DUKE GEOLOGICAL LABORATORY

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TRIPS ON THE ROCKS

Guide 12: Cameron's Line and the Hodges Complex, West Torrington, Connecticut

Trip 13: 23 September 1990; Trip 38: 27 April 1996

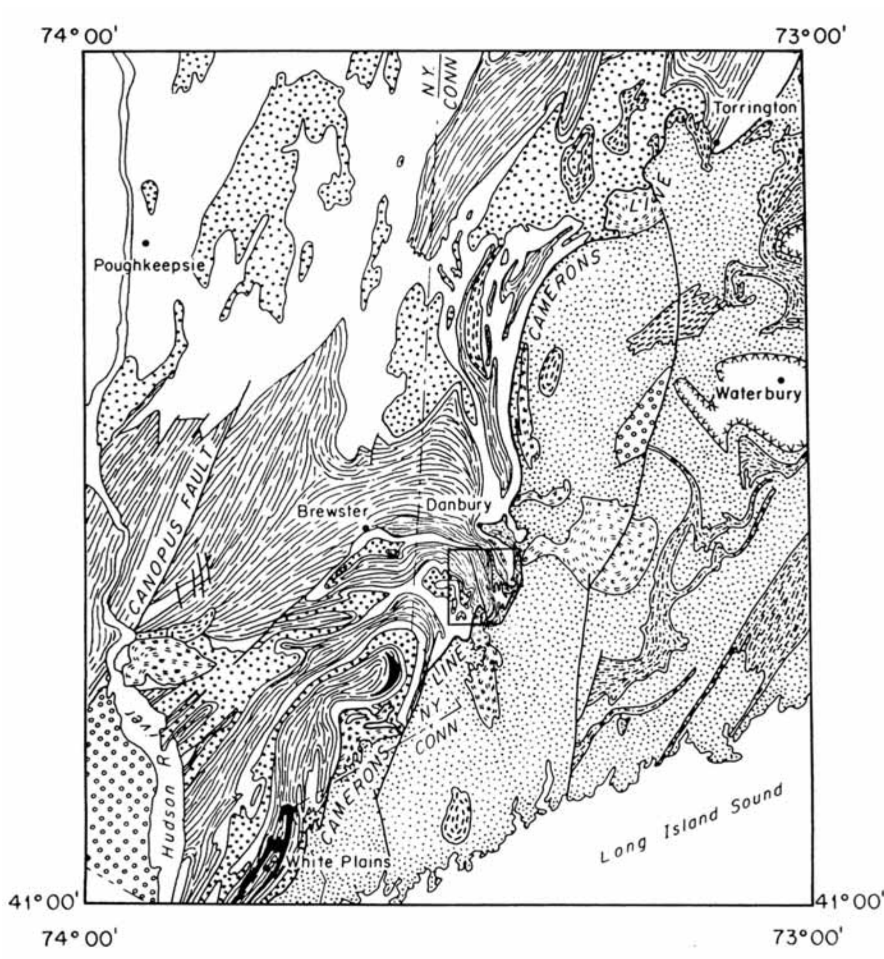

Figure 1 – Geologic sketch map of Cameron's Line and related plutons in western Connecticut. (T. R. Spinek and L. M. Hall, 1985, fig. 1, p. 220.)

Field Trip Notes by:

Charles Merguerian and John E. Sanders

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INTRODUCTION

Today's field trip to the Hodges Complex, West Torrington, Connecticut differs from most previous trips in that we plan to concentrate on the geologic relationships within a small area situated in the crystalline highlands of western Connecticut (Figure 1, on cover). Our particular focus is on the igneous rocks of the Hodges Complex and their relationships to the major tectonic feature known as Cameron's Line. The rocks that we shall examine have been subjected to numerous episodes of deformation.

We shall concentrate on demonstrating the relationships of the structural features made by these episodes of deformation to Cameron's Line and the plutons of the Hodges Complex. In the process, we shall illustrate the significant geologic information that can be obtained from an analysis of plutons and their surrounding rocks. For example, the relative date of emplacement of a pluton is determined by studying the cross-cutting relationships between pluton and the rocks it intrudes (the so-called "country rocks"). In many cases, the absolute time of crystallization of the plutonic rocks can be determined by means of radiometric dating techniques. Finally, the heat from the pluton may create notable mineralogical- and textural changes (known as contact metamorphism) in the country rocks. Other features within the pluton may be of interest, such as a sequence of intrusion of different kinds of igneous rocks; indications that cumulus-type igneous rocks formed by differential movement of crystals in the magma and the formation of layers comparable to those in sedimentary rocks; and xenoliths, pieces of the country rock that became detached and surrounded by the magma (here, the "city rock" completely surrounds the "country rock").

In recent years, much interest has centered on the kinds of mafic- and ultramafic igneous rocks that form at spreading centers on mid-oceanic ridges and then become part of the oceanic crust and -lithosphere. A complete succession of the igneous rocks formed at a spreading ridge is an ophiolite complex. When one encounters such igneous rocks, it is important to determine whether the rocks are parts of the oceanic crust that were mechanically emplaced in the solid state as parts of a great overthrust or whether these rocks formed as a pluton, in which case they would have crystallized from a fluid that may have originated in the mantle, but moved upward to their present locations as a body of magma. Beginning in 1973, CM began examining and mapping the unusual rocks of the Hodges Complex in order to demonstrate whether they represent tectonically emplaced slivers of oceanic lithosphere (an ophiolite complex) or whether they represent plutons, intruded at depth and now exposed at the Earth's surface after millions of years of uplift and erosion. CM has written numerous publications on the bedrock geology of western Connecticut and many of the field descriptions presented below have appeared elsewhere in the scientific literature (Merguerian, 1977, 1983, 1985, 1987).

One of the results of deformation of rocks is the formation of features known as folds and faults. We describe some of the general principles that apply to analysis of geologic structures, paying particular attention to the effects of high confining pressures on resulting structure, and will visit localities where participants can see many kinds of structural features that formed at depths of 20 km and greater based on metamorphic index minerals. Next, we discuss the regional geology of the field-trip route, placing particular emphasis on the plate-tectonic development of this part of the Appalachian mountain belt. We then describe the geology of the West Torrington quadrangle and end with a detailed description of our eight field trip stops planned for today.

We will drive eastward from the Academy across to the FDR Drive and northward on the Drive to the Major Deegan Expressway. From here, we will travel north to the Saw Mill River Parkway to I-684 (North). We continue north on I-684 and cross deeply eroded- and highly folded rocks of the Manhattan Prong. Eventually, we will drive across Cameron's Line, a major structural-stratigraphic dislocation within the Manhattan Prong. Near Brewster, New York, we will drive eastward on Route I-84 to Exit 8 (Route 6) for a brief rest stop, then continue east on I-84 to Conn. Route 8 (North). Route 8, a beautiful transect through the crystalline highlands of western Connecticut, will take us to our first stop of the day near the towns of Litchfield and Harwinton, Connecticut.

Our trip will begin at an outcrop in Harwinton to examine the highly aluminous schist and interlayered gneiss that are characteristic of the Hartland Formation of western Connecticut. From there we will examine the rocks around West Torrington in detail to see the relationships among plutons, Cameron's Line, and surrounding rocks.

So sit back comfortably in the vans, enjoy the scenery, and read your guidebooks for the necessary background for today's On-The-Rocks trip to West Torrington. Consult Table 1 as you read the following discussion. It is a time chart showing geologic time subdivisions shown on the bedrock maps herein, with estimates of numbers of years for their boundaries and a list of some important local geologic events. Table 2 summarizes the major local geologic units (stratigraphy) in terms of layers designated by Roman numerals.

GEOLOGIC BACKGROUND

Under this heading, we discuss geologic structure, the bedrock units, the glacial deposits, and the drainage history of our field-trip route.

GEOLOGIC STRUCTURE - A PRIMER

Geologists use terminology to confuse the layman and to enable them to amass a huge library of terms that are undeniably useless in most social situations. Luckily, our On-The-Rocks trips are an exception. We will not try to bury you in a mountain (how about a deeply eroded mountain range?) of terms to help you understand the major types of structures and geologic features that you will read- and hear about today. But, if you are to understand what we are talking about, you need to know some important definitions. What we hope to do in this section is make the reader aware of the kind of information that enables a geologist to infer that deformation has taken place, to define some of the major geologic structural features formed as a result of tectonic activity, and to evaluate the evidence upon which geologists establish the time when deformation took place. We begin with sedimentary strata and work our way upward through geologic structures to lithosphere plates. The term geologic structure refers to any feature made as a result of deformation related to tectonic activity. Nowadays, we ascribe most tectonic activity to the motion of the Earth's lithosphere plates.

Along the way, we examine some mechanical aspects of rock deformation. Up next are descriptions of folds, faults, effects on sedimentary strata of deformation, structures in sedimentary- vs. metamorphic rocks, and tectonostratigraphic units. We conclude this section with a summary of methods for geologic dating of episodes of deformation.

Strata

The most-important single feature used by geologists to infer that a body of rocks has been deformed is the primary attribute known as stratification. During normal deposition, or settling from a fluid in a rainfall of particles, a thick body of more-or-less featureless sediment may be deposited. The presence of original sedimentary layers, technically known as strata, implies that conditions of deposition changed. As a result, most geologists appreciate the fundamental point that layers in sedimentary rocks imply CHANGE in big letters. The change may have been in the parent area of the sediment, in the sizes of particles supplied, or in the style of deposition.

Thick layers are known as beds and thin layers as laminae. (The word laminae is the plural of lamina. The attribute word applicable to sediments displaying laminae is lamination. Please avoid the temptation to perpetuate the widespread usage indulged in by geologists who don't seem to know their attributes from a hole in the ground, namely the use of lamination in the plural form when they are discussing laminae.)

The particle sizes within a stratum may be uniformly distributed across the bed or may display grading in which larger particles are present at the base of a particular layer and the sizes diminish or "grade" upward into finer particles (Figure 2). A graded bed is the result of a kind of a "lump-sum distribution" from a current carrying a wide range of particles and depositing them within a short time span, largest first and progressively smaller ones later. A common kind of current that deposits layers showing size grading is a gravity-induced turbidity current that flows down a subaqueous slope and crosses a flat part of the basin floor.

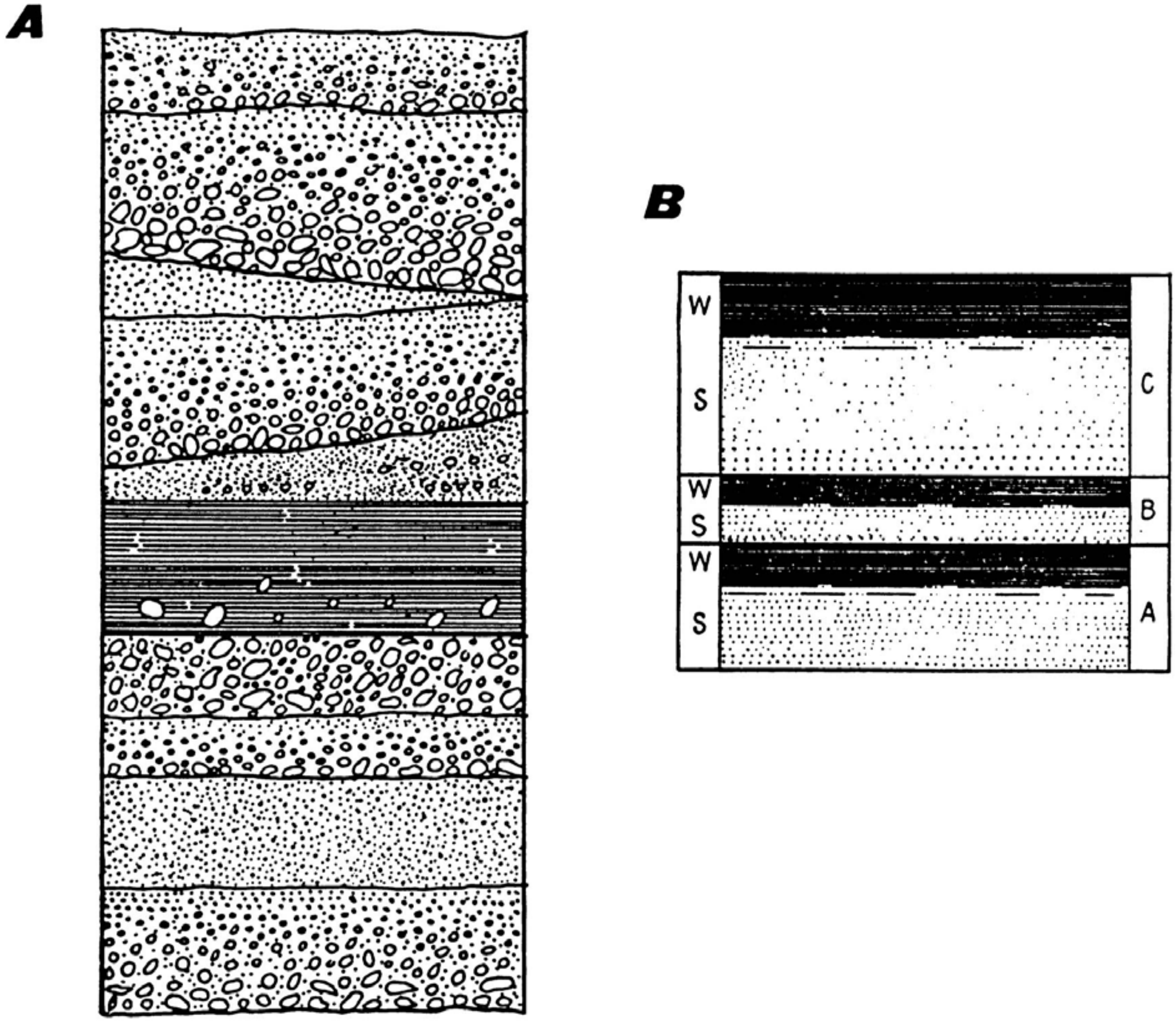


Figure 2. Contrasting kinds of sediments showing internal upward-fining grading.
A) Conglomeratic layers, some graded, some uniform, interbedded with shale containing scattered pebbles. Nonmarine Wamsutta Formation, (Pennsylvanian), E side of Great Pond, Braintree, MA. (R. R. Shrock, 1948, fig. 43, p. 84.)
B) Schematic view of varved silt and -clay deposited in a proglacial lake showing light-colored, coarser graded silty layers deposited during the short summer season when lake is free of ice (marked with S) alternating with dark-colored clay layers deposited during the much-longer winter season when the lake surface is frozen over (marked with W). (R. R. Shrock, 1948, fig. 44, p. 85.)

