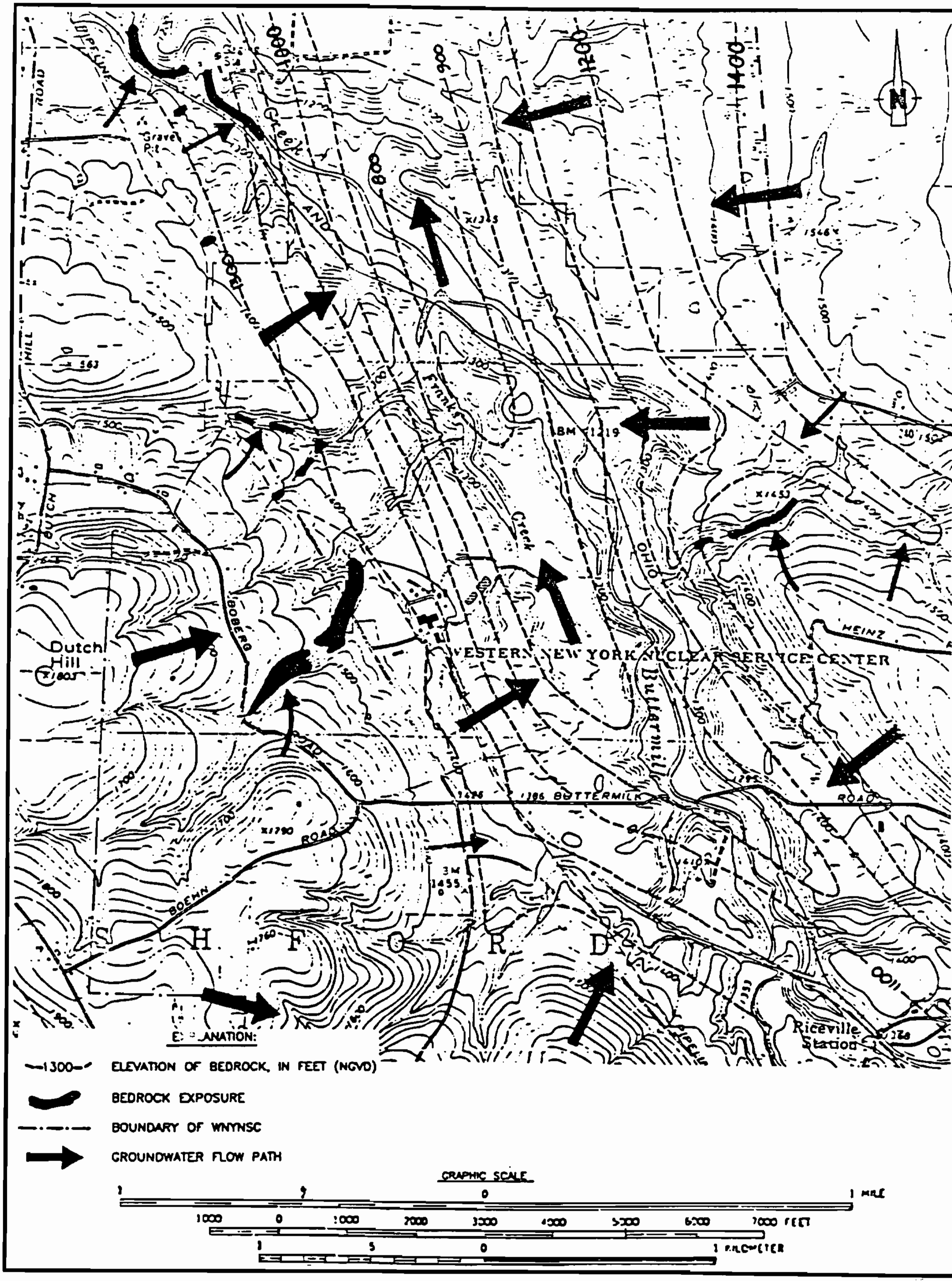

Figure 6: Groundwater flow pattern in shallowest bedrock, from Teifke (1993), Part 1, Fig. 6-3.

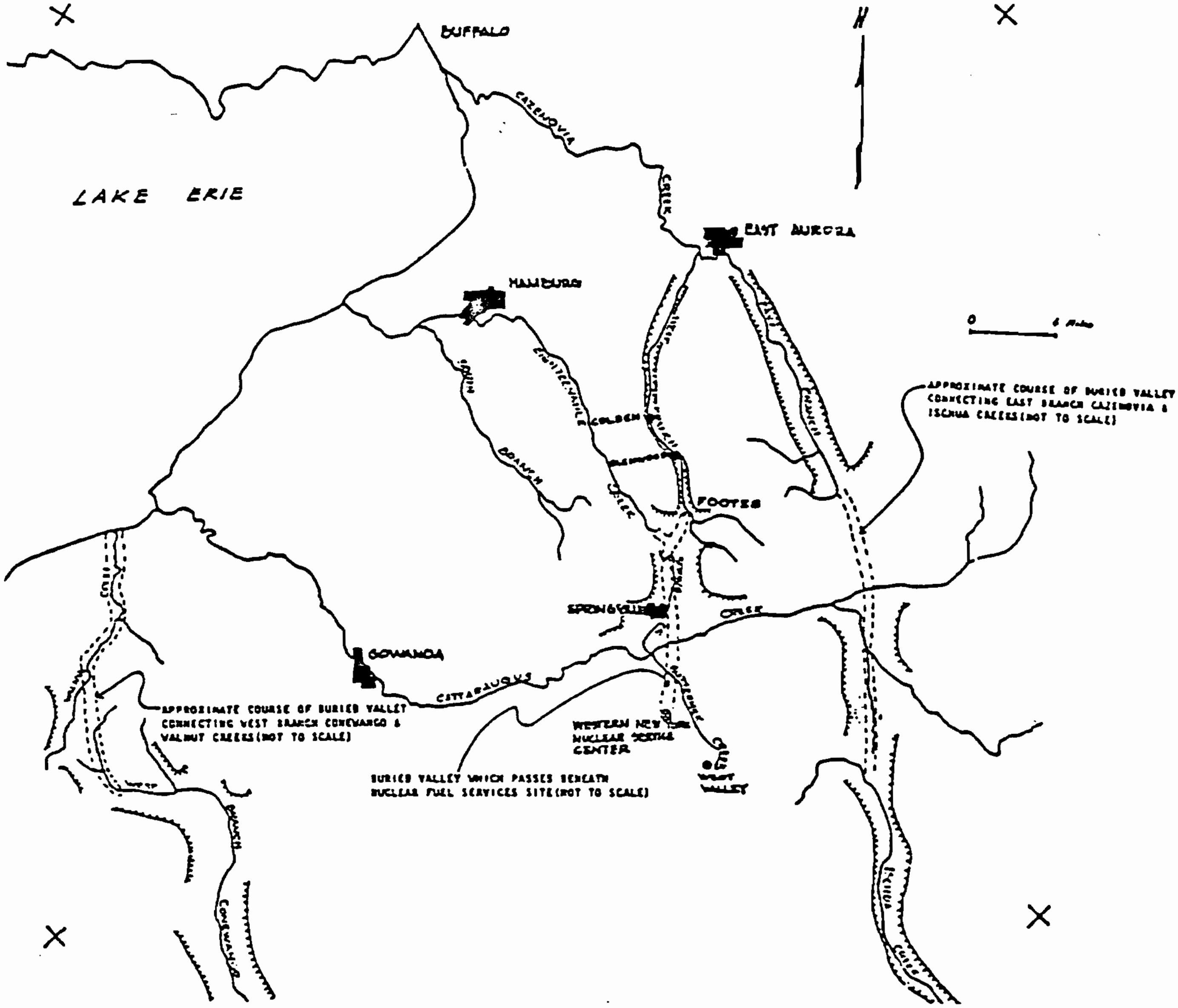

Figure 7: Interconnections among buried bedrock valleys, from Teifke (1993), Part 2, Fig. 2-2.

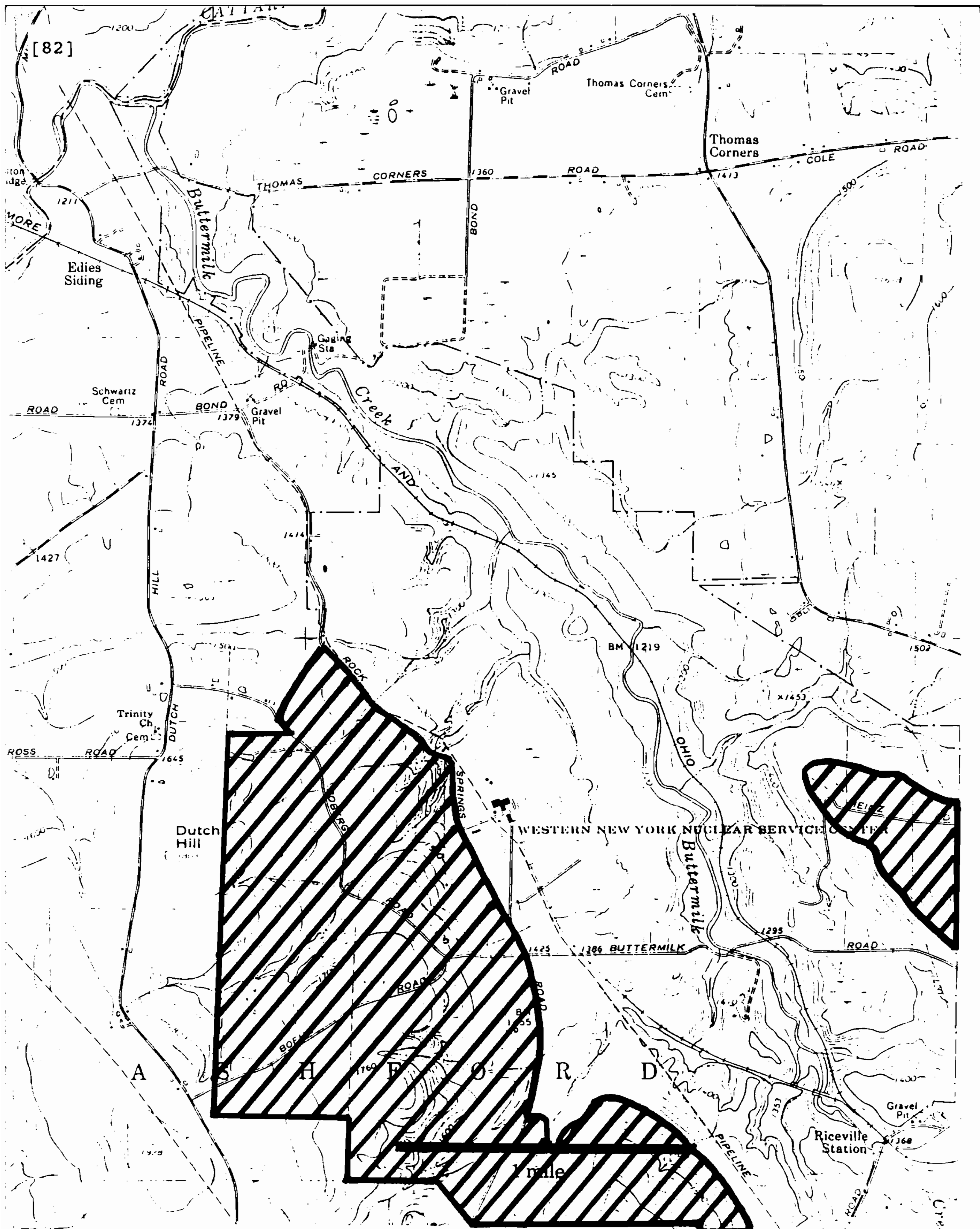

Figure 8: Approximate boundaries of deep aquifer recharge areas within the Western New York Nuclear Service Center