Fundamental principles about strata were recognized in the 1660s by Nicolaus Steno. He proposed four "rules" for understanding strata of which we include only the following two:

Steno Rule No. 1: Most strata are deposited with an original orientation that is horizontal. (We explain some important exceptions to this rule farther along.) Therefore, strata that are not horizontal usually lost their horizontality as a result of tectonic activity.

Steno Rule No 2: The oldest stratum is at the bottom and successively younger strata occupy higher positions. Two important corollaries of this rule are that each stratum was spread out, one at a time, at the Earth's surface. The materials forming the stratum, therefore, buried a former surface of the Earth. In turn, the top of the stratum was itself such a surface. The top, or face, of a stratum was initially in the up position. (Therefore, if strata are vertical, the tops of the strata indicate the former up direction.) As is discussed in a following section, certain features on the bottoms, within, or on the tops of strata, enable the former top direction to be determined unambiguously. Such features are known as geopetal criteria (Shrock, 1948).

Strata are such fundamental reference surfaces that we emphasize the importance of being able to recognize strata. Where all the particles are about the same size, bedding may not be so easy to identify. This is especially true with uniformly fine sediments (silt size, for example) or with uniformly coarse sediments (boulder gravel). In some cases, such recognition is self evident: materials of contrasting composition or -particle size (Figure 3) form distinct layers that are set off from adjacent layers by prominent surfaces along which the rock separates easily. These are termed bedding-surface partings. In many exposures, the most-prominent partings visible are the bedding-surface (or "bedding-plane") partings. In other exposures, however, tectonic activity has imposed a secondary (structural) parting that may be more prominent than the bedding-surface partings. (Such secondary partings are discussed farther along.)

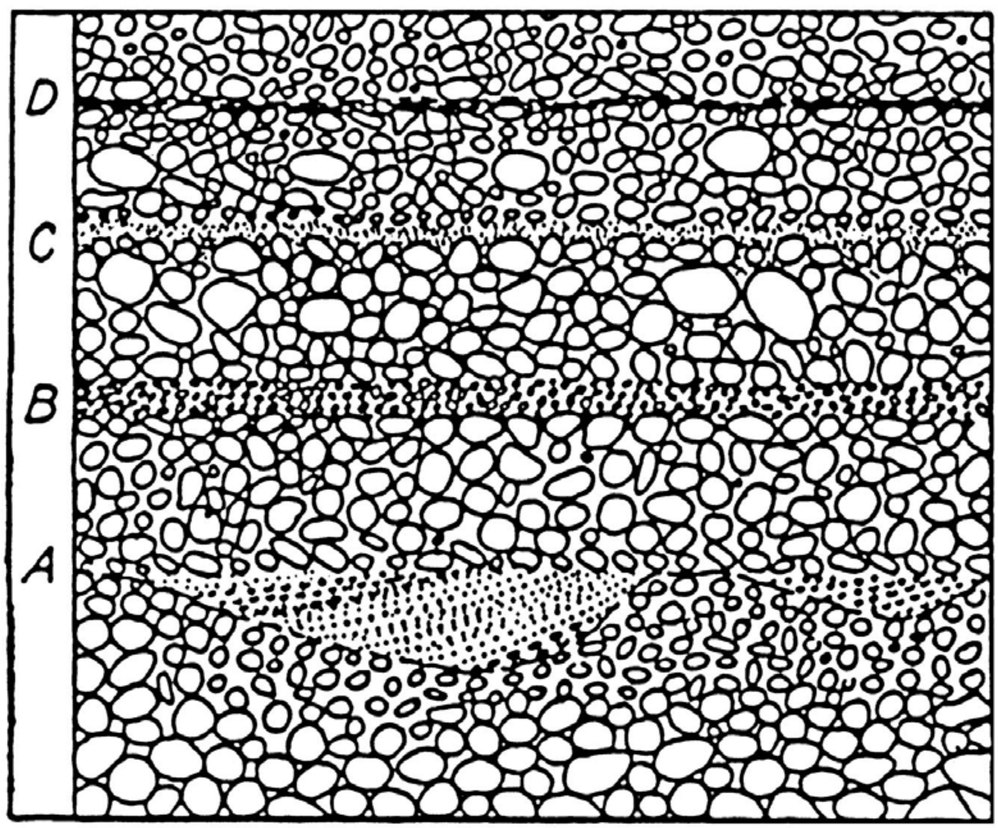


Figure 3. Sketch of gravel (or conglomerate) showing how bedding is revealed by finer sediment in planar layers (such as B, C, and D) or in lenses (as in A). (R. R. Shrock, 1948, fig. 3, p. 12.)

Always make careful note of the feature or features that you have used to support your identification of the bedding. Some such features include changes of color, changes of particle sizes (Figure 4), aligned shells of invertebrates, differences in degree of cementation, or whatever.

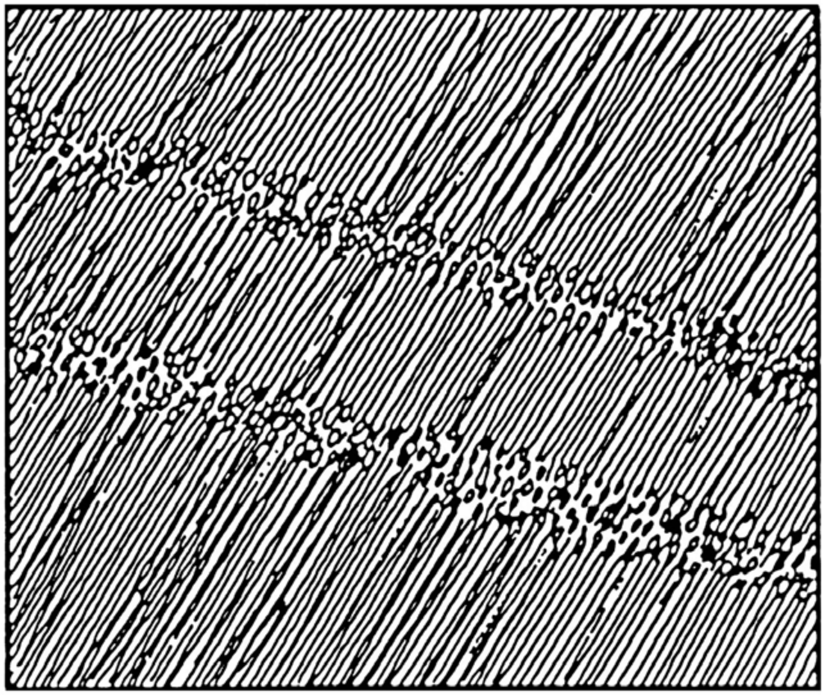


Figure 4. Sketch of slate (dark color) with prominent parting (slaty cleavage) dipping steeply to the left; two layers composed of silt-size sediment (stipple) show that bedding dips gently to the right. (R. R. Shrock, 1948, fig. 7, p. 15.)

Sedimentary Structures

During deposition in a variety of environments, primary- and secondary sedimentary structures can develop above-, below-, and within strata.

During high-energy transport of sand-size or coarser particles (defined as cohesionless sediment), a moving current reacts with the sediment it is transporting and over which it is flowing to create repeating patterns of linear- or curvilinear relief features having long axes that may be transverse to- or parallel with the direction of the current. These are collectively designated as bed forms (Figure 5). In many cases, the tops of sandstone beds display such bed forms, which are named ripples if their relief and crestal-separation distances are measured in centimeters or up to a few tens of centimeters or dunes, if their dimensions measure in meters, tens of meters, or even kilometers. Many bed forms are asymmetric; they slope gently into the current on their upcurrent sides and steeply downcurrent on their downcurrent sides. The shearing-drag effect of the current on the cohesionless-sediment substrate causes these bed forms to migrate downcurrent. They migrate bodily as sediment is eroded from their upcurrent sides and added to their downcurrent sides. The result is a distinctive kind of internal cross strata in which the layers are concave up, tangential at their bases, and truncated at their tops. If the crests of the bed forms are linear, then downcurrent migration creates planar cross strata. (See Figure 5, A.) If the crests of the bed forms are sinuous and concave downcurrent, then downcurrent migration creates trough-type cross strata. (See Figure 5, B.)

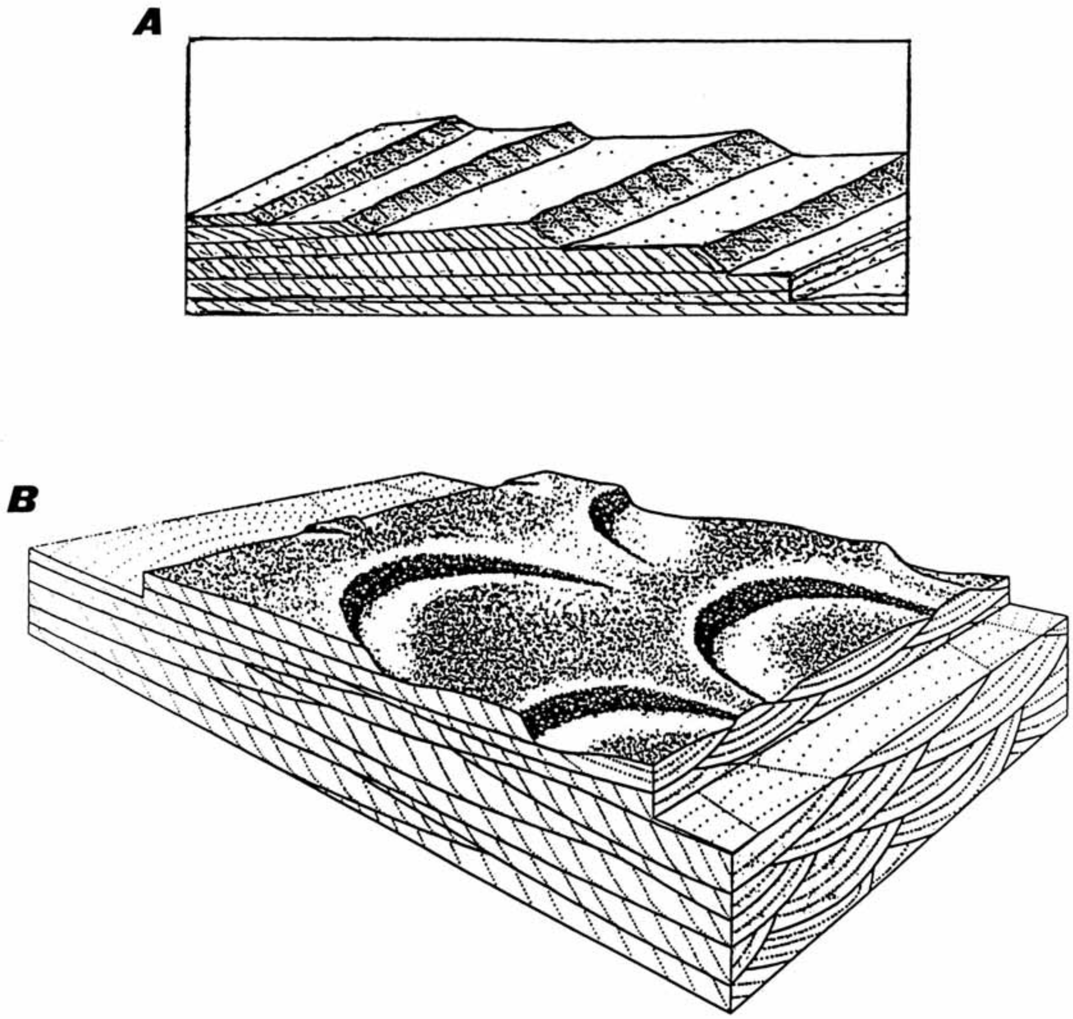


Figure 5. Sketches of contrasting shapes of bed forms created by a current flowing from left to right and cross strata resulting from their downcurrent migration.
A) Linear bed forms create planar cross strata. (G. M. Friedman, J. E. Sanders, and D. C. Kopaska-Merkel, 1992, fig. 5-19A, p. 166.)
B) When cuspate (lunate) megaripples migrate downcurrent trough-type cross strata form. In sections that are parallel to the current, trough cross strata and planar cross strata look about alike. The difference between them is immediately apparent in sections normal to the current. (H. E. Reineck and I. B. Singh, 1980, fig. 52, p. 43.)

Not all cross strata result from downcurrent migration of rhythmic bed forms composed of cohesionless sediment (as in Figure 6, A and D). In some places, local depressions in the bottom are filled in by sand transported from one side and deposited in inclined cross strata

(Figure 6, B). In other places, cross strata are deposited at the fronts of sediment embankments where a water current encounters a deeper place. The embankment lengthens in the downcurrent direction as sediment is deposited in inclined layers (cross strata) along the growing front of the embankment (Figure 6, C and E).

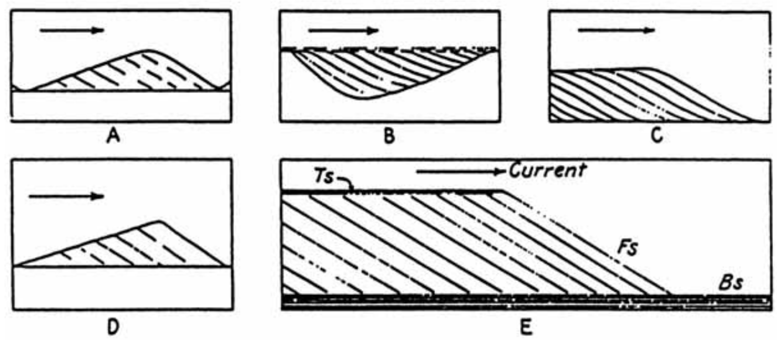


Figure 6. Sketches showing various settings in which cross strata dipping to the right can be deposited by a current flowing from left to right.
A) and D), Longitudinal profiles through individual bed forms that have migrated to the right (downcurrent) by at least one wavelength, thus preserving as internal cross strata many former downcurrent faces.
B), Longitudinal profiles through cross strata that have filled in an asymmetric depression.
C) and E), Longitudinal profiles through cross strata deposited by downcurrent growth of an embankment (such as a delta lobe), built where the flow enters slightly deeper water. (R. R. Shrock, 1948, fig. 207, p. 245.)

Once a current has established a pattern of asymmetric ripples, various kinds of ripple cross laminae are deposited depending on the abundance of sediment. At one extreme, the ripples may migrate and no new sediment is added (Figure 7, A). At the other extreme are ripples that persist as more sand-size- and other sediment is added from the suspended load of the current. Addition of sediment to a field of active ripples creates a kind of rolling-type stratification known as climbing-ripple strata (Figure 7, B).

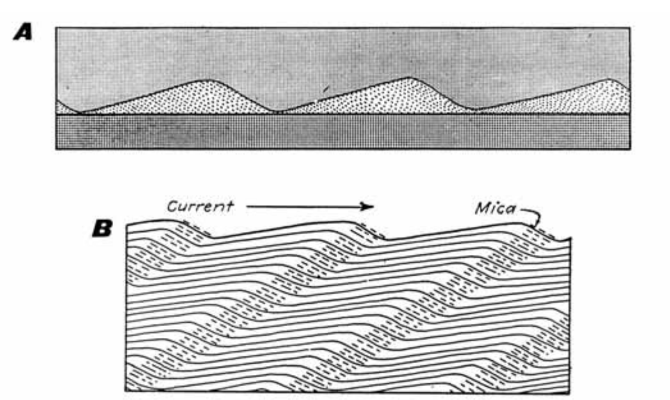


Figure 7. Sketches of ripple cross laminae formed under varying conditions of sediment supply.
A) Cross laminae formed by migration of ripples from left to right when no new sediment is added from the current. (R. R. Shrock, 1948, fig. 57, p. 103.)
B) Climbing-ripple laminae formed when sediment falls out from suspension while the current is still moving fast enough to form ripples. Concentration of mica on the downcurrent faces of ripples creates a large-scale "false bedding" dipping upcurrent (to the left). (R. R. Shrock, 1948, fig. 60, p. 105, based on J. B. Woodworth, 1901b.)

In some settings, the bed-form pattern is not one of regularly spaced linear ridges, but of irregular convex-up hummocks. Deposition in a field of hummocks yields hummocky strata (Figure 8).

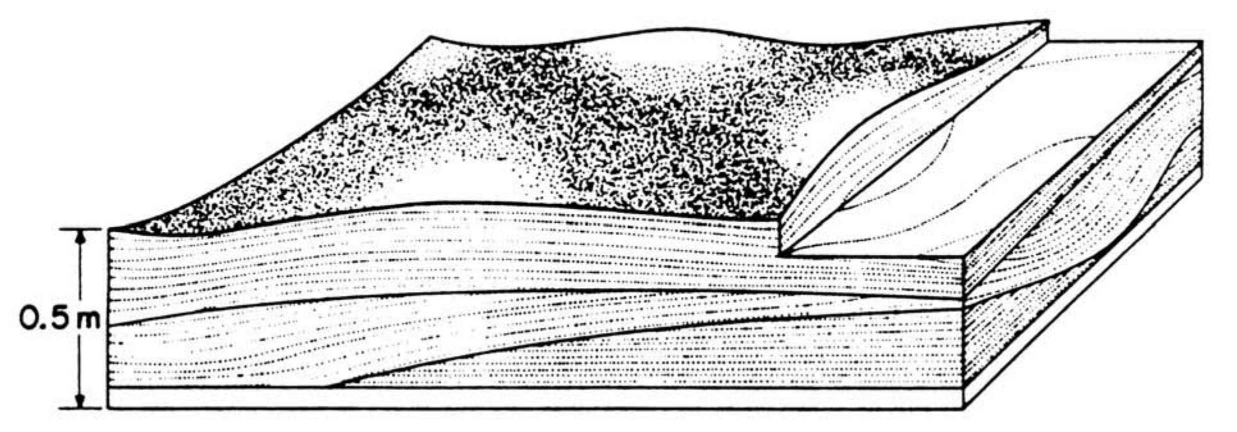


Figure 8. Sketch of hummocky strata. (R. H. Dott, Jr. and Joanne Bourgeois, 1982, fig. 1, p. 663.)

Where current direction oscillates, as it does every few seconds beneath shoaling waves or every few hours in some parts of the intertidal zone, the result may be symmetrical ripples that display pointed crests and broadly rounded, concave-up troughs (Figure 9).

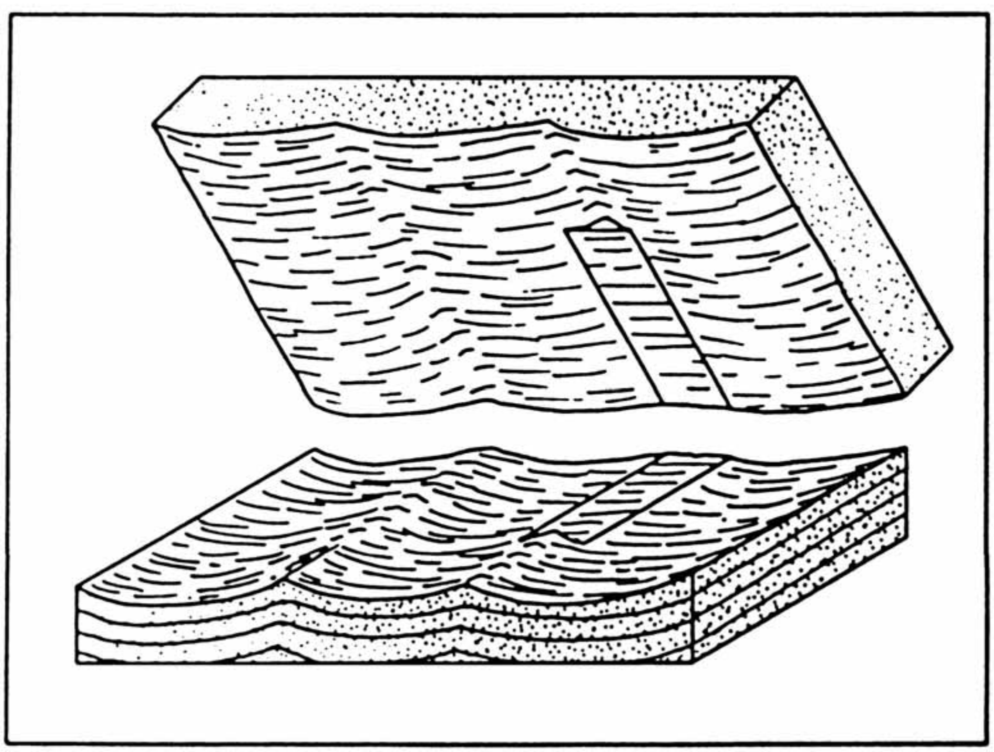


Figure 9. Sketch of symmetrical ripples.

A current carrying sand in suspension that crosses a substrate composed of cohesive fine sediment (as contrasted with the coarser, cohesionless sediment discussed previously) may interact with the bottom to form scour marks or tool marks. These features, sculpted in the cohesive "mud," are usually preserved in the geologic record as counterparts on the base of the overlying sandstone bed. All features found on the bases (i. e., the "soles") of such sandstone beds are collectively designated as sole marks. Patterns of sole marks vary, but many are elongated parallel to the direction of the current (Figure 10).

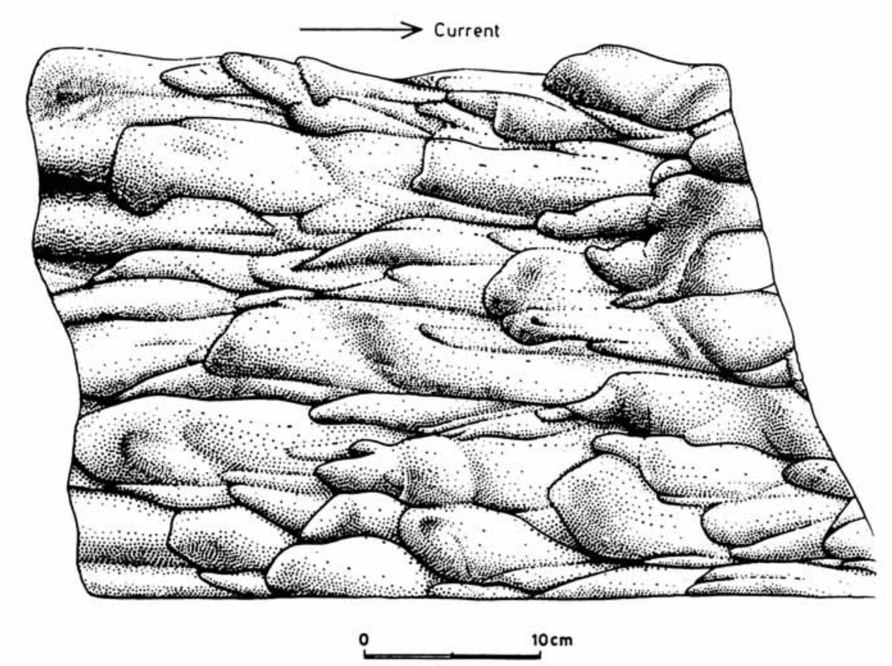


Figure 10. Sketch of counterparts of flutes, sole marks on base of Miocene sandstone bed, Apennines, Italy. (P. E. Potter and F. J. Pettijohn, 1977, fig. 5-2, p. 160, from E. ten Haaf, 1959, fig. 12.)

Cross strata, hummocky strata, and asymmetric current ripple marks deposited by moving currents yield valuable clues for unraveling the paleocurrent directions in which the ancient currents flowed. Many such features are also useful for indicating the original facing direction.

Secondary sedimentary features are developed on already deposited strata and include mud (or desiccation) cracks, rain-drop impressions, and animal footprints. Figure 11 is a composite diagram illustrating common sedimentary structures.

Together, these primary- and secondary sedimentary structures help the soft-rock structural geologist unravel the oft-asked field questions--namely.... Which way is up? and Which way to the package store? The direction of younging of the strata seems obvious in horizontal- or gently tilted strata using Steno's principle of superposition. But steeply tilted-, vertical-, or overturned beds can be confidently unravelled and interpreted structurally only after the true topping (stratigraphic younging) direction has been determined. As we may be able to demonstrate on this field trip, simple observations allow the card-carrying geologist to know "Which way is up" at all times.

It's now time to turn to some geometric aspects of the features formed as a result of post-depositional deformation of rocks in the Earth. We start with a brief lead-in discussion concerning the mechanical aspects of deformation and the strength of materials.

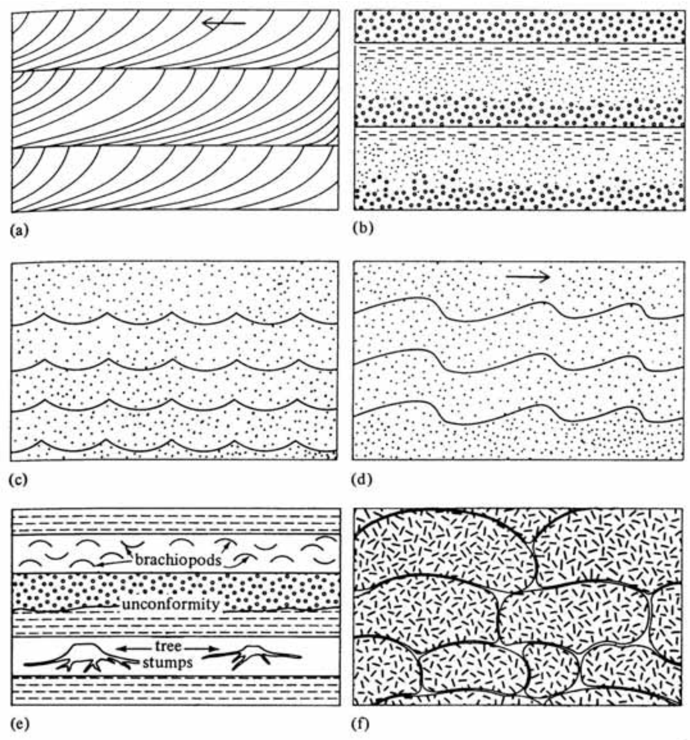


Figure 11. Diagrammatic sketches of of primary sedimentary structures (A through E) and cross sections of pillows (F) used in determining topping (younging) directions in layered rocks.

Mechanical Aspects of Deformation

We begin with some concepts and definitions based on the engineering discipline known as strength of materials. Given today's sophisticated laboratory apparatus, it is possible to subject rocks to temperatures- and pressures comparable to those found deep inside the Earth.

Imagine taking a cylinder of rock out of the Earth and torturing it in a tri-axial compression machine to see what happens. Some geologists get a big charge out of this and tell us (the field geologists) that they really understand how rocks behave under stress. [CM feels that they need to perform these experiments over a longer time frame than a few generations of siblings will allow and thus relies more on observation of field relationships than rock-squeezing data to gain a feel for the complex nature of how rocks are deformed in nature.]

Despite the limitations of the experimental work, measurements in the laboratory on specimens being deformed provide some fundamental definitions. One key definition is the elastic limit, which is the point at which a test specimen no longer returns to its initial shape after the load has been released. Below the elastic limit, the change of shape and/or volume (which is known as strain) is proportional to the stress inside the specimen. Above the elastic limit, the specimen acquires some permanent strain. In other words, the specimen has "failed" internally. Irrecoverable strain manifests itself in the distortion of crystal lattices, grain-boundary adjustments between minerals composing the rock, and minute motions along cleavage- or twin planes.

When differential force is applied slowly (or, according to CM, over long periods of time), rocks fail by flowing. This condition is defined as behaving in a ductile fashion (toothpaste being squeezed out of a tube is an example of ductile behavior). Folds are the result of such behavior. If the force is applied under low confining pressure or is applied rapidly (high strain rates), rocks do not flow, but fracture. This kind of failure is referred to as rocks behaving in a brittle fashion (as in peanut brittle). The result is faults and/or joints. Once a brittle failure (fracture) has begun, it will propagate and may produce offset thus forming a fault surface. Joint surfaces commonly exhibit distinctive "feathers" which show the direction of joint propagation.

In some cases, during deformation, rocks not only undergo simple strain, but also recrystallize. New metamorphic minerals form and newly formed metamorphic minerals acquire a parallel arrangement. More on metamorphic textures later. From the laboratory studies of rock deformation, a few simple relationships are generally agreed upon regarding brittle- and ductile faulting and these are discussed below.

When subjected to differential forces, under high confining pressures and elevated temperatures, rocks (like humans) begin to behave foolishly, squirming in many directions and upsetting the original orientation of primary- or secondary planar- and linear features within them. Geologists try to sort out the effects of deformation by working out the order in which these surfaces or linear features formed using a relative nomenclature based on five letters of the alphabet: D, F, S, L, and M. Episodes of deformation are abbreviated by (Dn), of folding by (Fn), of the origin of surfaces (such as bedding or foliation) by (Sn), of the formation of linear features (such as mineral streaking or intersection lineations produced by intersections of S1 and S0) by Ln, and of metamorphism by (Mn), where n is a whole number starting with 1 (or in some cases, with zero). Bedding, for example, is typically designated as S_0 (or surface number zero) as it is commonly overprinted by S_1 (the first foliation). To use this relative nomenclature to describe the structural history of an area, for example, one might write: "During the second deformation (D_2), F_2 folds formed with the development of an L_2 mineral lineation. An axial-planar S_2 schistosity developed and crosscut both an early foliation (S_1) and bedding (S_0). These features were produced under progressive M_1 metamorphic conditions."