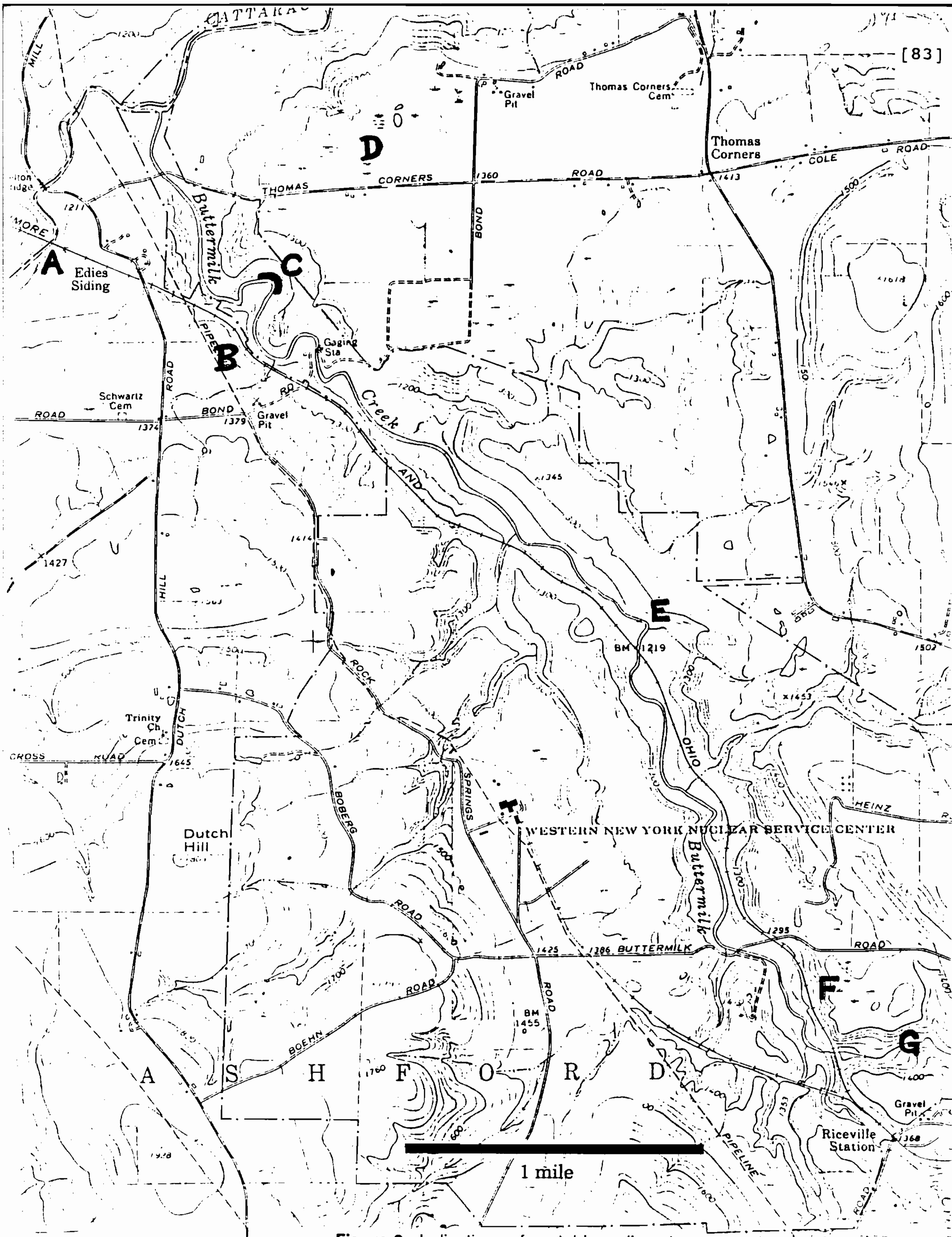

Figure 9: Indications of unstable sediments.

APPENDIX I: Liquefaction potential of sediment layers at the Western New York Nuclear Service Center

Liquefaction potential was one of the issues raised five years ago in scoping comments for the EIS that is now being prepared for site closure. Of the 23 scoping comments in Vaughan (1989), for example, four dealt with liquefaction and/or seismically-induced slumping. Those four comments were as follows:

> B2. In any alternative that leaves the main plant building standing, impacts associated with seismic events must be considered. Murray et al. (1977) provide a good starting-point. Incidentally, note their comment on p. 6 that the foundation (consisting of approx. 500 piles) under the building's massive cell structure "was repaired due to the inadequate bearing capacity of the surface layers of the soil at the site." Note also that the belief expressed by Dong and Ma (1978), p. 14, that "liquefaction should not occur" during seismic events, is not well-supported for the various layers underlying the building; it should be reassessed with current techniques such as those outlined in Liao et al. (1988).
>
> B5. In any alternative that leaves tanks 8D1 through 8D4 in situ (and presumably filled with cement), impacts associated with seismic events must be considered. Previous studies by Davito et al. (1978) and Liaw et al. (1979) assumed that tanks 8D2 and 8D4 contained HLLW rather than relatively heavier cement. Also, the belief expressed by Davito et al. (1978), p. 17, that "liquefaction should not occur" during seismic events is not well-supported for the various layers underlying the tanks and should be reassessed with current techniques such as those outlined in Liao et al. (1988).
>
> B9. As mentioned above, the possibility of soil liquefaction during severe seismic events needs to be evaluated in the immediate vicinity of the tanks and the main plant building. More generally, the possibility and potential impacts of soil liquefaction during a severe seismic event need to be considered for the whole area in and around the Project Premises, including the burial grounds. It appears, in other words, that the topographic relief is sufficient that <u>flow slides</u> or <u>mass movement</u> involving many acres of land might occur <u>if</u> soil liquefaction occurred beneath the area in question. In the vicinity of the burial grounds, for example, the geologic cross-section is shown in Nicholson and Hurt (1985), Fig. 4; the thick layer of Lavery Till would probably not be subject to large-scale liquefaction, but any occurrence of liquefaction in the underlying lacustrine layer of silt and sand could also have serious consequences, possibly involving a downslope movement of

large blocks (or the whole intact mass?) of the overlying Lavery Till that contains the burial grounds. This is not to say that liquefaction would occur during severe seismic events, but the possibility needs to be evaluated by means of current techniques such as those outlined in Liao et al. (1988).

B10. Regardless of whether large-scale liquefaction occurred, it is likely that other geomorphic processes would occur during a severe seismic event. For example, a severe seismic event would probably trigger slumping or mass wasting in areas where slumping or mass wasting is already known to occur. The scale and impacts of such processes need to be evaluated. Given the known area of slumping or mass wasting in the ravine of Buttermilk Creek northeast of the burial grounds, it is necessary to address the question of whether the ravine might be substantially blocked by earthquake-induced slumping, causing Buttermilk Creek to seek a new path. On a smaller scale, known areas of slumping or mass wasting immediately adjacent to the burial grounds would probably be affected by a severe seismic event; the impacts must be evaluated. Any erosion-control structures erected along Buttermilk Creek and its tributaries would probably be damaged or impaired in function by a severe seismic event; the impacts should be evaluated.

The Geology Environmental Information Document prepared by Teifke (1993) is expected to provide much of the geologic documentation for the EIS that is now being prepared. The section on soil liquefaction by Teifke (1993, Part 3, pp. 13-14) is as follows:

The geologic substrate at the WVDP [West Valley Demonstration Project] comprises, in order of decreasing stratigraphic position: 1) a fluvial-alluvial complex consisting of gravels, muddy gravels, and muddy sandy gravels; 2) dense and compact clayey glacial till; and 3) a sequence of thinly bedded and fairly well sorted glaciolacustrine sands, silts, and clays.

The fluvial-alluvial sequence underlies the north plateau and the main plant process building and related structures. The fluvial-alluvial sequence would be more likely to experience liquefaction than the other on-site units, owing to its age, depth of burial, mode of emplacement, textural properties, and hydrogeometry.

Dames & Moore recently completed an analysis of the fluvial-alluvial sequence to determine the potential for seismically induced liquefaction. Twenty-eight sample locations were characterized according to methods developed by Seed et al. (1983) and Liao et al. (1988). In these methods, the N-values generated when driving

> the sampling tubes into the soil are compared to empirically derived field performance curves. When evaluated by the criteria of Seed et al., only one of the twenty-eight samples was shown to have a moderate probability of liquefaction during expected levels of peak ground acceleration. According to the rankings of Liao et al., seven samples would have a moderate probability (10% to 30%) of liquefaction, and the remaining twenty-one samples would have a low probability (<10%). The Dames & Moore study results indicate that the probability of liquefaction for the north plateau substrate, under conditions associated with a a magnitude 5.25 earthquake, ranges from less than 1% to approximately 30%.
>
> The clayey (Lavery) till underlies the entire WVDP, including the north plateau, where it occurs beneath the fluvial-alluvial sequence. The cohesive texture, density, depth of burial, and the N-values, which are influenced by these properties, indicate that this unit is not susceptible to liquefaction. <u>The liquefaction potential of the glaciolacustrine sequence has not been evaluated.</u> The sensitivity of this sequence to seismically induced acceleration is expected to be relatively minor, as the depth at which this unit is found diminishes the potential for liquefaction.
>
> On the north plateau, most of the main plant complex is founded at or slightly below grade. The fuel storage pool (FSP), cask unloading pool (CUP), and several other process cells are founded below grade at elevations as low as 1,368 feet NGVD. All of these structures are founded on groups of steel H-piles driven several tens of feet into the underlying Lavery till and do not bear on the alluvial materials. They therefore would not be affected by liquefaction of the alluvium. The vault containing HLW tanks 8D-1 and 8D-2 is founded in the Lavery till at 1,362 feet NGVD, about 30 feet below the base of the alluvium. Thus, the integrity of tanks 8D-1 and 8D-2 would not be affected by liquefaction of the alluvium. On the south plateau, the host formation for waste disposal is the Lavery till. (emphasis added)

Two points should be noted in Teifke's analysis of liquefaction potential. First, the fluvial-alluvial sequence may be subject to liquefaction under seismic acceleration. Ossman (1992, pp. 52-64) provides details of the tests on which this conclusion is based. However, at the end of the above-quoted analysis, Teifke concludes that liquefaction of the fluvial-alluvial sequence would not affect waste-management facilities on the north plateau because their foundations are emplaced in the deeper till. At face value this conclusion seems valid, but, in view of the potential consequences, it seems risky to rely on a geologist's offhand opinion rather than have an engineer take a closer look at the problem. Two points raised by Vaughan (1989, paragraphs B2 and B5) should be noted, particularly the reference to the fact that the original foundation of the main plant building was "repaired" and the comment about the weight of cement in tanks 8D1 through 8D4. It should also be noted that Teifke, in the concluding paragraph quoted above, shows his unfamiliarity with the high-level waste tanks 8D1 and 8D2. He refers to a single vault (there are two, one for each tank) and he fails to note various facts such as the "flotation incident" and ad hoc corrective measures that took place when the vaults and tanks were first installed. (See Davito et al. 1978, including appendices.) These facts may or may not affect the analysis of liquefaction potential, but they should at least be known to the analyst.

The other point that should be noted in Teifke's analysis of liquefaction potential is that the glaciolacustrine unit beneath the Lavery Till has <u>not</u> been tested. This is a serious omission, particularly in view of the above-quoted scoping comment (Vaughan 1989, paragraph B9) which specifically noted a need for such testing. Teifke's comment about the depth of the unit is unconvincing, especially given the configuration of the glaciolacustrine unit relative to the valleys of Franks and Buttermilk Creeks. See Figure 2.

Liquefaction potential of the (glacio)lacustrine unit that lies beneath the Lavery till must be determined by testing. A simple assumption that its sensitivity to seismic events is "relatively minor" is inadequate. Vague generalities about the effect of depth on liquefaction potential are also inadequate. If there are relevant factors such as presence and pressure of groundwater in the lacustrine unit, these should be measured in a defensible manner and presented explicitly and quantitatively as factors that affect liquefaction potential.

Liquefaction potential must also be addressed for deeper lacustrine units in the buried bedrock valley beneath the Western New York Nuclear Service Center.

In summary, Teifke's analysis of liquefaction potential of sediment layers at the Western New York Nuclear Service Center is not adequate. A more complete and adequate analysis is needed for the EIS.

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Appendix G

Vaughan, R.C. (2005). "Fault Relationships and Basement Structure, Cattaraugus Creek Watershed, Western New York State," Thesis Proposal #2, presented to Department of Geology, State University of New York at Buffalo.

THESIS PROPOSAL #2

Fault Relationships and Basement Structure, Cattaraugus Creek Watershed, Western New York State

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January 30, 2005

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Approved: ________________________________Dr. Robert D. Jacobi, committee member

Approved: ________________________________Dr. Gary S. Solar, committee member

Abstract

Seismic lines will be run to investigate two known but poorly characterized basement faults in the Cattaraugus Creek watershed in western New York State. Soil gas lines may also be run. The work will determine the master-abutting relationship between these faults and investigate their relationship to other structural and drainage features. One of these faults strikes ENE and, by analogy to the Rome Trough, may be linked to Iapetan rifting if found to be the master fault. The other fault, which apparently strikes NNE, appears to be related to the Attica Splay of the Clarendon-Linden Fault. Determination of whether it is the master or abutting fault will aid in the interpretation of the Attica Splay, which remains a poorly understood member of the Clarendon-Linden system of faults. Characterization of these faults will be useful not only in the context of the tectonic history of the Grenville basement but also in assessing the future of radioactive wastes located within the Cattaraugus Creek drainage basin at the Western New York Nuclear Service Center near West Valley, NY.

Introduction

Reactivation of Precambrian faults and sutures through overlying sediments may reveal details of the tectonic history of Eastern North America, especially in intracratonic regions such as the Appalachian Plateau. Jacobi (2002) has used remote sensing, seismic, and outcrop data to map a number of basement faults and suspected faults in New York State. The drainage basin of Cattaraugus Creek in western New York State provides additional opportunities to investigate fault systems that appear to be expressions of older basement faults. The major drainageway of this creek – the 500-foot-deep Zoar Valley gorge near Gowanda, NY, that cuts anomalously ENE across the prevailing N-S glacial valleys of upstate New York – may itself be structurally controlled. If so, its excavation ca. 12,000 years B.P. by glacial meltwater would have consisted mainly of erosional removal of fractured blocks along a reactivated feature rather than grain-by-grain downcutting through competent rock.

Beneath the sedimentary cover of the Cattaraugus Creek watershed, the Grenville basement has not been extensively studied but contains old sutures and faults associated with the Elzevir-Frontenac Boundary Zone and the Attica Splay of the Clarendon-Linden Fault system. These features typically strike NNE (Fakundiny et al. 1978, Forsyth et al. 1994). Other basement faulting in the vicinity of the Cattaraugus Creek watershed may strike ENE and be linked to Iapetan rifting, as Jacobi (2002) has proposed for the suspected Mayville fault that approaches the Cattaraugus Creek watershed from the west.

To date, seismic investigation within the Cattaraugus Creek basin has been done for gas and oil exploration and for hazard investigation associated with the West Valley nuclear waste site that lies within the watershed. The work proposed in this dissertation is based on two existing seismic lines in the basin that were not addressed by Jacobi (2002). These two existing seismic lines have shown two faults, one apparently striking NNE, the other striking ENE, that can be projected to a point of intersection immediately south of the village of Sardinia, NY. The point of intersection, located near the Erie-Cattaraugus county line, is approximately 2 km south of the 2001 Vibroseis line on which the first of these faults was identified and approximately 8 km west of the westernmost of several dynamite lines shot in 1984-1985 to identify the ENE-trending fault. The work proposed here will run additional seismic lines near the point of intersection (near the confluence of Cattaraugus and Elton Creeks) in order to determine which of these two faults is the master fault that continues past the point of intersection. In addition, the work proposed here will use seismic and/or other methods to investigate the relationship of these two faults to additional structural features that have been observed in and near the Cattaraugus Creek drainage basin. One such additional feature is the apparent northeastward continuation of the suspected Mayville fault, which shows intermittent surface expression in outcrop north of the Cattaraugus Creek watershed.

Proposed identification of the master fault and related work

Neither of these two faults known from seismic data has been widely recognized, but both can be tentatively related to other structural features and lineaments. Identification of the throughgoing master fault will help demonstrate or disprove these relationships and thus contribute to the tectonic and glacial-drainage history of western New York State. The work may also be of interest due to the proximity of both faults and their point of intersection to the Western New York Nuclear Service Center which is located between Springville and West Valley, NY.

The first of these faults, apparently trending NNE, consists of two steeply W-dipping subparallel fault surfaces located at shot points 740 and 766 on line WVN-1, immediately east of Sardinia, NY. Both fault surfaces extend upward from basement through the Trenton and Onondaga Fms. and possibly to the base of the alluvium, with down-on-the west displacements of 8 m and 87 m, respectively, at the Onondaga Fm. (Bay Geophysical 2001.) As shown in Figure 1, this pair of faults near Sardinia is on-trend with the so-called Attica Splay of the Clarendon Linden Fault. The SSW-trending Attica Splay was studied and characterized in the 1970s by two Vibroseis lines (Fakundiny and Pomeroy 2002), the nearest of which was about 30 km NNE of Sardinia. Fakundiny (pers. comm., 2003) believes that it is likely that the pair of Sardinia faults is part of this same fault, i.e., the southwestward continuation of the Attica Splay. There is some evidence that the Attica Splay lies along a major suture within the Grenville (Culotta et al. 1990, Forsyth et al. 1994, Ouassaa and Forsyth 2002), in which case it may be reasonable to expect that the Attica Splay/Sardinia Fault is the master fault that continues past the point of intersection. The work proposed here will address this question.

If Attica Splay/Sardinia Fault is the master fault, then other faults that abut it are not likely to continue past the point of intersection. This would imply that the ENE-trending fault mentioned above does not continue westward past its point of intersection with the Attica Splay/Sardinia

Fault. However, there is some evidence to the contrary, as discussed in the next paragraph. For the purpose of discussion, this ENE-trending fault will be called the Arcade Center Fault.

The Arcade Center Fault, as mapped by Hodge (n.d.) and shown here as Figures 2-3, consists of two subparallel ENE-striking fault surfaces that extend through the Trenton Fm. and continue upward at least through the Onondaga Fm. Overall offset is on the order of 60 m, down on the south. This pair of faults was identified by a series of seismic lines shot in 1984-1985 near Arcade Center, a hamlet in the town of Arcade, NY (see Hodge n.d., Hodge and Eckert 1985, Stead 1979, and supporting files in UB Geophysics lab). This Arcade Center Fault can be projected about 8 km westsouthwestward to the aforementioned point of intersection with the Attica Splay/Sardinia Fault. If it were projected further westsouthwestward past the point of intersection (in other words, if the Arcade Center Fault were the master fault), it would then continue along the somewhat meandering but essentially linear course of Cattaraugus Creek toward Gowanda, NY. En route to Gowanda it would pass through and provide a structural-control explanation for the deep (150 m) bedrock gorge known as Zoar Valley and may provide an explanation for the ability of glacial meltwater to have downcut such a pathway through bedrock, and to have removed such a volume of rock, in the available time. Of equal or greater significance, the westsouthwestward projection of the Arcade Center Fault would coincide with the so-called Cattaraugus Creek Feature, a pair of faults south of Springville, NY, which have recently been interpreted from Vibroseis line BER 83-2A, run in 1983., as shown in Fig. 4. (Bay Geophysical 2001.) These two Cattaraugus Creek faults, which extend upward from basement through the Trenton Fm. and possibly through the Onondaga Fm., show up to 40 m of offset, down on the south. The sense of motion is thus consistent with the projected Arcade Center Fault. Additional features that may be structurally related are small thrust faults (~1 m offset) which are on-trend with the projected Arcade Center Fault and Cattaraugus Creek Feature and which have been observed in the Cattaraugus Creek gorge near Springville by the author and other geologists (Nevergold, Jacobi, Baker) and mapped by Guidetti (2002); ENE-striking multiplex joint(s) observed in the same gorge by the author; paraffinic residue observed in an abandoned well near the confluence of Buttermilk and Cattaraugus Creeks by Jacobi and others in 1999 (Bembia, pers. comm.); and the E-W orientation of azimuthal resistivity in studies by Mayer and Baker (2002) in the vicinity of the Cattaraugus Creek Feature.

None of these proposed correlations is conclusive, but in combination they support the hypothesis that the Arcade Center Fault continues westsouthwestward past its point of intersection with the Attica Splay/Sardinia Fault, then continues past Springville where it is seen as the Cattaraugus Creek Feature on line BER 83-2A, and provides structural control for the Zoar Valley bedrock gorge. The work proposed here will investigate whether this is true. It will identify whether the Attica Splay/Sardinia Fault or the Arcade Center Fault/Cattaraugus Creek Feature is the throughgoing master fault. Determination of the master fault will confirm or rule out the structural relationships proposed above for both faults.

Two additional features will also be investigated. One is an ENE-trending line of faults or large pop-ups seen in outcrop, apparently a continuation of the suspected Mayville Fault of Jacobi (2002), observed along strike from that fault and extending northeastward along a narrow linear zone toward Strykersville, NY (particularly near New Oregon and Colden, NY). This linear fault zone parallels not only the ENE-oriented bedrock gorge of Zoar Valley but also the more distant Rome Trough, suggesting an Iapetan rift origin for all. The other feature to be investigated is a possible NW-trending fault, inferred from well logs, that extends through Otto, NY, to the mouth of Cattaraugus Creek and possibly into Lake Erie toward Canada. The nature and possible relationship of these features will be addressed.

First and second rounds of testing

The first round of work proposed here would consists of E-W seismic lines run on *Genesee Road* between Savage Road and NY 16, *Card Road* between Vangilder and McKinstry Rds., *Block Road* between Town Line and McKinstry Rds., and *Gooseneck Road* between Fritz and McKinstry Rds., in order to characterize the Attica Splay/Sardinia Fault and its possible southwestward continuation, and also N-S seismic lines run on *NY 16* between Genesee and Eddy Rds., *Van Slyke and Town Line Roads* between Middle and Block Roads, and on *NY 240* between NY 39 and Cole Rd., in order to characterize the Arcade Center Fault and its possible westsouthwestward continuation along Cattaraugus Creek. These lines, totaling about 19 miles (30 km) in length, would be run with rented Vibroseis or other equipment and would then be interpreted to characterize the point of intersection and to determine the throughgoing master fault. Depending on the interpretation, a second round of seismic would be run to clarify uncertainties and to determine the relationships of these faults to other features listed above. Near-surface seismic may also be run with university-owned equipment to improve the characterization (strike and dip) of thrust faults observed in the Cattaraugus Creek gorge near Springville. Soil gas measurements (Jacobi and Fountain 1996) will be checked against seismic data in known fault locations and, if good correlation is shown, will be used to delineate faults beyond the extent of the seismic coverage. Stratigraphy will be obtained from well logs on file with the New York State Department of Environmental Conservation.