Folds

If layers are folded into convex-upward forms we call them anticlines. Convex-downward fold forms are called synclines. In Figure 12, note the geometric relationship of anticlines and synclines. Axial planes (or axial surfaces) physically divide folds in half. Note that in Figure 12, the fold has been deformed about a vertical axial surface and is cylindrical about a linear fold axis which lies within the axial surface. The locus of points connected through the domain of maximum curvature of the bedding (or any other folded surface of the fold) is known as the hinge line (which is parallel to the fold axis). This is geometry folks; we have to keep it simple so geologists can understand it.

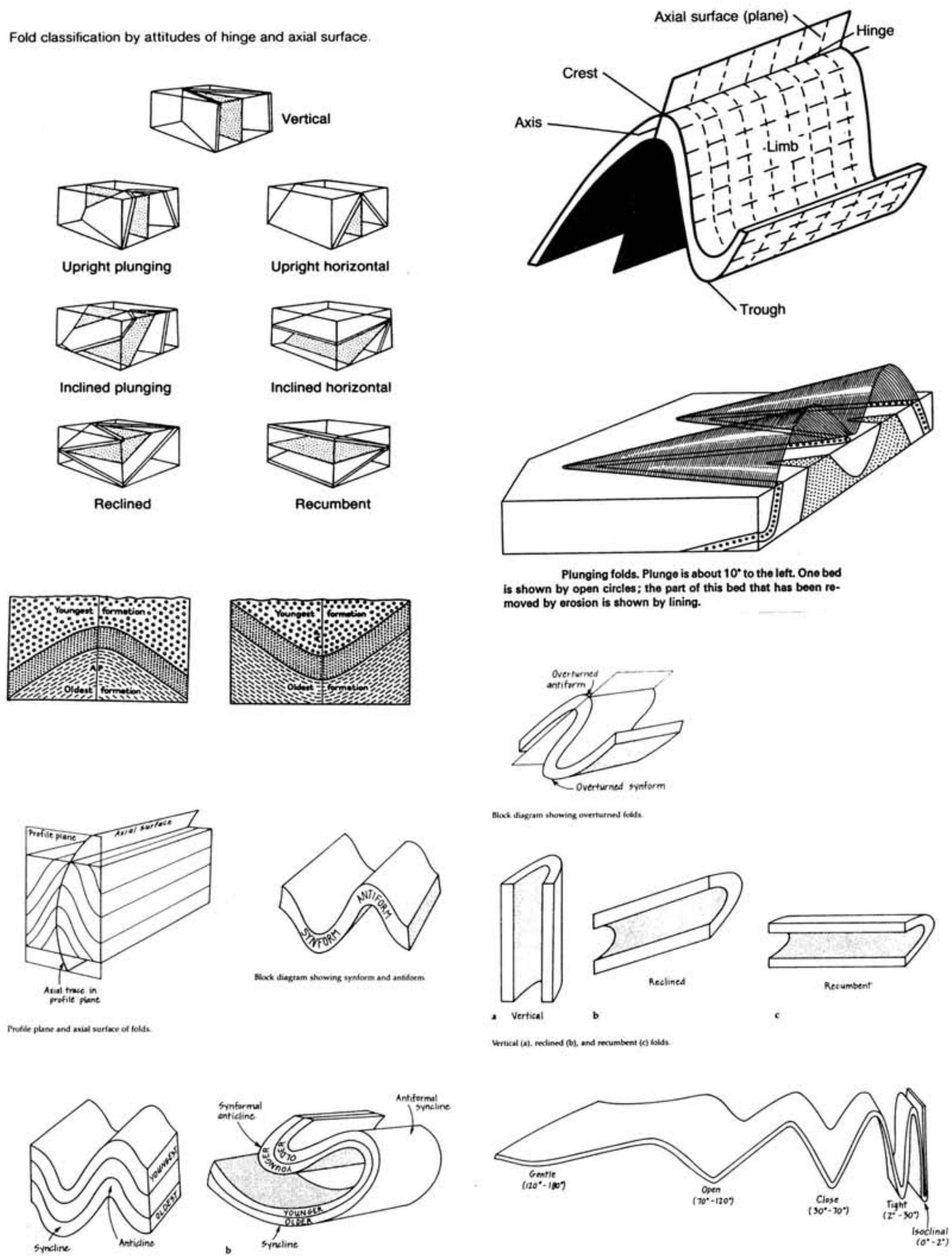


Figure 12. Composite diagram from introductory texts showing various fold styles and nopersonsclature (non-sexist terminology) as discussed in the text.

In eroded anticlines, strata forming the limbs of the fold dip away from the central hinge area or core (axis) of the structure. In synclines, the layers forming the limbs dip toward the hinge area. Given these arrangements, we expect that in the arches of eroded anticlines, older stratigraphic layers will peek through whereas in the eroded troughs of synclines, younger strata will be preserved (Figure 13).

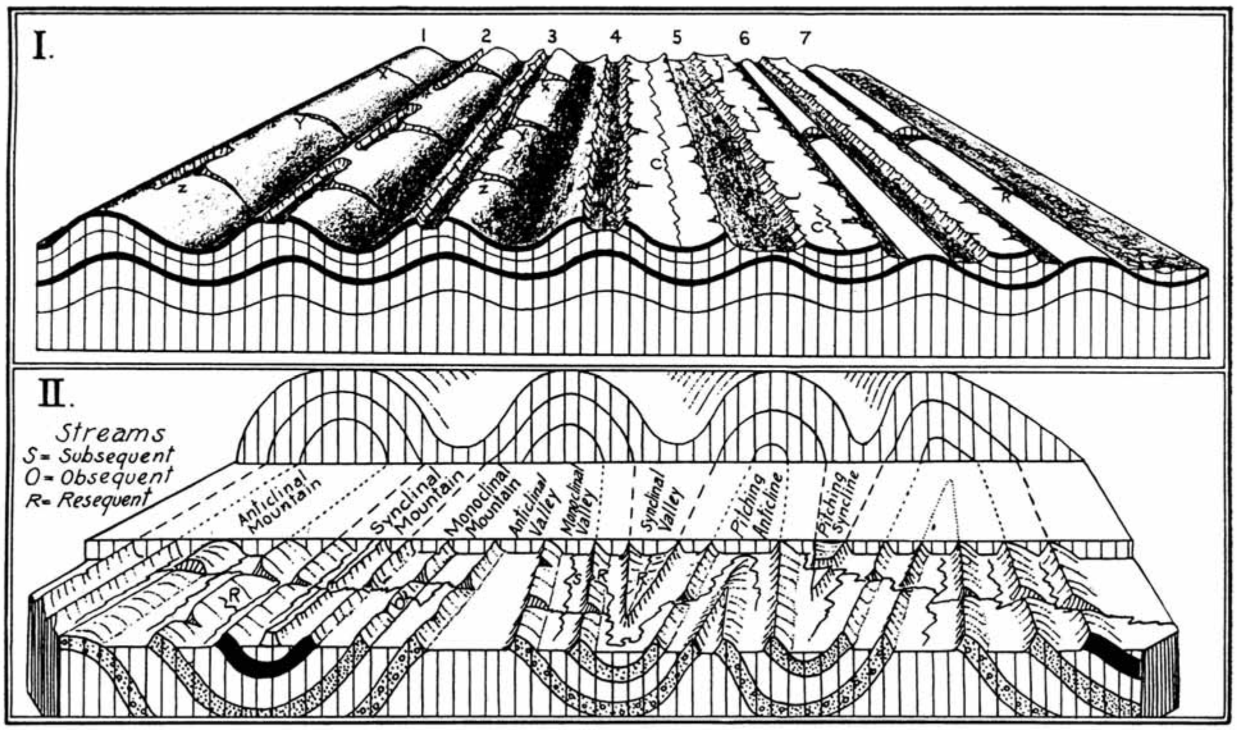


Figure 13. Schematic block diagram through eroded anticlines and -synclines showing how resistant layers form linear ridges and weak layers, linear valleys. The relationships shown here are common in the Appalachian Valley and Ridge Province of central Pennsylvania. (Lobeck, 1939, unnumbered figure, p. 588.)

In metamorphic terranes, field geologists are not always sure of the correct age relationships of the metamorphosed strata. Therefore, it is helpful to make use of the general terms antiform and synform which describe the folds by whether they are convex upward (antiform) or concave upward (synform) but do not imply anything about the relative ages of the strata within them.

Realize that in the upright folds shown in Figure 12, axial surfaces are vertical and fold axes, horizontal. This is a very rare case. Generally the axial surfaces are not vertical and the fold axes are not horizontal (Compare figures 12 and 13).

The scale of folds varies through an enormous range from tiny features that can be present in a hand specimen to great structures measuring many kilometers across and hundreds of kilometers long. A large anticline having a broad wave length and displaying smaller folds along its limbs is known as an anticlinorium. A companion large-scale synclinal feature is a synclinorium.

After the pioneering work of William Smith in 1812, who first mapped a large-scale plunging structure (we would call this a synclinorium today) in the southern England lowland areas, the Appalachians were the first mountains in which major geologic structures known as folds (anticlines and synclines) were demonstrated (by the Rogers brothers, H. D., and W. B., in the middle of the nineteenth century from their studies in Pennsylvania). To be sure, small folds had been recognized where seen in coastal exposures in numerous localities in western Europe. But, it was a giant step (and anything but an intuitively obvious leap) from seeing small folds in cross section to the reconstruction of very large-scale folds based on working out the stratigraphic relationships of Paleozoic strata, thousands of meters thick, underlying strike ridges that extend for tens, even hundreds of kilometers. (See Figure 13.)

Folds could care less about the orientation of their axes or axial surfaces and you can certainly imagine that axial surfaces can be tilted, to form inclined or overturned folds. Or the axial surfaces may be sub-horizontal, in which case the term recumbent folds is used. In both overturned folds and recumbent folds, the fold axes may remain subhorizontal. (See Figure 12.) It is also possible for an axial surface to be vertical but for the orientation of the fold axis to range from horizontal to some angle other than 0ø (thus to acquire a plunge and to produce a plunging fold). Possible configurations include plunging anticlines (or -antiforms) or plunging synclines (or -synforms). Vertical folds (plunging 90ø) are also known; in them, the terms anticline and syncline are not meaningful. In reclined folds, quite common in ductile shear zones, the fold axes plunge directly down the dip of the axial surface.

In complexly deformed mountain ranges, most terranes show the superposed effects of more than one set of folds and faults. As a result of multiple episodes of deformation, the ultimate configuration of folds can be quite complex (i. e., plunging folds with inclined axial surfaces and overturned limbs).

We need to mention one additional point about the alphabet soup of structural geology. Seen in cross section, folds fall into one of three groups, the S's, the M's, and the Z's. Looking down plunge in the hinge area of a northward-plunging anticlinal fold, for example, dextral shearing generates asymmetric Z folds on the western limb and sinistral shearing forms S folds on the eastern limb. Usually only one variety of small, asymmetric folds will be found on a given limb of a larger fold. Therefore, if one notices a change in the pattern from S folds to Z folds (or vice versa), one should be on the lookout for a fold axis. The hinge area is dominated by M folds (no sense of asymmetry).

One final note on folding -- it is generally agreed, in geologically simple areas, that axial surfaces form perpendicular to the last forces that ultimately produced the fold. Therefore, the orientation of the folds give some hint as to the direction of application of the active forces (often a regional indicator of relative plate convergence). In complex regions, the final regional orientation of the structures is a composite result of many protracted pulses of deformation, each with its unique geometric attributes. In these instances, simple analysis is often not possible. Rather, a range of possible explanations for a given structural event is commonly presented.

Faults

A fault is defined as a fracture along which the opposite sides have been displaced. The surface of displacement is known as the fault plane (or fault surface). The enormous forces released during earthquakes produce elongate gouges within the fault surface (called slickensides) that may possess asymmetric linear ridges that enable one to determine the relative motion between the moving sides (Figure 14, inset). The block situated below the fault plane is called the footwall block and the block situated above the fault plane, the hanging-wall block.

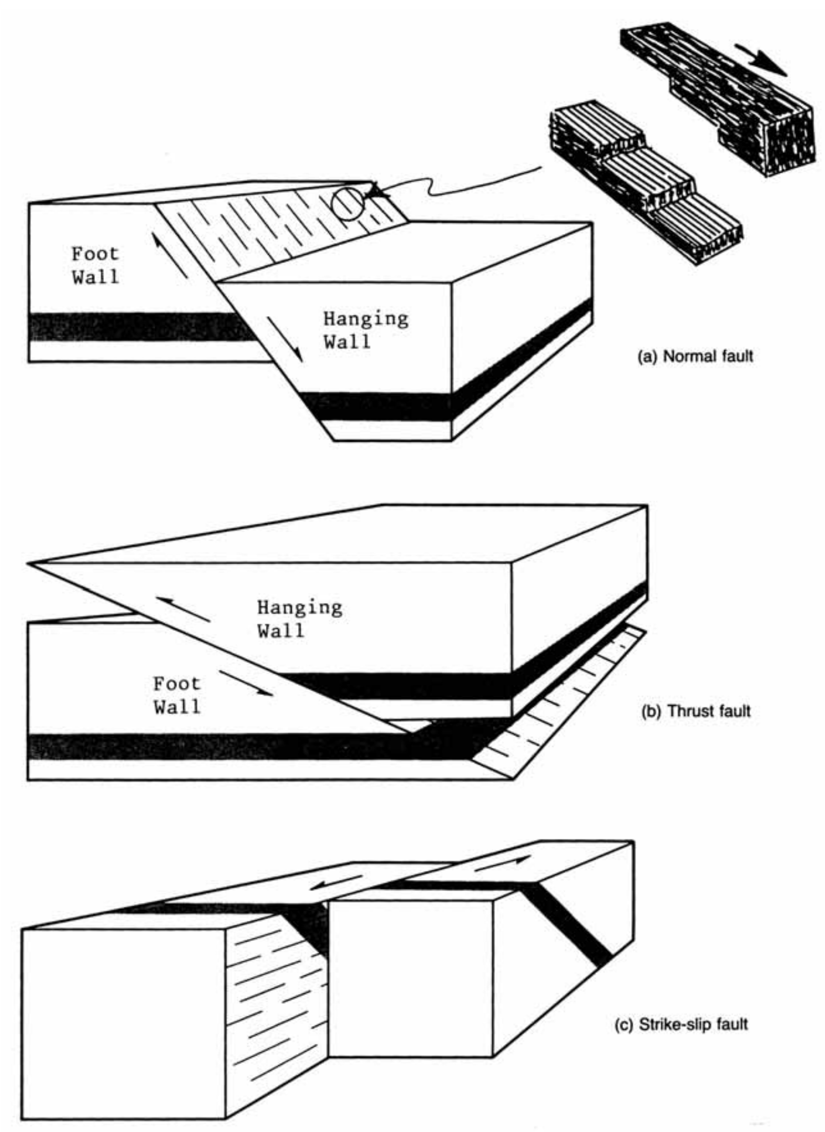


Figure 14. The three main types of faults shown in schematic blocks. Along a normal fault (A) the hanging-wall block has moved relatively downward. On a thrust fault (or reverse fault) (B) the hanging-wall block has moved relatively upward. Along a strike-slip fault (C), the vertical reference layer (black) has been offset by horizontal movement (left-lateral offset shown here). Inset (D) shows segments of two blocks along a slickensided surface show how the jagged "risers" of the stairsteps (formed as pull-apart tension fractures) can be used to infer sense of relative motion. [(A), (B), (C), Composite diagram from introductory texts; (D), J. E. Sanders, 1981, fig. 16.11 (b), p. 397.]