Investigation of the northeastward continuation of the suspected ENE-trending Mayville fault may be included in the second round, using both seismic and soil-gas lines. Such lines would be run along *New Oregon Road* immediately north of the hamlet of New Oregon (where the suspected fault is evident in outcrop), along *NY 240* between Colden and Glenwood, NY (no surface expression has been observed here, but the dogleg in Cazenovia Creek suggests past motion on this fault), and along *NY 16* between South Wales and Holland, NY.

Conclusions

Seismic and soil-gas lines will be run to characterize two basement faults in the Cattaraugus Creek watershed and to investigate their relationship to other structural and drainage features. Results will be integrated in the written dissertation to identify the master fault and discuss its origin, its likely response to past and present stress fields, its possible relationship to Iapetan rifting, and its relationship to other nearby features such as Zoar Valley (for the Arcade Center Fault/Cattaraugus Creek Feature) or the Grenville basement suture (for the Attica Splay/Sardinia Fault). Additional analysis and interpretation will be needed if neither or both of the faults continues through the point of intersection. The work will contribute to the structural understanding and tectonic history of the Grenville basement in this intracratonic region which includes the Cattaraugus Creek drainage basin and the Western New York Nuclear Service Center that lies within it.

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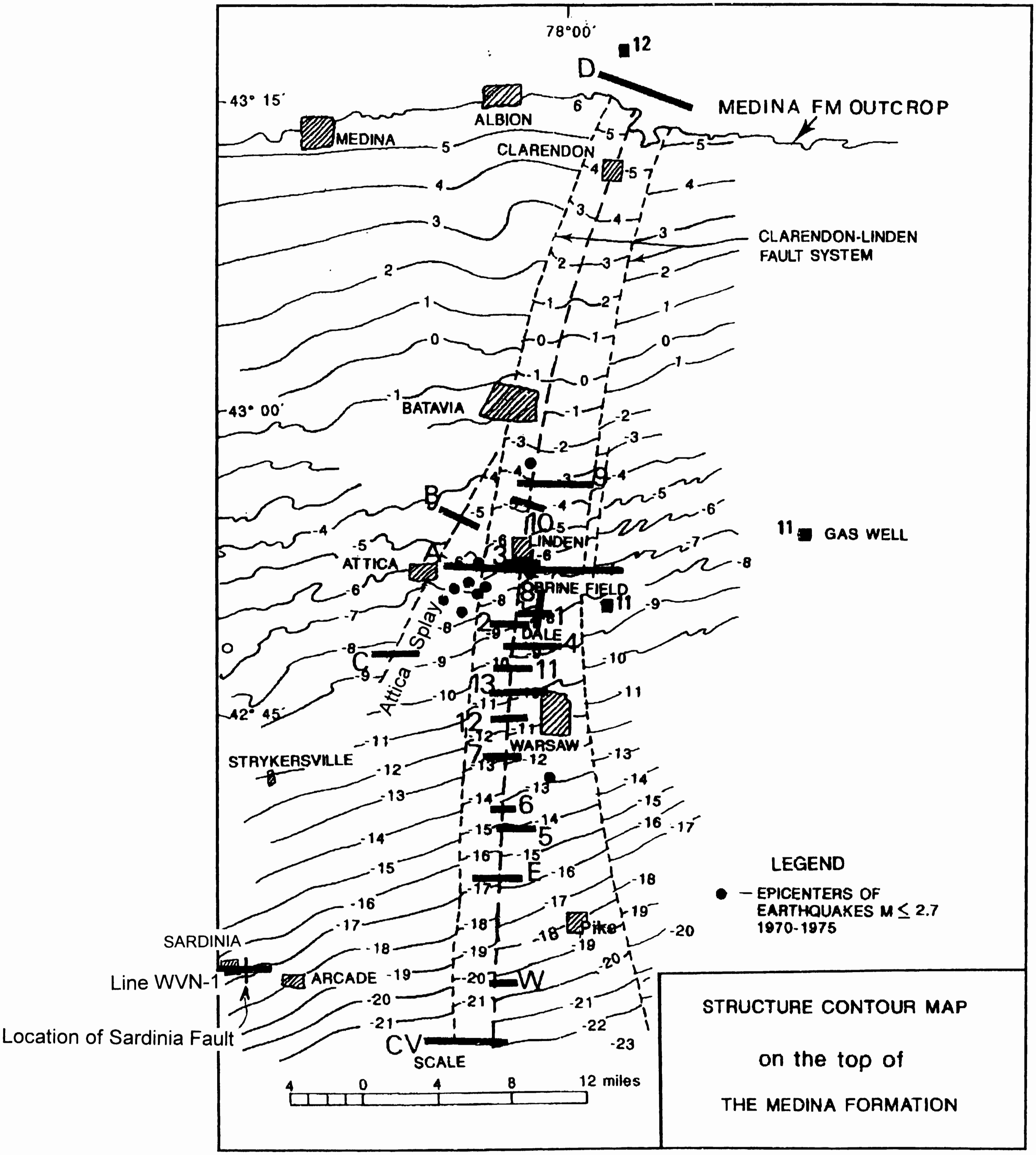

Figure 1: Spatial relationship of Sardinia Fault to Attica Splay (adapted from Fakundiny and Pomeroy 2002)

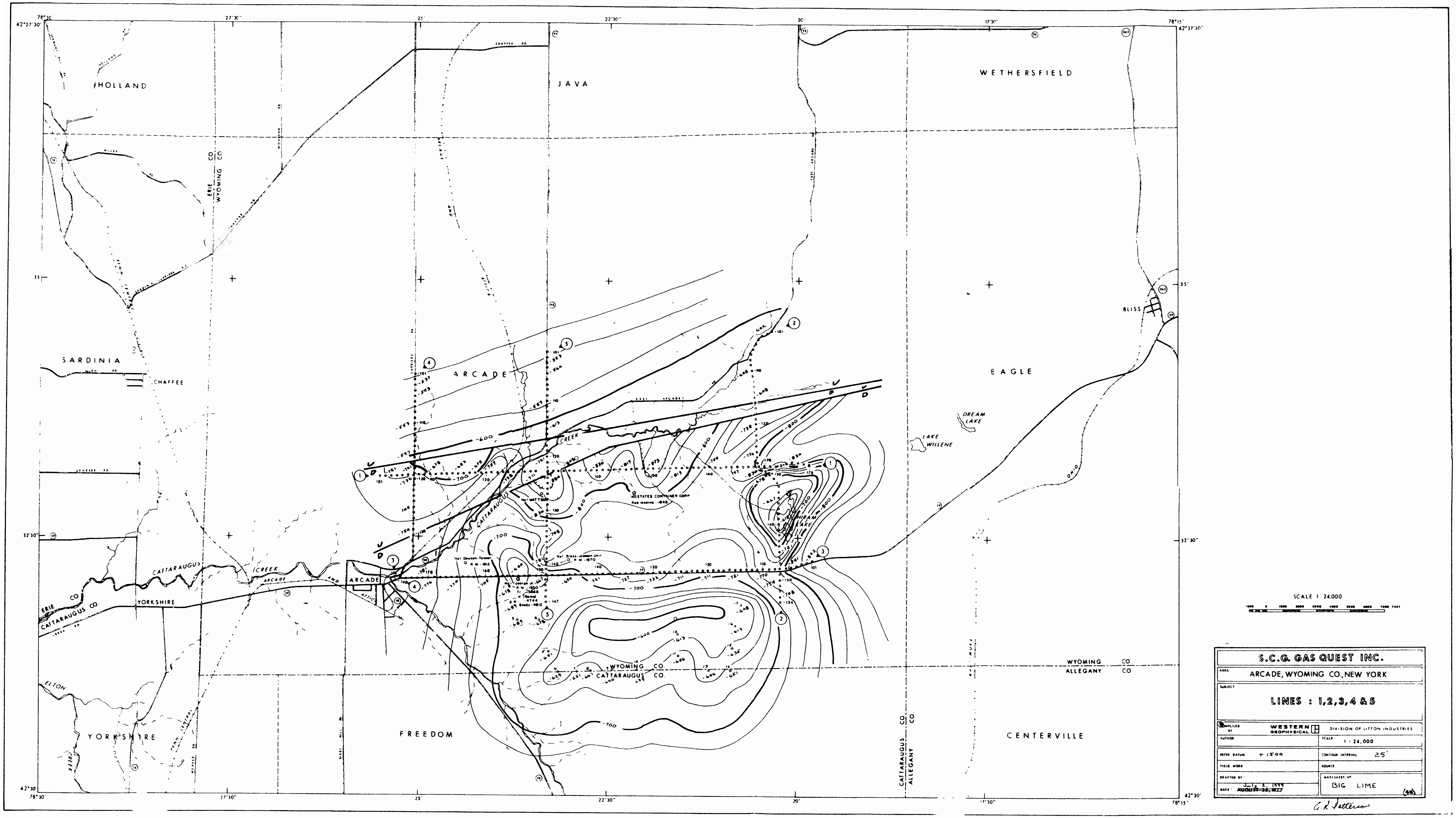

Figure 2: Structure Contours on the Onondaga Formation (from Weston Geophysical, 1979A)

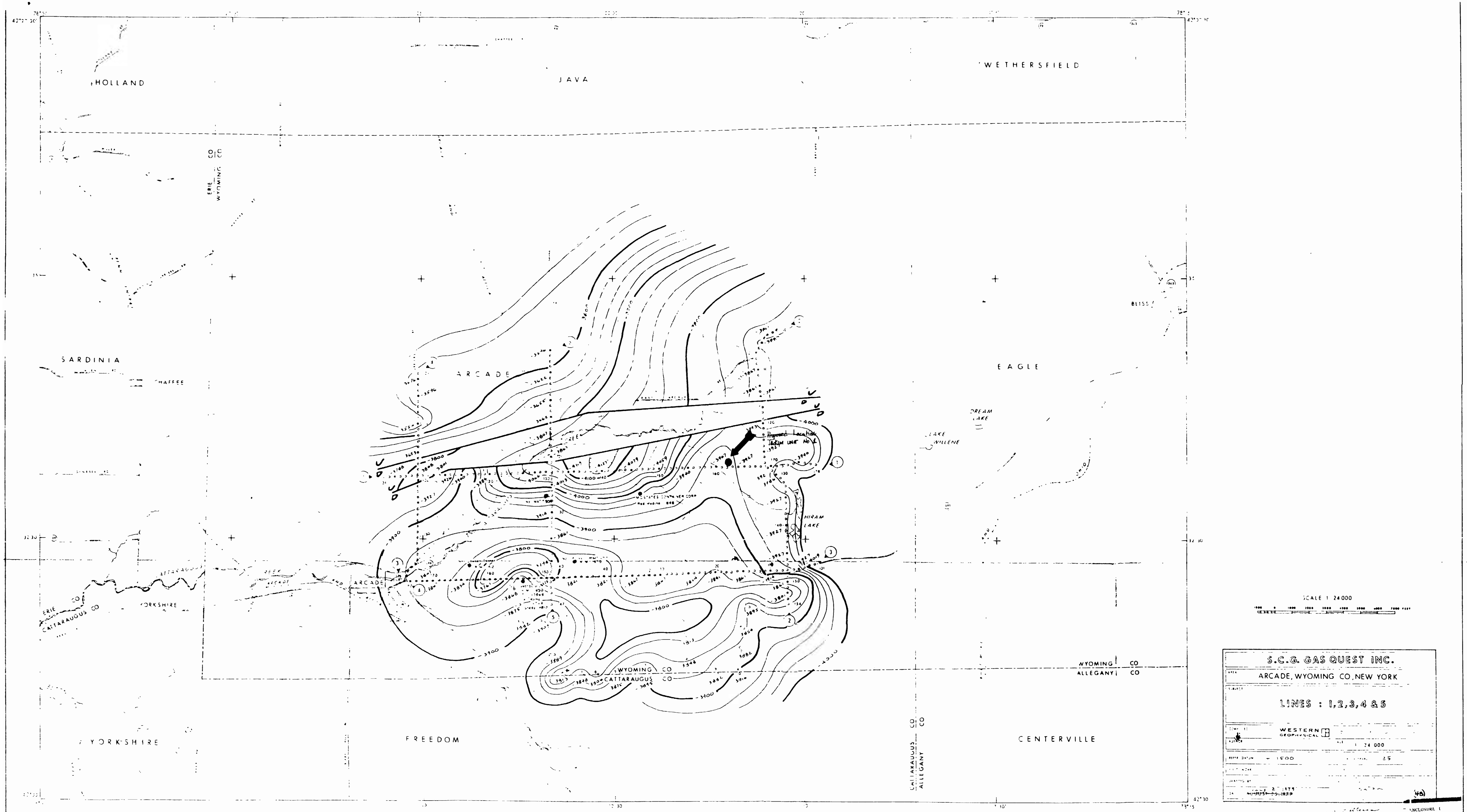

Figure 3: Structure Contours on the Trenton Formation (from Weston Geophysical, 1979B)