Normal- and Reverse Faults

Imagine extending (stretching) or compressing a block-like portion of the Earth's crust. Extensional force may causes the crust to rupture along a fracture that is not vertical. In this case, the block above the dipping fracture, known as the hanging-wall block, will slide down the fracture surface producing a normal fault. [See Figure 14 (a).] Compressive forces drive the hanging-wall block up the fracture surface to make a reverse fault. A reverse fault with a low angle (<30ø) is called a thrust fault. [See Figure 14 (b).] In all of these cases, the slickensides on the fault will be oriented more or less down the dip of the fault plane and the relationship between the tiny "risers" that are perpendicular to the striae make it possible to determine the relative sense of motion along the fault. Experimental- and field evidence indicate that the asymmetry of slickensides is not always an ironclad indicator of relative fault motion. As such, displaced geological marker beds or veins are necessary to verify relative offset. Fault motion up- or down the dip (as in normal faults, reverse faults, or thrusts faults) is named dip-slip motion.

Low-angle Thrusts

A low-angle thrust is a special kind of reverse fault that initially contained one or more segments that are parallel to the originally horizontal strata. Such low-angle faults have also been referred to as overthrusts, but this term implies a sense of motion that may not be correct. In order to beg the question of whether motion was one of overthrusting or underthrusting, P. B. King (1960) advocated use of the term low-angle thrust.

The large-scale repetition of strata on low-angle thrusts was first shown in the Scottish Highlands. Soon thereafter, spectacular examples were found in the Alps and, indeed, in nearly all mountain chains. Studies in the Appalachians made possible new understanding between thrusts and folds. Examples were found illustrating all gradations from small breaks across the axes of overturned folds (Figure 15) to what are known as imbricate thrusts in which the deformed strata and the overthrusts dip southeastward at about the same angles; during deformation, the right-way-up strata of the northwest (normal) limbs of two synclines have been brought together and the southeastern (overturned) limbs and the central parts of the intervening anticlines have vanished (Figure 16).

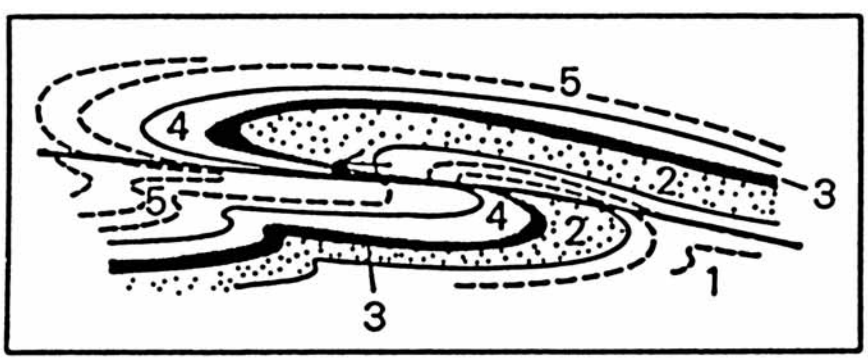
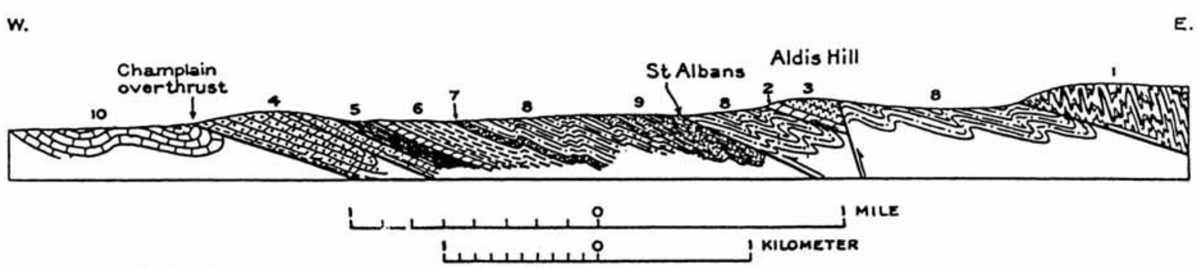


Figure 15. Overturned fold broken by a low-angle thrust fault, schematic profile section. Units are numbered in order of decreasing age, from 1 (oldest) to 5 (youngest). (J. E. Sanders, 1981, fig. 16.14, p. 398.)



Section from west to east near St. Albans, Vermont, showing the principal thrusts at that latitude and the three sequences of Paleozoic formations separated by the thrusts. 1, Undifferentiated pre-Cambrian and lower Paleozoic; 2, dolomite and schist, probably late pre-Cambrian; 3, Lower Cambrian quartzite (eastern sequence); 4–9, central sequence; 4, Lower Cambrian dolomite; 5, Lower Cambrian slate and dolomite; 6, Middle Cambrian slate; 7, conglomerate, base of Upper Cambrian (thickness exaggerated); 8, Upper Cambrian slate; 9, Lower Ordovician slate; 10, Ordovician of the western sequence. The coarse conglomerate at the base of the Ordovician in the central sequence (No. 9) is not exposed in the vicinity of this section

Figure 16. Imbricate thrusts that are essentially parallel to the strata that have been duplicated; example from Paleozoic strata in northwestern Vermont. (C. R. Longwell, 1933, fig. 14, p. 63.)

Studies in the Appalachians led John L. Rich (1934) to propose the concept of "bedding thrusts." By this term, he referred to overthrusts along which two contrasting segments can be recognized: (1) segments that are parallel to the bedding; and (2) segments that cut across bedding at steep angles (Figure 17). (These segments that cut bedding at steep angles have subsequently been named ramps.) What was totally different about Rich's analysis is the relationship between thrusts and folds. Because of the geometric arrangement of the ramps and the beds, any forward displacement causes the strata of the upper block to be folded. As the strata are pushed against the ramp, they become parallel to it, forming one limb of a ramp-related anticline. Where the strata that have been displaced past the ramp return to the next bedding-parallel segment of the thrust surface, they dip downward toward this surface, thus forming the second limb of the ramp-related anticline. Where later deformation has not obscured the relationships, Rich's mechanism creates flat-topped anticlines whose widths are direct functions of the amount of displacement on the thrust and intervening flat-bottom synclines whose widths are determined by the spacing between adjacent ramps.

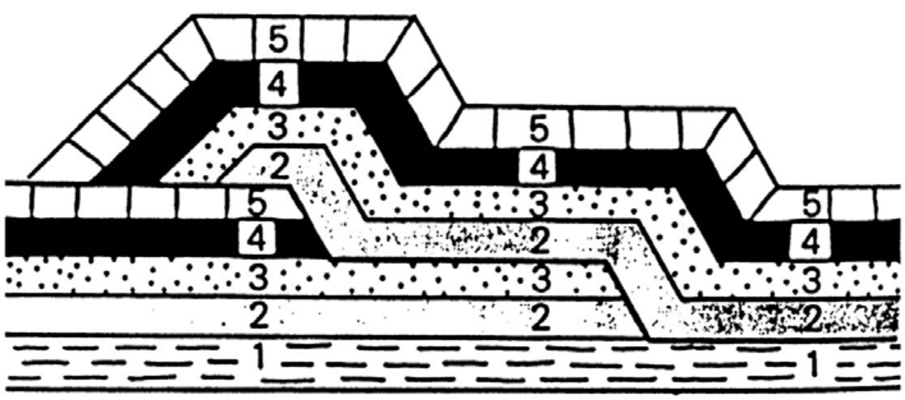


Figure 17. Folds formed by movement on a bedding-plane thrust according to the mechanism proposed by J. L. Rich; schematic profile section with units numbered in order of decreasing age from 1 (oldest) to 5 (youngest). (J. E. Sanders, 1981, fig. 16.1e, p. 390.)

For about 35 years, the only person who seems to have understood the fundamental new point that John L. Rich was trying to make was John Rodgers. In numerous important papers, Rodgers (1949, 1950, 1953, 1963, 1964, 1970) established what he referred to as the "thin-skinned" mechanism of Appalachian deformation. A synonym of the "thin-skinned" style of Appalachian deformation was the term "no-basement" style. That is, the deformation of the strata took place independently of the basement. Therefore, in between the deformed strata and the non-involved basement was a surface of detachment, or d‚collement. Where the basement was involved, the term "thick-skinned" style was applied. These two styles were thought to be mutually exclusive.

Major new understanding of the importance of John L. Rich's insights have resulted from the discovery of large reserves of petroleum in the Rocky Mountains. For many years, searchers for petroleum avoided drilling in places where the strata had been overthrust. They believed that deformation on a scale that creates overthrusts would destroy any petroleum present in the strata. Therefore, overthrust belts ranked high on the list of places to avoid. Much to everybody's surprise, major discoveries were made by drilling through overthrusts to petroleum traps formed in the strata on the lower block (Gries, 1983). After several giant gas fields had been discovered, "overthrust belts" were stricken from the "no-no" list and quietly moved to the head of the list of places to be explored. As a result, seismic profiles were shot across mountain chains and deep holes were drilled through complex geologic structures. The result has been the acquisition of great quantities of new subsurface information from places that probably would have remained "terra incognita" forever had it not been for petroleum exploration.

All of which brings us back to the Appalachians. The next major point about Appalachian overthrusts is the displacement of basement rocks over the strata. In other words, the basement becomes "involved." (What's that story about ham and eggs? The hen is "involved" but the pig is "committed.") In this respect, northwestern New Jersey and adjacent southeastern New York provides some critical evidence that most Appalachian geologists have overlooked. In the discussions about "thin-skinned" vs. "thick-skinned" deformation, this part of the Appalachians seems to have been studiously avoided The evidence pointing to basement involvement in the northern Appalachian overthrusts was presented by Isachsen (1964). Isachsen argued that on a regional scale, the Proterozoic basement rocks had been thrust over the Paleozoic sedimentary strata (Figure 18). He inferred that this displacement had taken place during the Late Ordovician Taconian orogeny. (We discuss the subject of age of deformation in a following section.)

A critical locality demonstrating the displacement of Proterozoic rocks over Paleozoic strata is the Musconetcong tunnel built by the Lehigh Railroad more than 100 years ago. Isachsen (1964, p. 822) cited K□mmel (1940) as his source of the information about this tunnel. As we pointed out in our On-The-Rocks Guidebook for Trip 12 to Franklin Furnace (17 June 1990), however, the correct citation is not K□mmel (1940) but rather Lewis and K□mmel (1915; fig. 3, p. 58).

Over the years, field geologists have noted special geologic features associated with thrust faults. Because they propagate at low angles with respect to bedding, thrusts commonly duplicate strata. In addition, thrust faults can displace strata for great distances and wind up transporting rock deposited in one environment above rocks deposited in markedly disparate environments. In such cases, we call the displaced strata of the upper plate above a thrust fault an allochthon or describe an entire displaced sequence of strata as an allochthonous terrane (see Tectonostratigraphic Units below). In other words, allochthonous rocks were not originally deposited where they are now found. By contrast, regions consisting of rock sequences that were originally deposited where they are now found constitute an autochthon or autochthonous terrane.

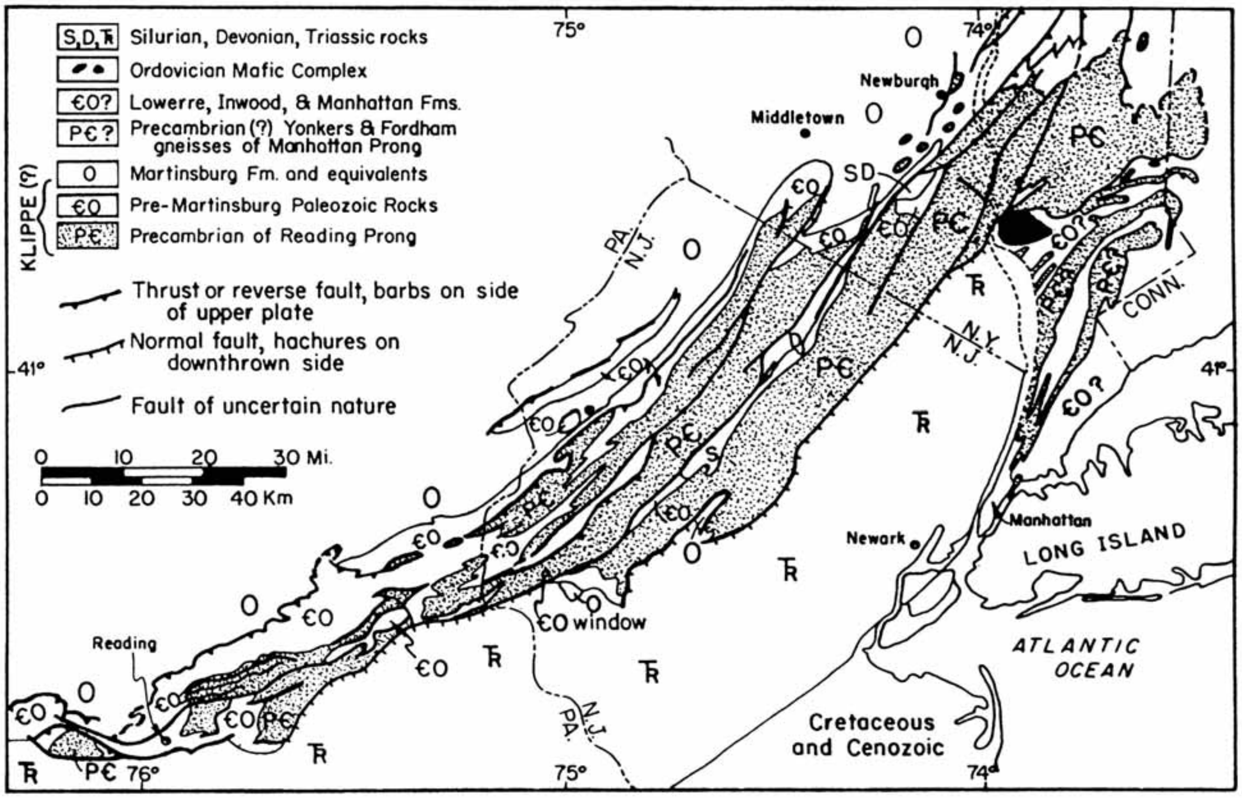


Figure 18. Sketch map showing extent of Reading Prong klippe [stippled areas]. (Yngvar Isachsen, 1964, fig. 5, p. 821.)

Interesting geometric patterns result from the erosion of overthrust sheets of strata that have been folded after they were overthrust. Where a "hole" has been eroded through the upper plate (allochthon), it is possible to peer downward through the allochthon and see the autochthon exposed in a window (synonyms: inlier, or, in the German tongue, fenster) surrounded by the trace of the thrust fault that was responsible for the dislocation (Figure 19). By contrast, if most of the upper plate has been eroded, only a remnant outlier or klippe may remain. (See Figure 19.) Both klippen and windows (or fensters) produce similar map-scale outcrop patterns. The difference is that the thrust surface typically dips toward the center of a klippe (a remnant of the allochthon) and away from the center of a window (which shows a part of the underlying autochthon).

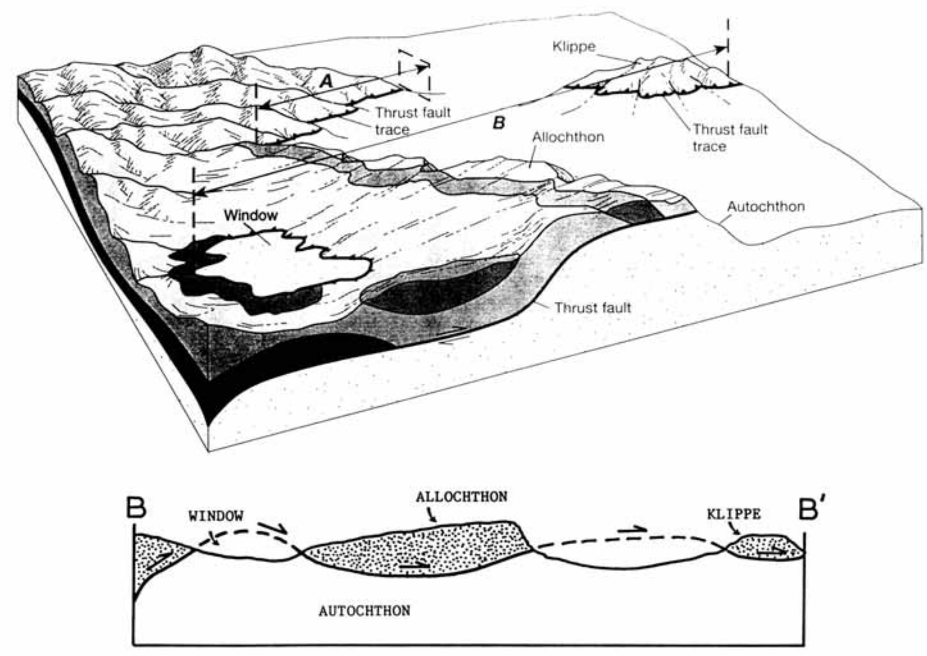


Figure 19. Block diagrams illustrating the relationships among major components of low-angle thrust sheets, including allochthons, autochthons, klippen, and windows. (R. J. Twiss and E. M. Moores, 1992, fig. 6.4, p. 99; section B-B' redrawn by CM.)

Bedding-plane thrusts are more-localized features but are geometrically the same as thrust faults in that they involve layer-parallel shortening of strata and produce low-angle imbrication of strata. They can easily be "missed" in the field but result in overthickening of strata and can produce anomalous stratigraphic thickness in sedimentary units. The field geologist can identify them by careful bed-by-bed examination of known sequences based on duplication of key- or marker beds and by identification of highly veined rocks adjacent to- or zones of gouge or slickensides along the surfaces of dislocation.

During episodes of mountain building associated with continuous subduction and/or collisions near continental margins, thrusting is typically directed from the ocean toward the continent. Accordingly, one of the large-scale effects of such periods of great overthrusting is to impose an anomalous load on the lithosphere which causes it to subside and form a foreland basin. These basins receive tremendous quantities of sediment which fill the basin with debris derived from erosion of uplifted areas within the active collision zone. In the late stages of convergence, forces transmitted from the collision zone into the developing foreland basin create a diachronous secondary stage of folding and continent-directed overthrusting of the strata filling the foreland basin. Thus, a thrust may override debris eroded from it (Sanders, 1995).

Strike-slip Faults

Rather than simply extending or compressing a rock, imagine that the block of rock is sheared along its sides (i. e., that is, one attempts to rotate the block about a vertical axis but does not allow the block to rotate). This situation is referred to as a shearing couple and could generate a strike-slip fault. [See Figure 14 (c).] On a strike-slip-fault plane, slickensides are oriented subhorizontally and again may provide information as to which direction the blocks adjoining the fault surface moved.

Two basic kinds of shearing couples and/or strike-slip motion are possible: left lateral and right lateral. These are defined as follows. Imagine yourself standing on one of the fault blocks and looking across the fault plane to the other block. If the block across the fault from you appears to have moved to the left, the fault is left lateral [illustrated in Figure 14 (c)]. If the block across the fault appears to have moved to the right, the motion is right lateral. Convince yourself that no matter which block you can choose to observe the fault from, you will get the same result! Naturally, complex faults show movements that can show components of dip-slip- and strike-slip motion, rotation about axes perpendicular to the fault plane, or reactivation in a number of contrasting directions or variety. This, however, is no fault of ours.

Distinctive Fault Rocks

Tensional-, compressional, or strike-slip faulting results in brittle deformational response at crustal levels above 10 to 15 km. Such faulting is episodic and accompanied by seismicicity and the development of highly crushed and granulated rocks called fault breccias and cataclasites (including fault gouge, fault breccia, and others). Figure 20 lists brittle- and ductile fault terminology as adapted from Sibson (1977) and Hull et al. (1986). Begining at roughly 10 to 15 km and continuing downward, rocks under stress behave aseismically and relieve strain by recrystallization during ductile flow. These unique metamorphic conditions prompt the development of highly strained (ribboned) quartz, feldspar porphyroclasts (augen), and frayed micas, among other changes, and results in highly laminated rocks called mylonites. (See Figure 20.)

The identification of such ductile-fault rocks in complexly deformed terranes can be accomplished only by detailed mapping of metamorphic lithologies and establishing their geometric relationship to suspected mylonite zones. Unfortunately, continued deformation under load often causes early formed mylonites to recrystallize and thus to produce annealed mylonitic textures (Merguerian, 1988), which can easily be "missed" in the field without careful microscopic analysis. Cameron's Line, a recrystallized ductile shear zone showing post-tectonic brittle reactivation, is an original ductile fault zone (mylonite) having a complex geologic history.

Effects on Sedimentary Strata of Deformation

The most-obvious effect of deformation on sedimentary strata is change of attitude: originally horizontal strata are no longer horizontal. Apart from such changes, other indicators of deformation include displacement of strata, disruption of strata, and rock cleavage (Figure 21).

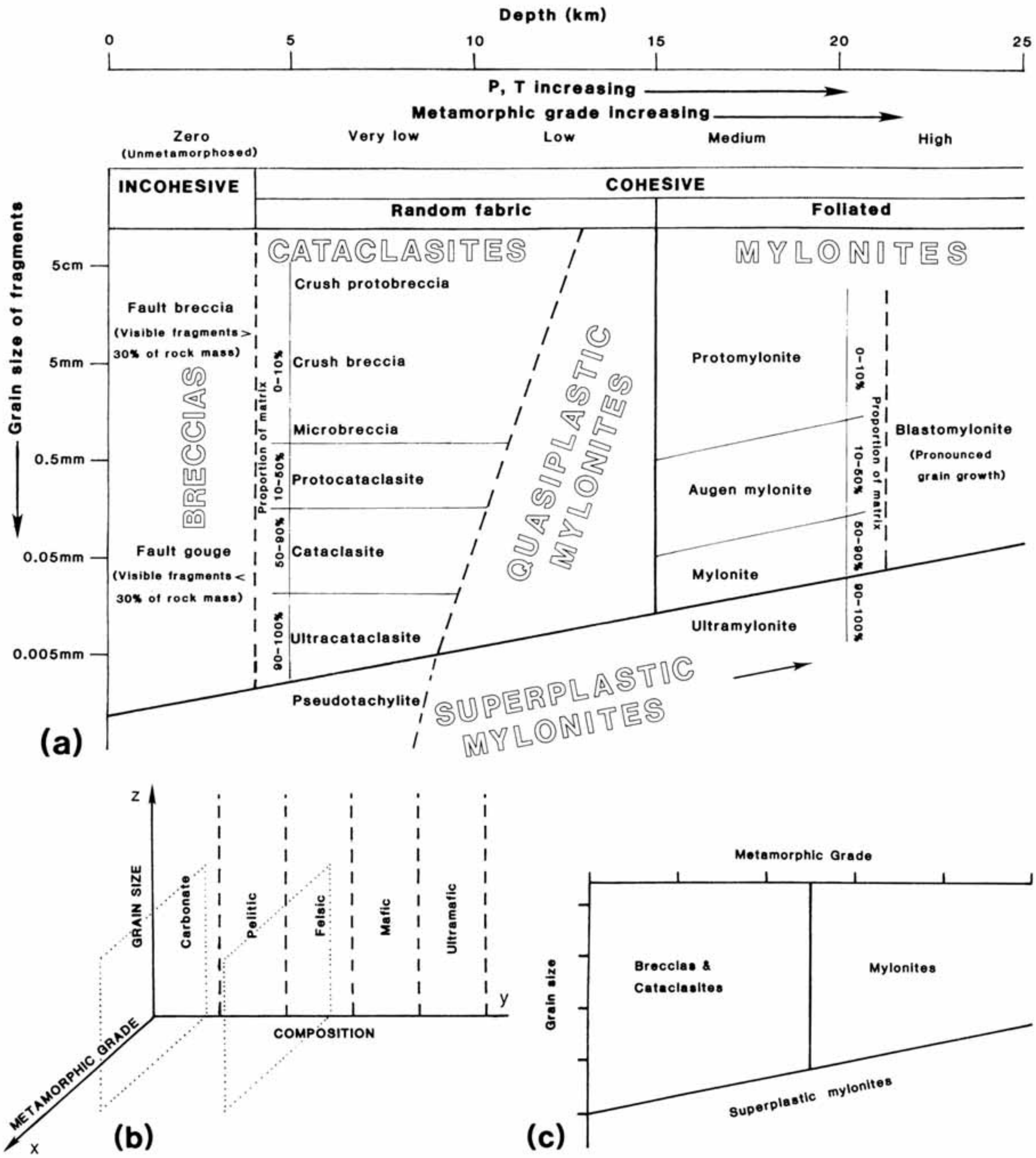


Figure 20. Fault-rock terminology. A) Classification of fault rocks that have been derived from quartzo-feldspathic lithologies (e. g. granite). (Adapted from R. Sibson, 1977.) B) The particle size - metamorphic grade - lithologic composition grid used for classifying fault rocks. (After J. Hull, R. Koto, and R. Bizub, 1986.) C) Fault-rock diagram for marl showing expanded mylonite (sic) and superplastic mylonite fields as compared to those shown on the diagram for granite in A) (R. Stephen Marshak and Gautam Mitra, 1988, fig. 11-23, p. 227.)

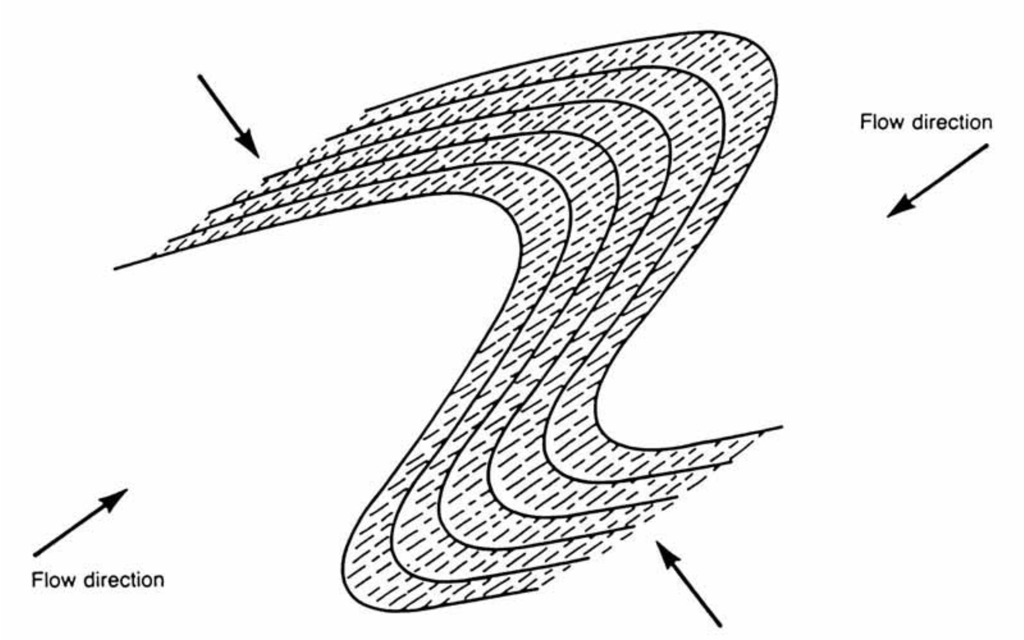


Figure 21. Sketch of slaty cleavage oriented parallel to axial plane of folds. (R. D. Hatcher, Jr., 1990, fig. 15-22, p. 335.)

Where two sets of cleavage have been superimposed upon each other (or more than two sets of cleavage on one another), the rocks tend to break into wedge-like, triangular shapes or into long, slender faceted "pencils." This property of shape is further promoted if the deformed rocks contain a planar parting parallel to bedding.

In regions where several episodes of deformation intense enough to form rock cleavage have been superimposed, it may be very difficult to unravel the relative ages of cleavage. If deformation has beem along subparallel trends, tlhis problem may become acute. Normally, cleavage direction parallels the axial surface of folds. That is, the layer-type minerals whose parallelism causes the cleavage are oriented parallel to the plane of maximum compression. (See Figure 21.)