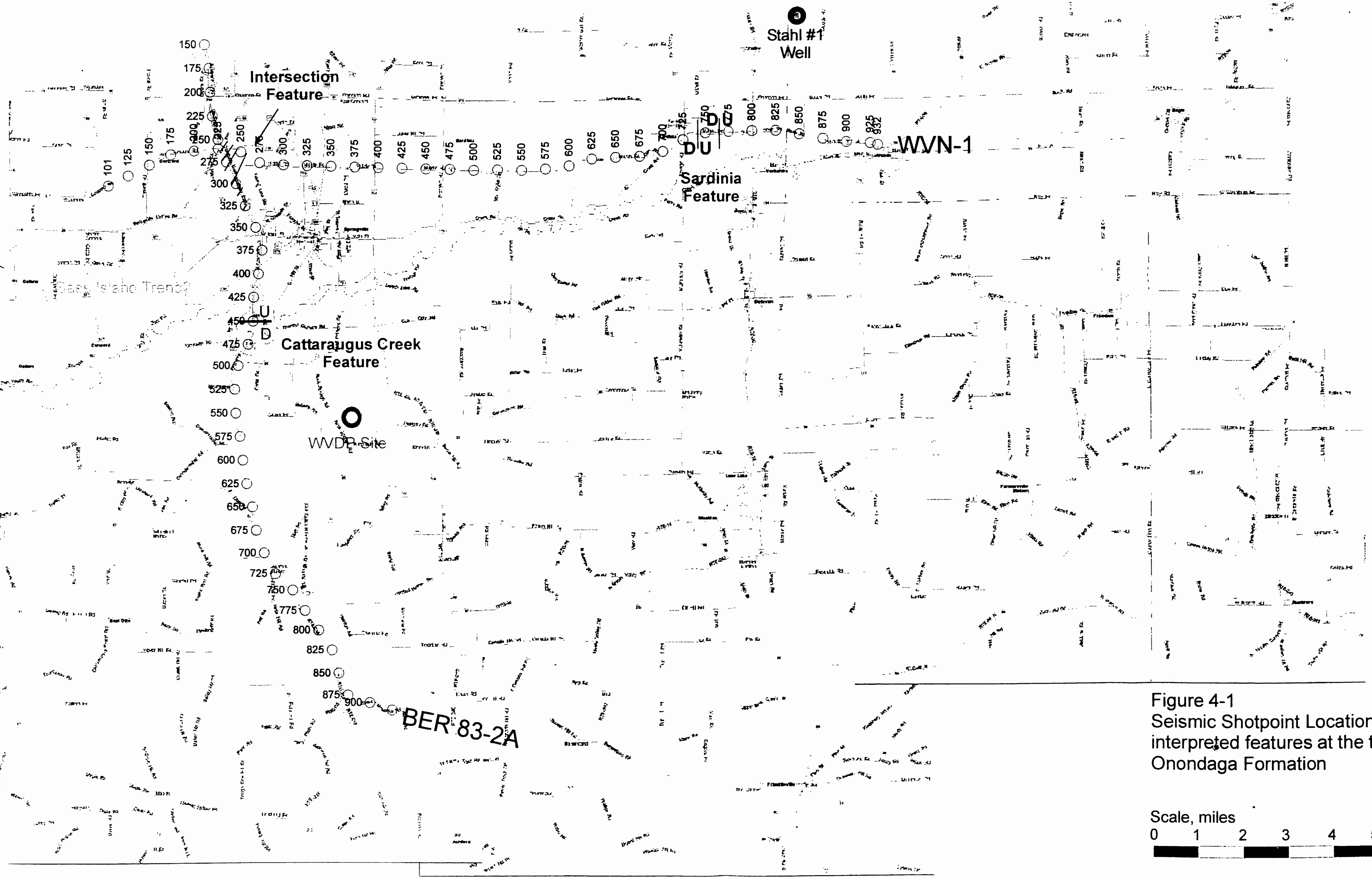

Figure 4-1
Seismic Shotpoint Location Map showing interpreted features at the top of the Onondaga Formation

Appendix H

Vaughan, R. and McGoldrick, K. (1993). “Structural Evidence for Deep, Northwest-Trending Fractures Under the Western New York Nuclear Service Center,” in *Geology Reports of the Coalition on West Valley Nuclear Wastes*, East Concord, NY: Coalition on West Valley Nuclear Wastes, 1994.

STRUCTURAL EVIDENCE FOR DEEP NORTHWEST-TRENDING FRACTURES UNDER THE WESTERN NEW YORK NUCLEAR SERVICE CENTER

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Introduction

Between 1969 and 1971, six deep injections known as shale-fracturing tests were conducted at the Western New York Nuclear Service Center. One of the tests led to disagreement on whether the liquid injected 1,010 feet below the surface had moved downward an additional 375 feet through a joint or fracture in the rock.

One of the engineers working on the shale-fracturing tests stated that "a vertical fracture had been predicted for the injection at 1,010 ft, but that it should fracture downward was a complete surprise. Presumably, the liquid injection found and followed a joint, but no joints were believed present in the shale with a vertical height of 375 ft." (de Laguna 1972/73, quoted in Sun and Mongan 1974, pp. 89-90)

Other engineers proposed alternative explanations that they found more plausible than a deep joint or fracture of 375 feet vertical height—yet no further investigation was done, and the question remains unresolved.

Based on evidence summarized here, we find it likely that the shale-fracturing test liquid did indeed encounter a deep joint or fracture of 375 feet vertical height. The evidence suggests a series of deep joints or fractures that are essentially parallel, and oriented about 315° magnetic, in the bedrock under the Western New York Nuclear Service Center near West Valley, N.Y. One of these joints or fractures is likely to have been the pathway through which the liquid in the 1971 test moved downward 375 feet.

The Structural Evidence

We find three main types of structural evidence for deep northwest-trending fractures in the vicinity of the Western New York Nuclear Service Center: **1)** Several sources suggest a northwesterly strike as a predominant direction for joints in local bedrock. **2)** Several linear topographic features that we consider structurally controlled are oriented at 315° magnetic on the local topographic quadrangle (Ashford Hollow). **3)** Major fractures of this orientation, and of considerable vertical height, have been observed in our recent field work near the Western New York Nuclear Service Center.

Specifically, the evidence in these three categories includes:

1a) Joint readings taken at the Western New York Nuclear Service Center by G. H. Chase of USGS, ca. 1969. Shown here as Figure 1(a), copied from Sun and Mongan (1974), Fig. 12.

1b) Linear topographic features observed on, and measured from, aerial photographs of the Buttermilk Creek drainage basin. Measurements by Dana et al. (1979). Shown here as Figure 1(b), copied from Dana et al. (1979), Figure 9. Dana et al. discuss (p. 41) and endorse the idea that linear features in glacial tills reflect joint patterns in the underlying bedrock.

1c) Joint readings taken by us during recent field work along Cattaraugus, Connoisarauley, and Buttermilk Creeks. Shown here as Figure 1(c).

1d) Orientations of cliff faces along Cattaraugus Creek, as recently measured and described by us. Shown here as Figure 1(d), compiled from Vaughan et al. (1993), Tables I and II. Although cliff faces are not necessarily identifiable as joints, the entire set of our cliff face orientations is included here because the downcutting of Cattaraugus Creek through the local bedrock appears structurally controlled, i.e., seems to have occurred preferentially along preexisting joints and fractures.

Despite the differences among the rose diagrams in Figures 1(a) through 1(d), all four show a <u>strong northwesterly component</u>. (Differences among the four rose diagrams can be attributed in part to a formidable sampling problem: How can a representative sample of joints be chosen? Our own observations indicate that the distribution of joint orientations varies considerably from place to place within just a few miles of the Western New York Nuclear Service Center. The observations of G. H. Chase in Fig. 1(a), for example, seem to apply strictly to outcrop along Buttermilk Creek. In any case, given such variation, and given the sporadic occurrence of outcrop, a representative sample of joint orientations is very difficult to define.)

2a) Line, or linear topographic feature(s), designated "2A" in Figure 2. (Figure 2 is a portion of the Ashford Hollow quadrangle.) This line generally coincides with the straight (and evidently structurally controlled) section of Buttermilk Creek downstream of the railroad bridge; with the tributary that enters Buttermilk Creek immediately downstream of the railroad bridge; and with the waterfall that descends into Buttermilk Creek downstream of the old Bond Road bridge. (The face of the waterfall is oriented about 315^{o} magnetic.) Belcher (1970) regards the straight section of Buttermilk Creek downstream of the railroad bridge as "a rigidly controlled channel..." The evidence for structural control of this section of the creek is, to say the least, very strong. What we are proposing here is that a single deep bedrock fracture, or a series of generally parallel fractures, follows the line designated 2A (and likewise the lines designated 2B, 2C, and 2D) in Figure 2.

2b) Line, or linear topographic feature(s), designated "2B" in Figure 2. This line coincides with a straight section of Cattaraugus Creek, located about one-half mile downstream of the Mill Street bridge; with a major fracture (see **3a**, below) that is oriented about 315° magnetic and is located along that same section of Cattaraugus Creek; with a section of Franks Creek adjacent to the SDA waste burial ground on the Western New York Nuclear Service Center; with a short, straight section of Buttermilk Creek that, at its downstream end, has experienced severe landsliding and slumping for many years; and with a small tributary of Buttermilk Creek where it is crossed by Heinz Road, immediately adjacent to the shale-fracturing test area described at the beginning of this paper. We presume the liquid injected at 1,010 feet depth in the shale-fracturing test entered the deep fracture, or one of several parallel fractures, along line 2B.

2c) Line, or linear topographic feature(s), designated "2C" in Figure 2. This line generally coincides with the waterfall that descends into Cattaraugus Creek immediately downstream of the U.S. 219 bridge (the face of the waterfall, identified as cliff A4 in Vaughan et al. 1993, is oriented about 310° magnetic) and with Buttermilk Creek at its confluence with Gooseneck Creek. The topographic evidence that this line reflects an underlying deep fracture is less compelling than the evidence for lines 2A, 2B, and 2D. Possible additional evidence is provided by the logs of two wells (74-DMB37 and 74-DMB42) that were drilled more or less along line 2C. Both wells encountered vertical and/or high-angle joints in the relatively short distance they penetrated into bedrock. The source of information for these wells (Bergeron 1985) provides no information on the strike of the joints.

2d) Line, or linear topographic feature(s), designated "2D" in Figure 2. This line coincides with two straight sections of a small tributary of Cattaraugus Creek (this tributary descends the waterfall described in the preceding paragraph) and with a straight section of Buttermilk Creek where it crosses Fox Valley Road at the southeast boundary of the Western New York Nuclear Service Center.

3a) Major fracture oriented about 315° magnetic that transects the waterfall or "slide" in cliff A2 shown in Fig. 2 of Vaughan et al. (1993). This fracture, or a parallel fracture, can also be seen in the face of cliff A2; it extends the full height of the cliff (about 30 feet).

3b) Major fracture, or series of en echelon fractures with slight offset, oriented about 315° magnetic, that extends the full height (about 100 feet) of the cliff designated C4 in Vaughan et al. (1993).

3c) Fracture oriented about 315° magnetic that transects a pop-up on the Ashford side of Cattaraugus Creek, between Springville Dam and Scoby Bridge (A6 of Vaughan et al. 1993).

3d) Major fracture, or series of en echelon fractures with slight offset, oriented about 310° magnetic, that extends the full height (about 30 feet) of a cliff on the west side of Connoisarauley Creek at its confluence with Nigh Creek.

Conclusions

Based on the various types of evidence summarized above, we find it likely that deep fractures, oriented about 315° magnetic, exist in the bedrock under and around the Western New York Nuclear Service Center. Joints and larger fractures with a strike of approximately 315° magnetic can be found over a fairly wide area; the trend appears more regional and less localized than some of the joint orientations represented in Figures 1(a) through 1(d). The size of some of the fractures observed in the field (particularly **3a** and **3b** above) and the length of the straight, apparently structurally controlled section of Buttermilk Creek (see **2a** above) provide some idea of the possible length and depth of these fractures.

Such fractures may be significant for two reasons. First, they may serve as pathways for fluid flow—either conduits for upwelling groundwater or, in the context of the nuclear site, paths for leakage of contaminants. Second, there may be very limited motion or shifting of bedrock along such fractures. (Substantial motion along northwest-trending fractures is unlikely in view of the regional stress. Whatever the origin of such fractures, they are now approximately normal to the prevailing NE-SW compressional stress. We do not rule out the development of other, more favorably oriented faults along which tectonic motion may occur—as Nur et al. (1993) propose has occurred elsewhere—possibly causing limited motion along northwest-oriented fractures.)

The evidence presented here for major northwest-trending fractures falls short of proof. We do not claim that such fractures have been shown to exist under and around the Western New York Nuclear Service Center. Proof or disproof is the next necessary step. Competent investigation designed to prove or disprove the existence of such fractures is clearly warranted by the evidence presented here. It is a necessary part of the geologic characterization of the Western New York Nuclear Service Center and must be included in the geologic characterization now underway for closure of the Center.

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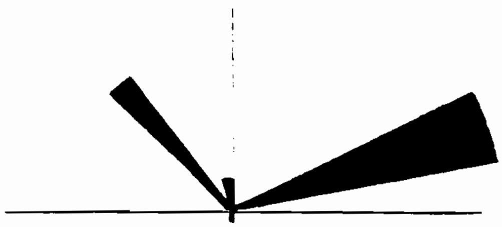

Figure 1(a): Trends of three principal joint sets reported at the West Valley site by G. H. Chase (Sun and Mongan 1974, Fig. 12).

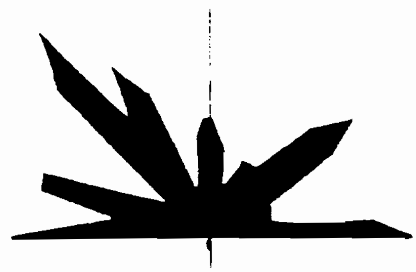

Figure 1(b): Linear topographic features observed in the drainage basin of Buttermilk Creek (Dana et. al 1979, Fig. 9).

Figure 1(c): Joints recently measured in outcrop along Cattaraugus, Connoisarauley, and Buttermilk Creeks.

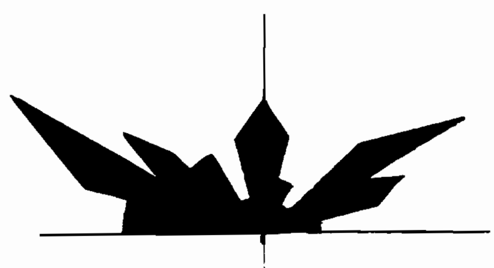

Figure 1(d): Orientations of cliff faces along Cattaraugus Creek (Vaughan et al. 1993, Tables I and II).

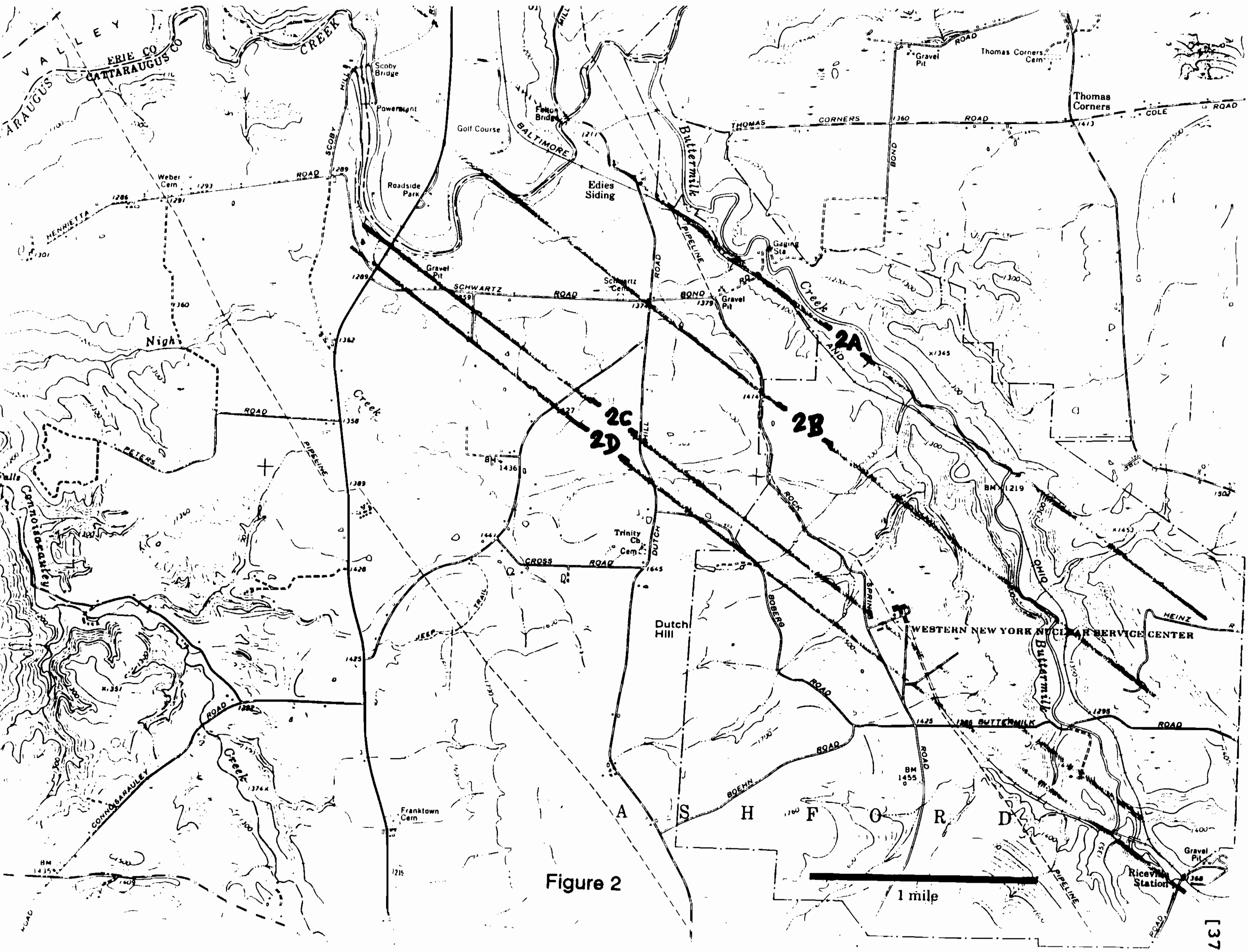
2A
2B
2C
2D
WESTERN NEW YORK NUCLEAR SERVICE CENTER
Buttermilk
Creek
Edies Siding
Thomas Corners
Thomas Corners Cem
Scoby Bridge
Powerplant
Golf Course
Roadside Park
Gaging Sta
Gravel Pit
Schwartz Cem
Trinity Ch Cem
Dutch Hill
Franktown Cem
Riceville Station
Weber Cem
ERIE CO
CATTARAUGUS CO
VALLEY
CREEK
Connoisarauley
Nigh
A S H F O R D
1 mile
THOMAS CORNERS ROAD
COLE ROAD
BOND
SCHWARTZ ROAD
BOND RD
ROCK AND SPRINGS
BOBERG ROAD
BOEHN ROAD
BUTTERMILK ROAD
CROSS ROAD
DUTCH HILL ROAD
HEINZ
OHIO
BALTIMORE
PIPELINE
SCOBY HILL
HENRIETTA
PETERS
CONNOISARAULEY
JEEP TRAIL

Figure 2

Appendix I

Vaughan, R.; McGoldrick, K.; Rauch, J.; Kent, C.; and Mathe, G. (1993). "Confirmation of Anomalous Westward Dip Between Springville and West Valley, N.Y.", in *Geology Reports of the Coalition on West Valley Nuclear Wastes*, East Concord, NY: Coalition on West Valley Nuclear Wastes, 1994.

CONFIRMATION OF ANOMALOUS WESTWARD DIP BETWEEN SPRINGVILLE AND WEST VALLEY, N.Y.

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Introduction

Strata in bedrock near Springville and West Valley, N.Y., are conventionally thought to dip southward at about 50 ft/mile. In a previous paper (Vaughan et al. 1993) we presented evidence that the dip is not southward but westward: 38 ft/mile at 265^{o} magnetic. In this paper we provide the results of two additional sets of measurements from the same area south of Springville. Although taken in a different manner, the new measurements lead to similar conclusions.

One new set of measurements indicates a dip of 35 ft/mile at 258.5^{o} magnetic; the other indicates 48 ft/mile at 254.5^{o} magnetic.

Thus, the three sets of measurements are mutually corroborative. All show a generally westward stratigraphic dip in this area between Springville and West Valley. In combination, the three sets of measurements imply a dip of roughly 40 ft/mile at about 260^{o} magnetic (252^{o} true).

An additional result of this work is the tentative identification of the Laona siltstone member in the outcrop south of Springville. Tesmer (1975) treats all outcrop along Cattaraugus Creek near Springville as part of the undifferentiated Canadaway shale formation. In Chautauqua and western Cattaraugus Counties he recognizes the Laona siltstone member as a distinct unit within the Canadaway; however, he believes that the Laona pinches out east of Gowanda. At elevations between 1205' and 1235' along Cattaraugus Creek, and at about 1280' along Buttermilk Creek, we have observed a prominent siltstone or sandstone unit that is similar in appearance to the Laona and is in about the right stratigraphic position. We consider it to be a continuation of the Laona. As such, it provides a useful marker for stratigraphic correlation and for further structural study.

Further structural study is clearly warranted. The local westward dip near Springville and West Valley must be reconciled with the southward regional dip, and also with the faults and other structural deformations that exist in the area.

Study Method

In our previous paper (Vaughan et al. 1993), we made reference to preliminary measurements on one siltstone bed visible in the outcrop that "indicated a dip of approximately 40 ft/mile to the west." The work described here is a continuation of those measurements.