Structures in Sedimentary- vs. Metamorphic Rocks

For hard-rock geologists working in metamorphic terranes, simple sedimentary observations will not allow the card-carrying geologist to know "Which way is up" at all. Rather, because of intense transposition and flow during ductile deformation, stratification, fossils for age dating, tops and current-direction indicators are largely useless except to identify their hosts as sedimentary protoliths. Thus, according to CM, "at the outcrop scale, metamorphism can best be viewed as the great homogenizer." Commonly during metamorphism, the increase in temperature and -pressure and presence of chemically active fluids severely alter the mineral compositions and textures of pre-existing rocks. As a result, in many instances, typical soft-rock stratigraphic- and sedimentologic analysis of metamorphic rocks is not possible.

Keep in mind that folding under metamorphic conditions commonly produces a penetrative mineral fabric with neocrystallized minerals (typically micas and amphiboles) aligned parallel to the axial surfaces of folds. Such penetrative metamorphic fabrics are called foliation, if primary, and schistosity, if secondary. Minerals can also become aligned in a linear fashion producing a metamorphic lineation. Such features can be useful in interpreting a unique direction of tectonic transport or flow direction. Because folds in metamorphic rocks are commonly tight- to isoclinal (high amplitude-to-wavelength aspect ratio) with limbs generally parallel to axial surfaces, a penetrative foliation produced during regional dynamothermal metamorphism will generally be parallel to the re-oriented remnants of stratification (except of course in the hinge areas of folds). Thus, in highly deformed terranes, a composite foliation + remnant compositional layering is commonly observed in the field. Departures from this common norm are important to identify as they tend to mark regional fold-hinge areas.

Tectonostratigraphic Units

In metamorphic terranes, tectonostratigraphic units can best be described as large-scale tracts of land underlain by bedrock with similar age range, protolith paleoenvironment, and structure. Such terranes are generally bounded by ductile-fault zones (mylonites), surfaces of unconformity, or brittle faults. Unravelling the collisional plate-tectonic history of mountain belts is greatly facilitated by identifying former cratonic (ancient crustal), continental-margin, continental-slope-, and rise, deep-oceanic, and volcanic-island tectonostratigraphic units. The major distinction in unravelling complexly deformed mountain belts is to identify former shallow-water shelf deposits (originally deposited on continental crust) and to separate them from deep-water oceanic deposits (originally deposited on oceanic crust). The collective adjectives miogeosynclinal (for the shallow-water shelf deposits) and eugeosynclinal (for the deep-water oceanic deposits) have been applied to the products of these contrasting depositional realms.

Geologic Dating of Episodes of Deformation

Geologists use many methods to establish the geologic date of deformation. These include analysis of surfaces of unconformity, obtaining the dates on formations containing pebbles- or inclusions of deformed rock, relationships to associated plutons, and radiometric ages on minerals that grew as a result of deformation.

Surfaces of Unconformity

Surfaces of unconformity mark temporal gaps in the geologic record and commonly result from periods of uplift and erosion. Such uplift and erosion is commonly caused during the terminal phase of regional mountain-building episodes. As correctly interpreted by James Hutton at the now-famous surface of unconformity exposed in the cliff face of the River Jed (Figure 22), such surfaces represent mysterious intervals of geologic time where the local evidence contains no clues as to what went on! By looking elsewhere, the effects of a surface of unconformity of regional extent can be recognized and piecemeal explanations of evidence for filling in the missing interval may be found.

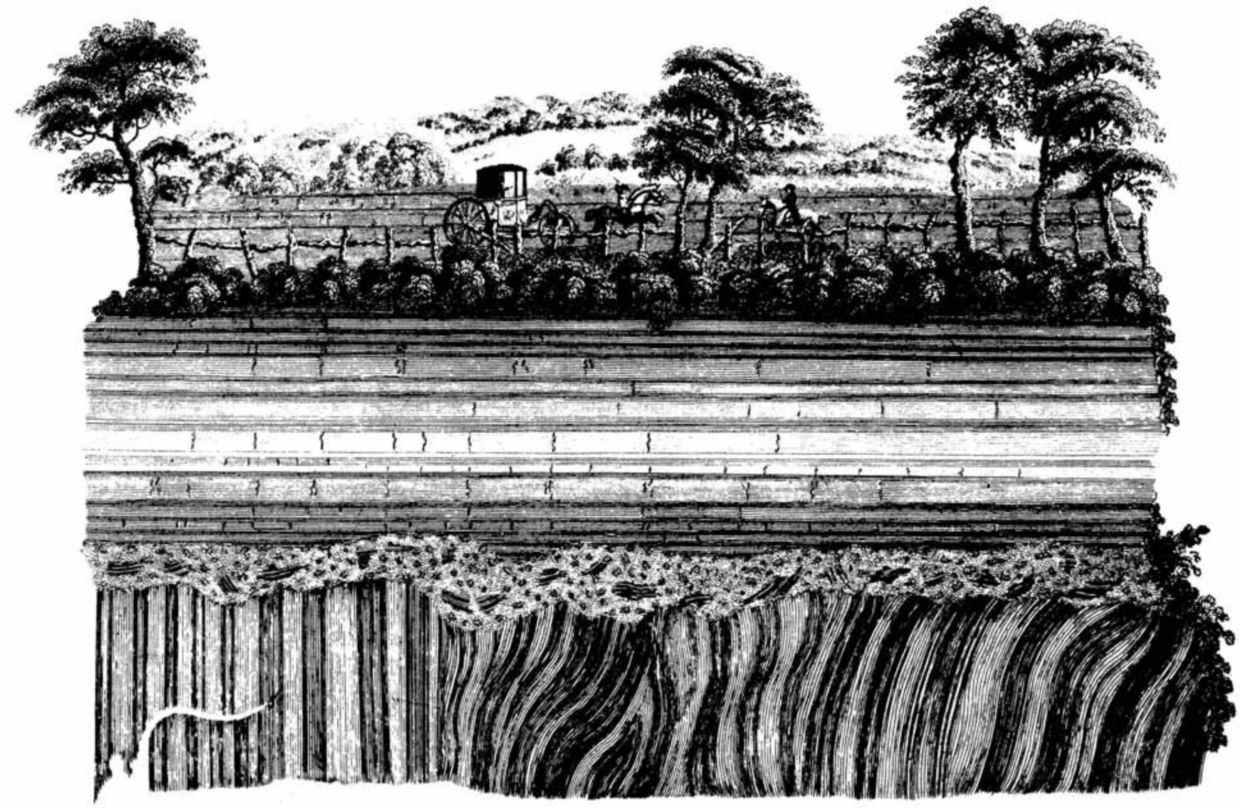


Figure 22. Unconformity with basal conglomerate along the River Jed, south of Edinburgh, Scotland. (James Hutton, "Theory (sic) of the Earth", 1795.)

Surfaces of unconformity resulting from erosion can be classified into three categories: (a) surfaces of angular unconformity, (b) surfaces of nonconformity, and (c) surfaces of disconformity (Figure 23). Along surfaces of angular unconformity (such as that James Hutton saw exposed in the banks of the River Jed), dipping strata below the surface have been truncated and thus angular discordance is present between the strata below- and above the surface of erosion. A surface of nonconformity separates sedimentary strata above from eroded igneous- or metamorphic rocks below. Surfaces of disconformity are the most-subtle variety; the separate subparallel sedimentary strata. They are commonly identified by paleontologic means, by the presence of channels cut into the underlying strata, or by clasts of the underlying strata in their basal part. The strata above a surface of unconformity may or may not include clasts of the underlying strata in the form of a coarse-textured, often bouldery basal facies.

Following the proposal made by L. L. Sloss (1963), surfaces of unconformity of regional extent within a craton are used as boundaries to define Stratigraphic Sequences.

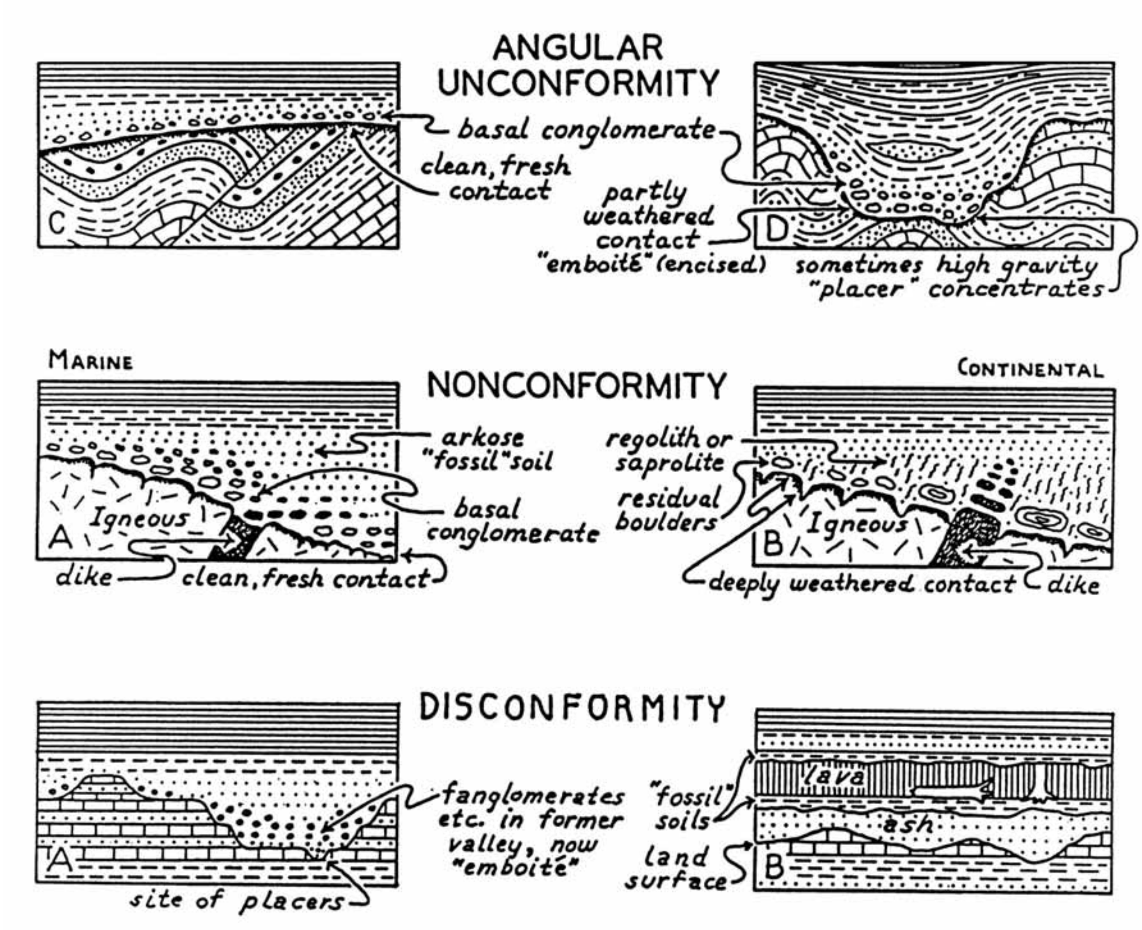


Figure 23. Varieties of geologic relationships along surfaces of unconformity, which mark gaps in the geologic record. (Drawings by Rhodes W. Fairbridge.)

Dating Formations that Contain Pebbles- or Inclusions of Deformed Rock

In certain situations, it is possible to find pebbles- or inclusions of deformed rock in another formation that can be dated. The date of the formation containing the pebble or inclusion places an upper limit on the date of deformation indicated by the pebble. For example, pieces of Martinsburg slate have been found as inclusions in the Late Ordovician igneous rocks in northwestern New Jersey. This proves that the age of the slate is pre-Late Ordovician. In other cases, pebbles of mylonite might be found in a datable conglomerate and the age of the conglomerate thus marks an upper limit on the age of the mylonite.

Relationships to Associated Plutons

Commonly orogenic episodes are accompanied by plutonic activity. Where a pluton cuts a fault, for example, the pluton clearly is younger. The date on the pluton thus sets an upper limit on the age of the fault. Plutons can be dated directly by a radiometric age on minerals, a whole-rock age, or indirectly by crosscutting relationships with formations of the country rock or by finding pebbles of the pluton in younger formations.

Radiometric Ages on Minerals that Grew as a Result of Deformation

Dating deformation by radiometric ages on minerals that grew as a result of deformation is analogous to obtaining radiometric ages on minerals from a deformation-associated pluton, but need not always involve a pluton. In some cases, micas or other minerals recrystallized as a result of deformation and the radiometric date yields the date of recrystallization. Other kinds of minerals may grow in veins whose emplacement accompanied an episode of deformation.

BEDROCK UNITS ENCOUNTERED ON TRIP ROUTE

Layers I and II: Crystalline Complex of Paleozoic and Older Rocks

As we begin our journey from the New York Academy of Sciences, a few thoughts about the rocks beneath our feet. The crystalline bedrock of New York City marks the southern terminus of an important sequence of metamorphosed Precambrian to Lower Paleozoic rocks of the Manhattan Prong (Figures 24 and 25) which widens northward into the New England Upland physiographic province of the Appalachian mountain belt. Originally, the New York City strata were, in part, deposited on the complexly deformed sequence of layered feldspathic and massive granitoid gneiss, amphibolite, and calc-silicate rocks of uncertain stratigraphy known as the Fordham and Yonkers Gneiss (Layer I). As such, the complexly deformed, Proterozoic Y and Z basement sequence (Layer I) represents the ancient continental crust of proto-North America that became a trailing edge, passive continental margin throughout the early Paleozoic Era. Interestingly, the current geologic setting of eastern North America, with deformed Paleozoic and older basement covered by Mesozoic and younger sediments, is analogous to the past (except for differences in age, paleolatitude, geothermal regime, and paleotectonics).

The Cambrian- to Ordovician bedrock units in western Connecticut and New York City (Layer II) now form a deeply eroded sequence of highly metamorphosed, folded and faulted sedimentary- and igneous rocks (Figure 26) which began life roughly 550-450 million years ago as thick accumulations of both shallow- and deep-water sediments adjacent to the Early Paleozoic shores of proto-North America (Figure 27). Layer II can be divided into two sub-layers, IIA and IIB.

The older of these, IIA, represents the ancient passive-margin sequence of the proto-Atlantic (Iapetus) ocean. These rocks can be subdivided into two facies that differ in their original geographic positions with respect to the shoreline and shelf. A near-shore facies [Layer IIA(W)] was deposited in shallow water and is now represented by the Cambrian Lowerre Quartzite and Cambro-Ordovician Inwood Marble in New York City and as the Cheshire Quartzite and Stockbridge Marble in western Connecticut and Massachusetts. These strata began life as sandy and limey sediments in an environment not significantly different from the present-day Bahama Banks.