Specifically, we have measured the height or elevation of two different strata. One of the strata is the aforementioned siltstone bed which is several inches thick, ranging up to 8.5 inches thickness in S_1. As a working label, we have referred to it as the "hard stratum."

About 70 feet higher in the stratigraphic column is the "big stratum," as we have termed it. This sandstone or quartz-rich siltstone occurs at or near the bottom of a group of strata that we tentatively identify as the Laona siltstone member.

In taking these measurements, the level of Cattaraugus Creek (or, in one case, Buttermilk Creek) was used as the vertical reference point. Most measurements were made diagonally upward from the creek level to the pertinent stratum. A 100-foot tape measure was employed in most cases, but, when necessary, two such tape measures were used in combination. The angle or dip of the measurement was determined by means of a sighting inclinometer.

Two of the measurements were made vertically from the rail of the U.S. 219 bridge to the pertinent stratum. Separate measurements were made from the bridge rail to the surface of the creek to ensure consistency in the vertical reference system.

Results

All measurements of the two strata are listed in Appendices I and II. Results are summarized in Tables I and II. See Figure 1, adapted from Vaughan et al. (1993), for the designations of the various cliffs on which measurements were taken. See also Table III for identification of the "slides" numbered S_1 through S_{29} in cliff A3. Several of the measurements were taken in these "slides."

Elevation of the hard stratum was measured in eight locations, as shown in Table I. Single measurements were made in three of these eight locations. Measurements were taken and then repeated at each of the other five locations. See Appendix I for details.

In cases where repeat measurements were made, the discrepancy between the first and second measurements ranged from a fraction of an inch (in S_{12} and S_{29}) to 4.3 feet (in A4). The second of two measurements at a given location is generally considered more accurate, mainly because the inclinometer was hand-held and therefore not very steady for some of the earlier measurements. Later measurements were made with the inclinometer resting on a rock or other convenient, stable surface.

Despite the reasons for preferring the later measurements, a simple average was taken in all cases where a second measurement was made.

As indicated in Table I, a least-squares fit shows the hard stratum dipping 35 ft/mile at a strike of 251^{o} true or 258.5^{o} magnetic. This agrees relatively well with the dip of 38 ft/mile at 265^{o} magnetic that was derived previously by an entirely different method (Vaughan et al. 1993).

The "big" stratum, located about 70 feet higher, was measured for elevation in six locations. Five of the measurements were single measurements, while one was repeated and averaged, as shown in Appendix II. Results are summarized in Table II. Based on a least-squares fit, the big stratum dips 48 ft/mile at 247^{o} true or 254.5^{o} magnetic. Again, this is relatively close to the prior results mentioned above.

Discussion

Although there are probably systematic as well as random errors in the measurements of the two strata, both types of error are believed to be negligible in terms of the overall results. For example, the use of creek elevations as a vertical reference point probably introduces both types of error. Creek elevations inferred from the topographic map may be in error by a few feet, but this should have little effect in view of the fact that we require our inferred elevations to change at a realistic rate along the course of the creek. Fluctuations in creek level will introduce a random error, but we have observed no more than a few inches difference in level over the season.

Aside from the vertical reference point, two possible sources of error are 1) measurement error and 2) confusion or misidentification of strata.

In spite of the difficult terrain, measurement error appears minimal as long as the inclinometer is positioned on a stable surface when readings are taken. Repeatability of measurements at the same location, as already discussed, provides some idea of the extent of measurement error.

Confusion or misidentification of strata is a potential problem. Our previous work (Vaughan et al. 1993) employed numerous dip measurements over short distances, such that a single stratum had to be traced only 45 to 50 feet at most. The current work is based on relatively few measurements of two individual strata over distances of thousands of feet. A measurement of the wrong stratum would be detrimental to the results.

The procedure of showing that two strata seen thousands of feet apart are in fact the same stratum usually falls short of proof; it relies instead on a combination of visual identification and inference. The inference is strengthened in the present instance by the partial overlap of measurement locations, i.e., the fact that some but not all of the measurement locations are common to both the hard stratum and the big stratum.

Measurements involving both strata were taken at three of the locations (A4, S_{29}, and A2). Four measurements taken at these three locations show the differences in elevation between the hard stratum and big stratum to be 73.0, 70.2, 68.4, and 69.1 feet. The average is 70.2 feet; the standard deviation is 2.0 feet. Thus, over the area studied, the big stratum remains an essentially constant distance of about 70 feet above the hard stratum. This consistent stratigraphic separation helps verify the identities of both the hard stratum and the big stratum at measurement locations A4, S_{29}, and A2.

At the various other locations, only one of the two strata was visible or accessible for measurement. The similarity of the least-squares fits for the hard stratum and big stratum shows that the results are consistent and not very dependent on the particular locations chosen for measurements. This consistency serves as a check against any serious misidentification of the strata being measured.

Faults, Pop-Ups, and Major Fractures

Structural features such as faults, pop-ups, and major fractures have been observed by us in outcrop south of Springville, both in the course of the work described here and in our previous work (Vaughan et al. 1993). These features include:

1. Single fault in A2, as indicated in Figure 1.

2. Two side-by-side faults in A2, as indicated in Figure 1.

3. Possible fault in C3, as indicated in Figure 1.

4. Pop-up in C3, seen as localized distortion in the strata several feet above creek level, a short distance downstream from the possible fault in C3.

5. Pop-up at the base of S_{26} in A3, seen as distortion in the strata a few feet above creek level, extending over a distance of about twenty horizontal feet, perhaps with an inch or more of vertical offset.

6. Pop-up at or near S_{18} in A3. The strata appear to be distorted and/or misaligned here, perhaps with a foot or more of vertical offset, in an area of the cliff that measures at least fifty vertical feet and at least a hundred horizontal feet. The steepness and deep incision of the cliff at S_{18} make it difficult to take either diagnostic measurements or diagnostic photographs.

7. Pop-up at the base of S_2 in A3, seen as a slightly open joint (strike about 265^{o} magnetic) with localized distortion in the adjoining strata.

8. Pop-up at the base of S_1 in A3, seen as a slightly open

joint (strike about 275° magnetic) with localized distortion in the adjoining strata.

9. Pop-up in A6, located a few feet above creek level and configured as a symmetrical gable-type pop-up (see Fakundiny et al. 1978, Fig. 12). At least two fractures (oriented about 300° and 315° magnetic) have cut through the pop-up and apparently act, or have acted, as stress relief.

10. Pop-up in C6, located a few feet above creek level and directly across the creek from (i.e., directly east of) the A6 pop-up described above. The exposed portion of the pop-up in C6 is substantially eroded away but appears, from what remains, to have been continuous with the pop-up in A6 before Cattaraugus Creek cut down through it. In other words, the original pop-up apparently had an essentially east-west axis. Minor excavation on the property of the Springville Electric Department would probably expose intact portions of this pop-up in C6.

11. Two or three major fractures, spaced about 20' apart and oriented about 315° magnetic, that appear as joints in the face of cliff A2 and transect the slide/waterfall at the upstream end of A2 near the single fault. We consider these fractures to be part of structural feature 2B proposed by Vaughan and McGoldrick (1993).

12. Major fracture, oriented about 315° magnetic, that extends the full height of cliff C4. This fracture is continuous but exhibits a few inches of horizontal (en echelon) offset; it also has about an inch of vertical offset, down-on-the-southwest.

13. Major fracture or cleft, oriented about 255° magnetic, near the top of cliff A3 above S_{13}. This fracture has been observed only from the base of the cliff. It is not clear whether it could be reached from the top for closer inspection.

14. Multiple, closely spaced joints or fracture planes that occur at, and essentially parallel to, the faces of the upper and lower falls in A4. These joints may be related to structural feature 2C proposed by Vaughan and McGoldrick (1993).

15. Multiple, closely spaced joints or fracture planes that occur at, and essentially parallel to, the face of the lower waterfall that descends to Buttermilk Creek a few hundred feet north of Bond Road. These joints may be related to structural feature 2A proposed by Vaughan and McGoldrick (1993).

Photos of the above features are available from the authors upon request.

A possibility that cannot be ruled out is that additional structural features such as small faults may lie between the current measurement locations. For example, there is an apparent discrepancy between the second and third measurements listed in Table II. These measurements were taken only 300' apart. To fit the dip implied by other measurements along the same east-west line, the big stratum should be several feet lower at the falls in A4 and/or several feet higher under the U.S. 219 bridge at A3.5. It is not yet clear whether the discrepancy is due to measurement error, confusion of strata, or a small fault with several feet of offset, down-on-the-east.

Conclusions and Directions for Further Study

Measurements on the two strata described here have confirmed the anomalous westward dip reported in our earlier paper (Vaughan et al. 1993). As noted at the end of that paper, further work is needed in several areas. Such work is needed for proper geologic characterization of the Western New York Nuclear Service Center, especially in the context of the joint federal-state EIS now being prepared for Phase II of the West Valley Demonstration Project and for site closure.

In general, the federal and state agencies involved in the EIS must gather and use geologic data that is real and defensible. Using data that is merely presumed to be true, such as the presumed southward stratigraphic dip that is shown here to be false, is not acceptable. The geologic setting of any nuclear waste site is closely related to its long-term performance and must therefore be well-understood.

In addition to the specific recommendations for further study made in our previous paper (Vaughan et al. 1993), we recommend that one or both of the strata described in this paper be used as stratigraphic markers beyond the geographic area studied here. Assuming that these strata do not pinch out, they should be available as reference points for drilling conducted in areas south and west of Springville, including large portions of the Western New York Nuclear Service Center.

One deep core sample has already been taken on the Western New York Nuclear Service Center in conjunction with the 1969-1971 shale-fracturing tests; the core, if still extant, should be checked for the strata described here. If the core is no longer available, a more detailed log than that given by Bergeron (1985, well 69-USGS1-5) should be sought so that the strata may be identified and used in mapping the subsurface structure.

Other borings into bedrock, both inside and outside the Western New York Nuclear Service Center, should also be conducted as part of the structural geology work needed for a defensible site characterization. Here again, one or both strata should be used if possible as stratigraphic markers.

West of Springville, the two strata in question can probably be used as subsurface markers in drillings done north or south of Cattaraugus Creek; however, along the course of the creek (i.e., through Zoar Valley), the two strata may appear in outcrop and may serve to elucidate the structure and stratigraphy. It is particularly important that the stratigraphic sequence immediately beneath the present valley floor be understood in terms of the downcutting rate of Cattaraugus Creek through Zoar Valley. We have observed relatively durable bedrock in the creekbed in several locations between Springville and Zoar Bridge. The durable rock in these locations no doubt acts as a knickpoint that limits downcutting of the creek. It is thus important to understand both the thickness of such durable rock and the nature of the strata that lie beneath it.

North and east of Springville, the two strata described here may be observable in outcrop on some of the hilltops. Again, the purpose of finding these strata would be (in conjunction with other techniques such as drilling and seismic reflection) to aid in mapping the subsurface structure. The structure to the north and east is clearly of interest, given the high probability that the closest approach of the Bass Island Trend is immediately north of Springville and given the fact that the Attica Splay of the Clarendon-Linden Fault approaches the Western New York Nuclear Service Center from the northeast.

In all, a great many questions remain to be answered about the structural geology under and around the Western New York Nuclear Service Center. We hope that the work described here will be useful in a small way to outline what needs to be done.

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R. Vaughan and K. McGoldrick (1993). Structural Evidence for Deep Northwest-Trending Fractures under the Western New York Nuclear Service Center Seven-page report dated September 13, 1993.

Table I: Coordinates (x,y,z) expressed in feet for **hard stratum**:

x = distance east from top of upper waterfall in A4
y = distance north from top of upper waterfall in A4
z = elevation above sea level

(65', 65', 1141.5')	A4
(550', 350', 1141.3')	C5
(700', -100', 1142.8')	S_1 in A3
(750', -100', 1141.6')	S_2 in A3
(1000', -125', 1146.4')	S_8 in A3
(1200', -150', 1148.7')	S_{12}/S_{13} in A3
(2000', 65', 1149.3')	S_{29} in A3
(3500', 1185', 1164.4')	A2

Best fit by least squares:

35 ft/mile at 251^o true or 258.5^o magnetic
z-intercept: 1139.0'

Table II: Coordinates (x,y,z) expressed in feet for **big stratum**:

(-600', 2000', 1214.3')	A5
(0, 0, 1213.1')	A4
(300', 0, 1208.5')	A3.5
(2000', 0, 1217.7')	S_{29} in A3
(3500', 1125', 1233.5')	A2
(8125', -200', 1280')	Falls above Buttermilk Creek

Best fit by least squares:

48 ft/mile at 247^o true or 254.5^o magnetic
z-intercept: 1207.5'

Table III: Guide to identification of the "slides" in cliff A3:

The "slides" are numbered, in consecutive order upstream, from the U.S. 219 bridge. Photos are available from the authors upon request.

s_1 is approx. 385' upstream from the centerline of the U.S. 219 bridge.

s_{12} is approx. 900' upstream from the centerline of the U.S. 219 bridge.

s_{16} is V-shaped.

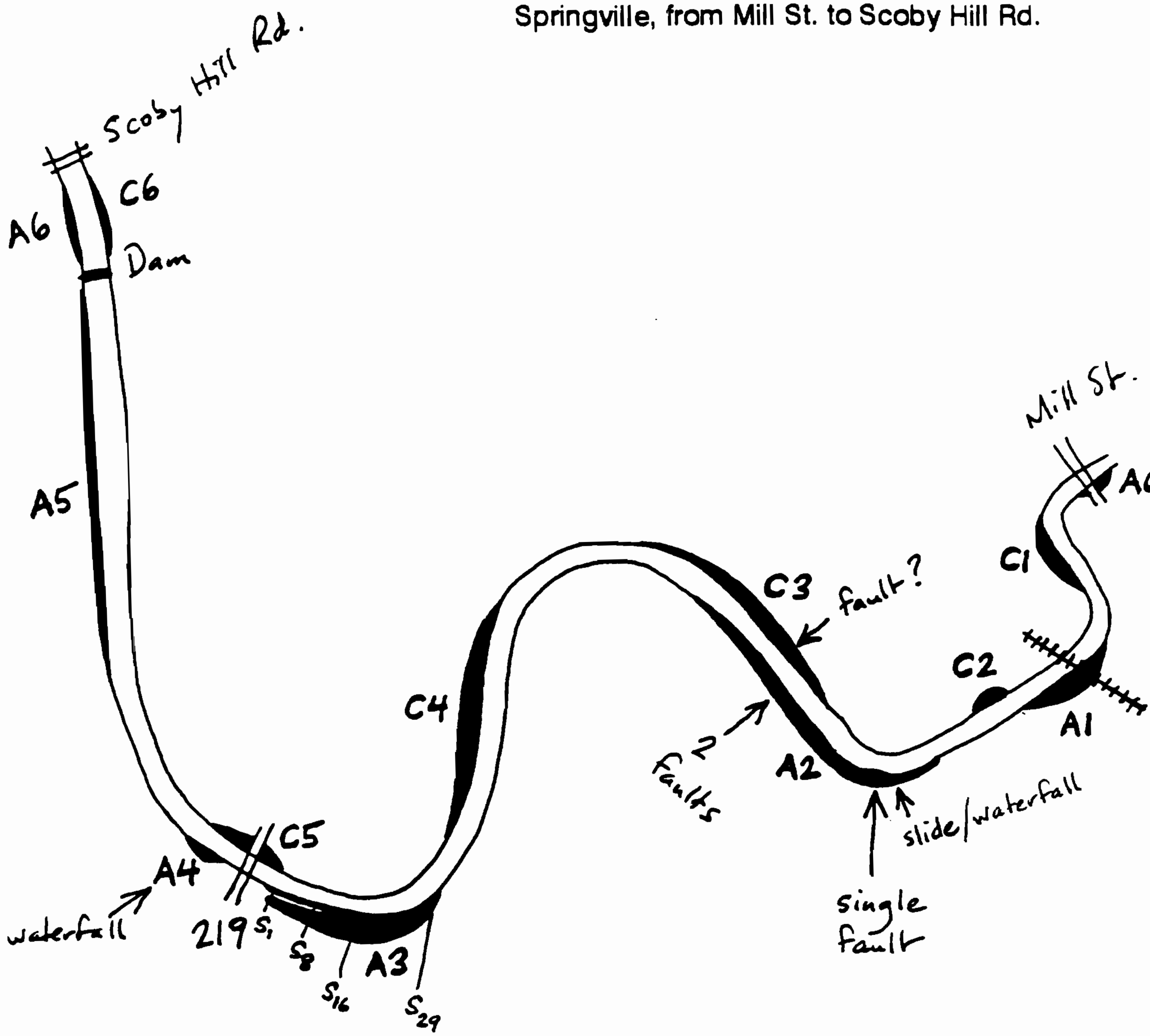
s_{18} is deeply incised, terminating in a flat rock face at its top end.

s_{21} and s_{26} are deeply incised.

s_{28} usually has water flowing down it. In upstream order, it is the last of the "slides" that descends directly to Cattaraugus Creek.

s_{29} does not descend directly to Cattaraugus Creek; it descends to a flood plain or terrace about 10' above creek level.

Figure 1: Cattaraugus Creek immediately south of Springville, from Mill St. to Scoby Hill Rd.



CLIFF DESIGNATIONS:

A0 through A6 on Ashford side

C1 through C6 on Concord side

APPENDIX I: Elevation data for **hard stratum**:

In **A4** [waterfall]:
(65', 65', 1141.5')
Creek elevation assumed from topo map to be 1089'
Two measurements: Hard stratum above creek level is
either $0.5 + 102 \sin 4^{o} + 76.3 \sin 38^{o} = 54.6'$ [7/15/93]
or $2.5 + 72.5 \sin 2^{o} + 79 \sin 35^{o} = 50.3'$ [11/13/93]
Average: 52.5'
Elevation of hard stratum above sea level is thus 1141.5'

In **C5** [under, and several feet E of, U.S. 219 bridge]:
(550', 350', 1141.3')
Bridge rail above C5 is 173.9' above creek level, as measured and calculated 10/30 and 10/31/93.
Creek elevation assumed from topo map to be 1089'
Two measurements: Hard stratum above creek level is
either $77 \sin 11^{o} + 95 \sin 28^{o} - 6 = 53.3'$ [7/15/93]
or $173.9 - 122.5 = 51.4'$ [10/30/93]
Average: 52.3'
Elevation of hard stratum above sea level is thus 1141.3'

In $\mathbf{S_1}$ in **A3** [about 385' upstream from U.S. 219 bridge]:
(700', -100', 1142.8')
Creek elevation assumed from topo map to be 1090'
One measurement: Hard stratum above creek level is
$1 + 73.2 \sin 45^{o} = 52.8'$ [7/3/93]
Elevation of hard stratum above sea level is thus 1142.8'

In $\mathbf{S_2}$ in **A3**:
(750', -100', 1141.6')
Creek elevation assumed from topo map to be 1090'
One measurement: Hard stratum above creek level is
$0.5 + 79.5 \sin 40^{o} = 51.6'$ [7/3/93]
Elevation of hard stratum above sea level is thus 1141.6'

In $\mathbf{S_8}$ in **A3**:
(1000', -125', 1146.4')
Creek elevation assumed from topo map to be 1090'
One measurement: Hard stratum above creek level is
$1 + 92 \sin 37^{o} = 56.4'$ [7/11/93]
Elevation of hard stratum above sea level is thus 1146.4'

In $\mathbf{S_{12}}$, or between S_{12} and S_{13}, in **A3**:
(1200', -150', 1148.7')
Creek elevation assumed from topo map to be 1090'
Two measurements: Hard stratum above creek level is
either $0.5 + 87 \sin 42^{o} = 58.7'$ [7/11/93]
or $1 + 98.2 \sin 36^{o} = 58.7'$ [7/15/93]
Average: 58.7'
Elevation of hard stratum above sea level is thus 1148.7'

APPENDIX I (cont'd): Elevation data for **hard stratum**:

In **S_{29}** in **A3**:
(2000', 65', 1149.3')
Creek elevation assumed from topo map to be 1091'
Two measurements: Hard stratum above creek level is
either $6 + 19 \sin 27^{o} + 64 \sin 43^{o} = 58.3'$ [9/25/93]
or $5.5 + 82.1 \sin 40^{o} = 58.3'$ [10/14/93]
Average: 58.3'
Elevation of hard stratum above sea level is thus 1149.3'

In **A2** [in slide/waterfall at upstream end of A2, near single fault]:
(3500', 1185', 1164.4')
Creek elevation assumed from topo map to be 1099'
Two measurements: Hard stratum above creek level is
either $1 + 100 \sin 39^{o} = 63.9'$ [7/3/93]
or $98 \sin 43^{o} = 66.8'$ [10/24/93]
Average: 65.4'
Elevation of hard stratum above sea level is thus 1164.4'

APPENDIX II: Elevation data for **big stratum [Laona]**:

In **A5**:
(-600', 2000', 1214.3')
Creek elevation assumed from topo map to be 1087'
One measurement: Big stratum above creek level is
194 sin 41^{o} = 127.3' [10/30/93]
Elevation of big stratum above sea level is thus 1214.3'

In **A4** [waterfall]:
(0, 0, 1213.1')
Hard stratum elevation assumed from other measurements to be 1141.5'
Two new measurements: Big stratum above hard stratum is
either 111.3 sin 41^{o} = 73.0' [10/14/93]
or 114 sin 38^{o} = 70.2' [10/24/93]
Average: 71.6'
Elevation of big stratum above sea level is thus 1213.1'

In **A3.5** [under U.S. 219 bridge]:
(300', 0, 1208.5')
Bridge rail above A3.5 is 176.0' above creek level, as measured and calculated 10/30 and 10/31/93.
Creek elevation assumed from topo map to be 1089'
One measurement: Big stratum above creek level is
176.0 - 56.5 = 119.5' [10/30/93]
Elevation of big stratum above sea level is thus 1208.5'

In **S_{29}** in **A3**:
(2000', 0, 1217.7')
Hard stratum elevation assumed from other measurements to be 1149.3'
One new measurement: Big stratum above hard stratum is
0.83 + 95.5 sin 45^{o} = 68.4' [10/14/93]
Elevation of big stratum above sea level is thus 1217.7'

In **A2** [in slide/waterfall at upstream end of A2, near single fault]:
(3500', 1125', 1233.5')
Hard stratum elevation assumed from other measurements to be 1164.4'
One new measurement: Big stratum above hard stratum is
91.5 sin 49^{o} = 69.1' [10/16/93]
Elevation of big stratum above sea level is thus 1233.5'

In **waterfalls that descend to Buttermilk Creek, N of Bond Rd.**:
(8125', -200', 1280')
Big stratum is in upper waterfall, at or sl. above railroad track.
Map and other sources imply track elevation between 1275' and 1280'
Buttermilk Creek elevation assumed from topo map to be 1145'
One measurement: Height of track above creek level is
85 sin 59^{o} + 28 sin 17^{o} + 84 sin 39^{o} = 133.9' [10/26/93]
Elevation of big stratum above sea level is thus about 1280'