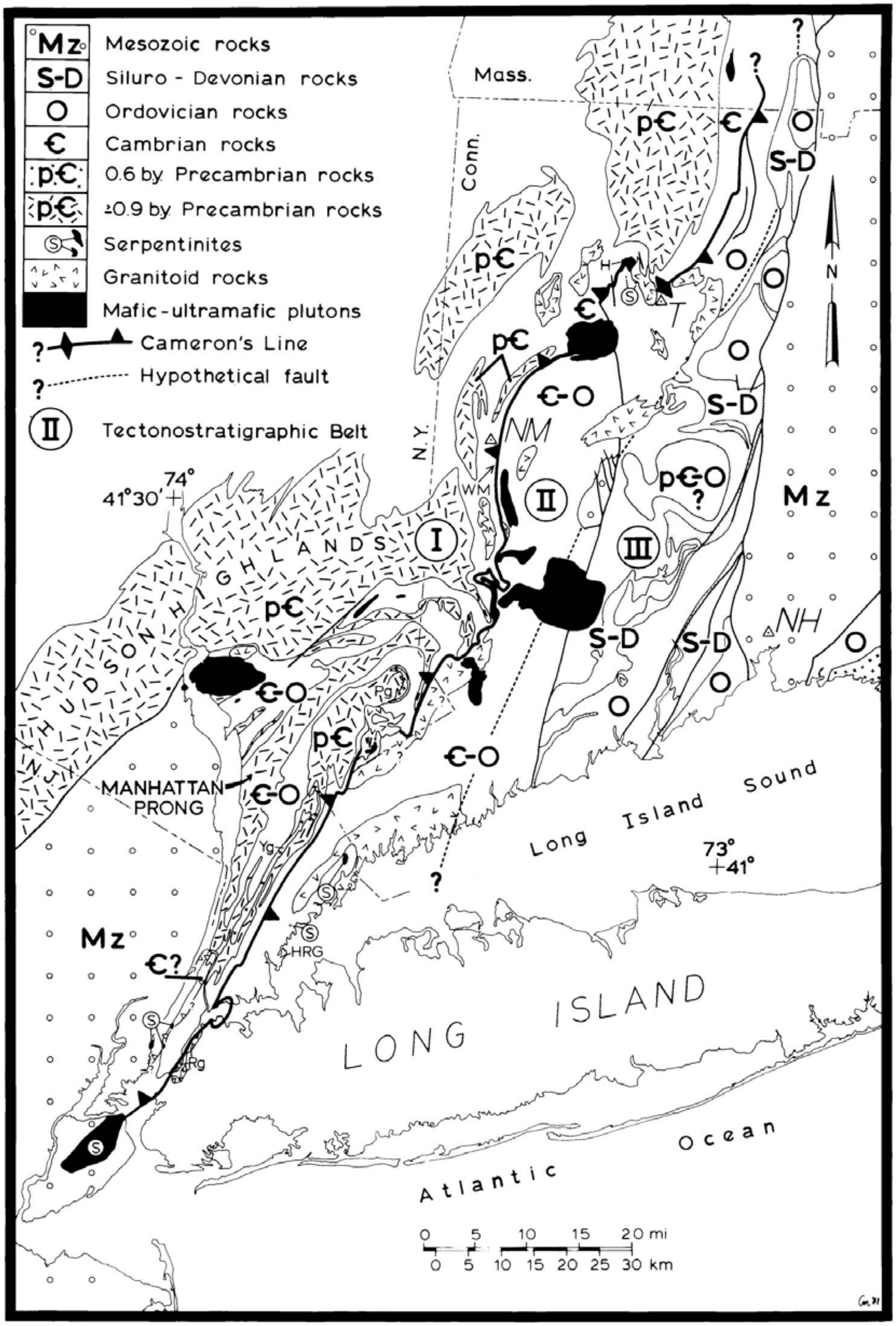


Figure 24. Geotectonic map of western Connecticut and southeastern New York. (Charles Merguerian, 1983, fig. 1, p. 342.)

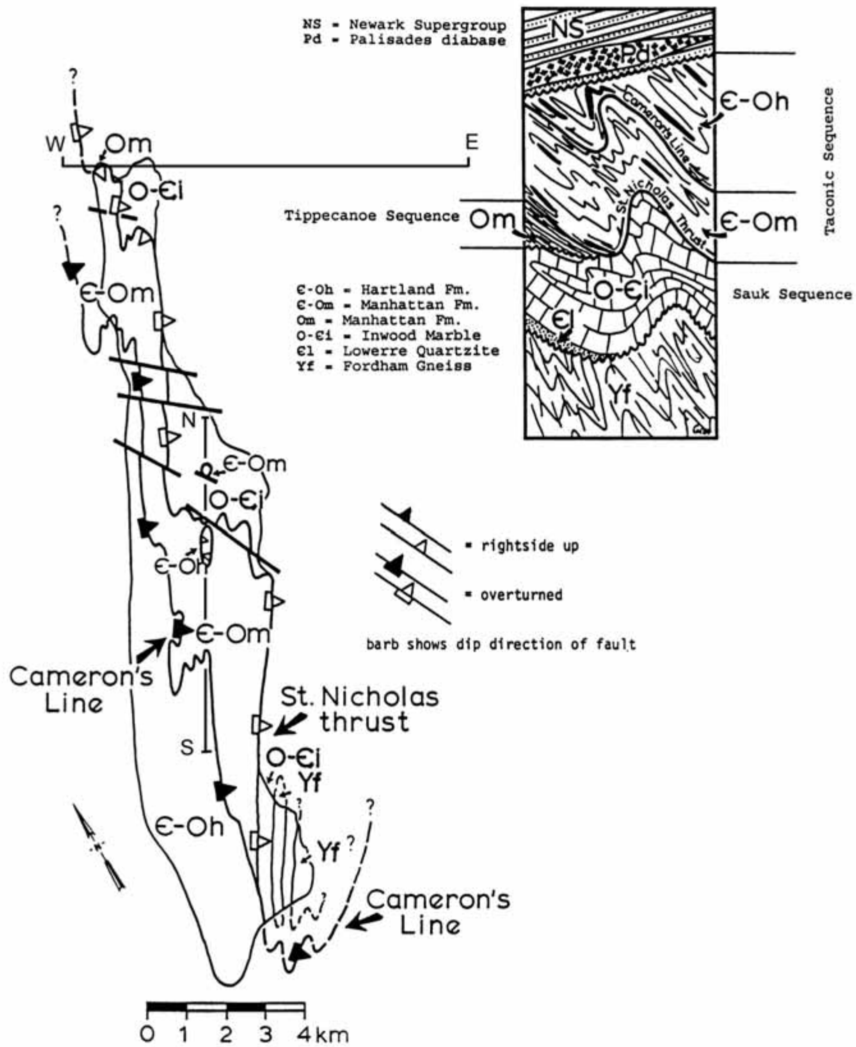


Figure 25. Geologic map and -section of Manhattan Island showing a new interpretation of the stratigraphy- and structure of the Manhattan Schist. (Drawn and mapped by Charles Merguerian.)

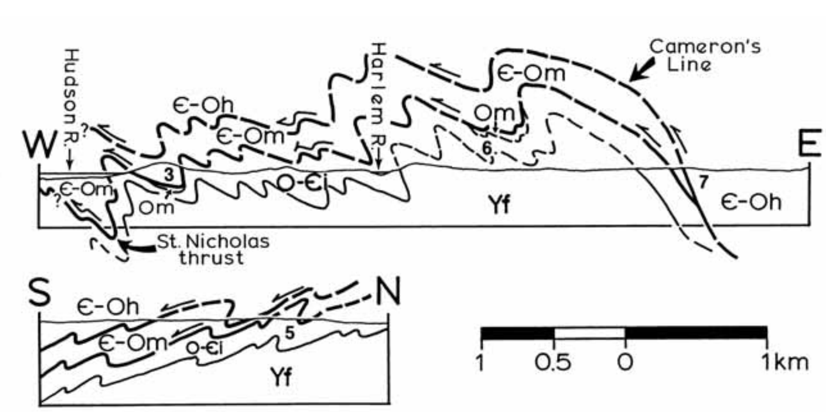


Figure 26. Folded overthrusts, Manhattan Island. Symbols defined on Figure 25. (Charles Merguerian, unpublished data.)

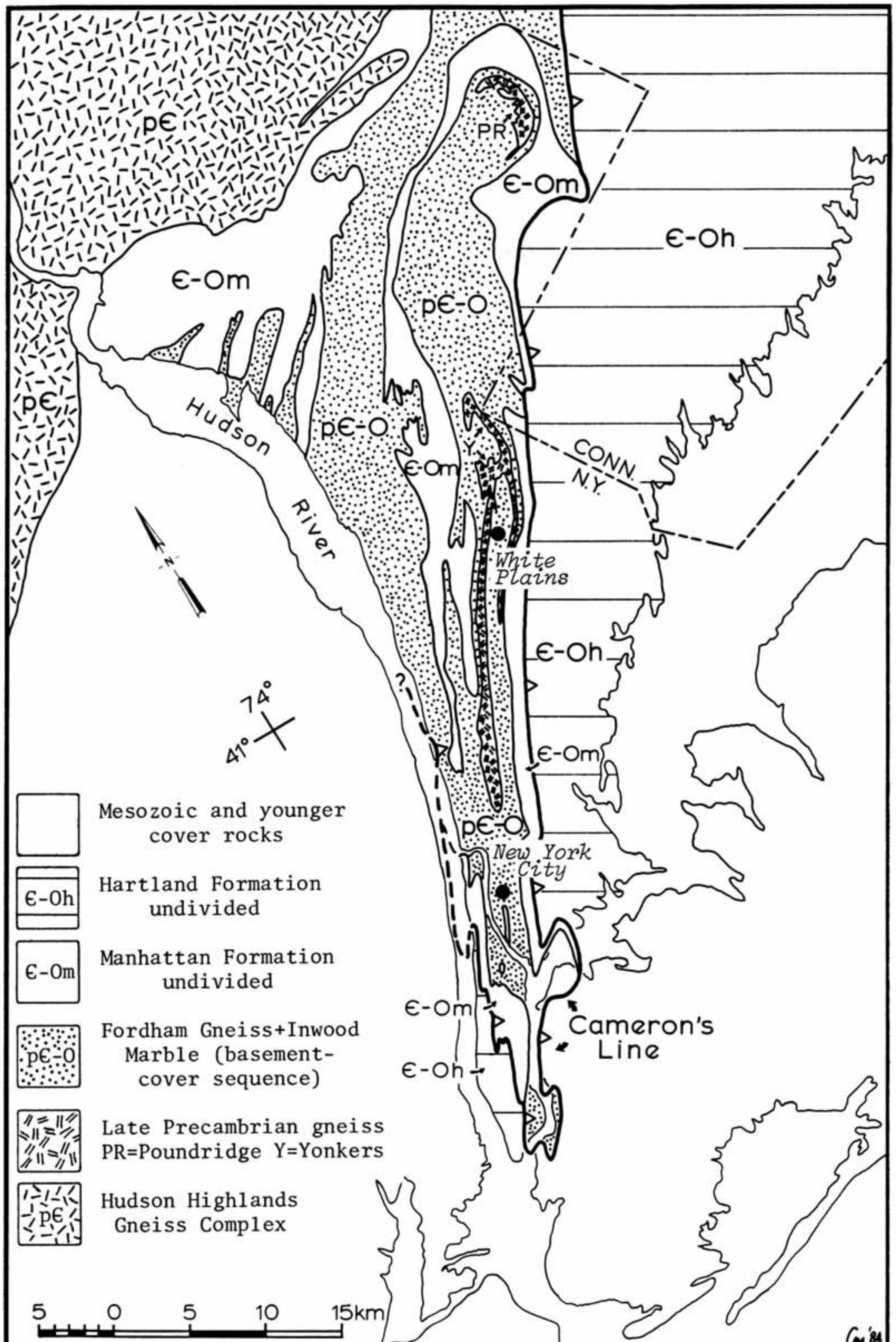


Figure 27. Simplified geologic map of Manhattan Prong showing the distribution of metamorphic rocks (Layers I and II) ranging from Proterozoic Y through Early Paleozoic in age. Most faults and intrusive rocks have been omitted. (D. G. Mose and Charles Merguerian, 1985, fig. 1, p. 21.)

Farther offshore, fine-grained terrigenous time-stratigraphic equivalents of the shallow water strata (shelf sequence) were evidently deposited under deep water on oceanic crust [Layer IIA(E)]. This sequence is also of Cambrian to Ordovician age and is known as the Taconic sequence in upstate New York, as units C-Ot and C-Oh of the Manhattan Schist(s) (Figure 25), and is described below as the Waramaug and Hartland formations, respectively, of western Connecticut.

Layer IIB is younger strata that rests depositionally above the western shallow water platform [Layer IIA(W)]. In eastern New York State, these rocks are mapped as the Waloomsac Schist, and Manhattan Formation. In New York City, it is the Manhattan Schist unit Om which, according to CM, is demonstrably interlayered with the Inwood Marble and contains thin layers of calcite marble (Balmville equivalent) at its base at Inwood Hill Park in Manhattan (Merguerian and Sanders 1988, 1991, 1991, 1993, 1993; NYAS On-The-Rocks Trips #3 16 21, 26 28). This field evidence is used to indicate that unit Om of the Manhattan Schist is in place where found and is therefore younger, or the same age, as Manhattan units C-Ot and C-Oh.

The Geology- and Tectonics of Western Connecticut

The crystalline terrane of western Connecticut consists of a diverse assemblage of Proterozoic to lower Paleozoic metasedimentary and metaigneous rocks of the Hartland Formation which can be traced from New York City (Layer IIA(E)) northward into the Connecticut Valley-Gaspe, synclinorium (a large-scale downfold or "syncline" that effects a broad portion of the Earth's crust) (Figure 29). In addition, rocks of Layer IIA(W) crop out in westernmost Connecticut and are continuous with lower Paleozoic rocks of southeastern New York. Separated by Cameron's Line, a major ductile shear zone in the New England Appalachians, these two major geological terranes [Layers IIA(W) and IIA(E)] dominate the geologic framework of western Connecticut (Figure 30).

J. G. Percival (Figure 31) was a melancholy naturalist, poet, U. S. Army surgeon, botanist, and a cunning linguist who collaborated with Webster on the first American dictionary. Clearly a character worthy of further discussion, the interested reader should consult Bell (1985) for more details. Of interest here, however, was the production of the first state geologic map of Connecticut in 1842 (Figure 32) by Percival who spent nearly seven years traversing Connecticut in a one-horse wagon, then by foot. Initially in 1835, Percival was accompanied by C. U. Shepard. By 1837, Shepard had published his comprehensive report on the economic mineral deposits of the state. Percival continued his work alone and eventually traversed the entire state along parallel East-West lines two miles apart. Percival's map (1842) is an excellent- and thorough document with a long written report (now a collector's item!) that has proven more correct than modern mapping in certain areas.

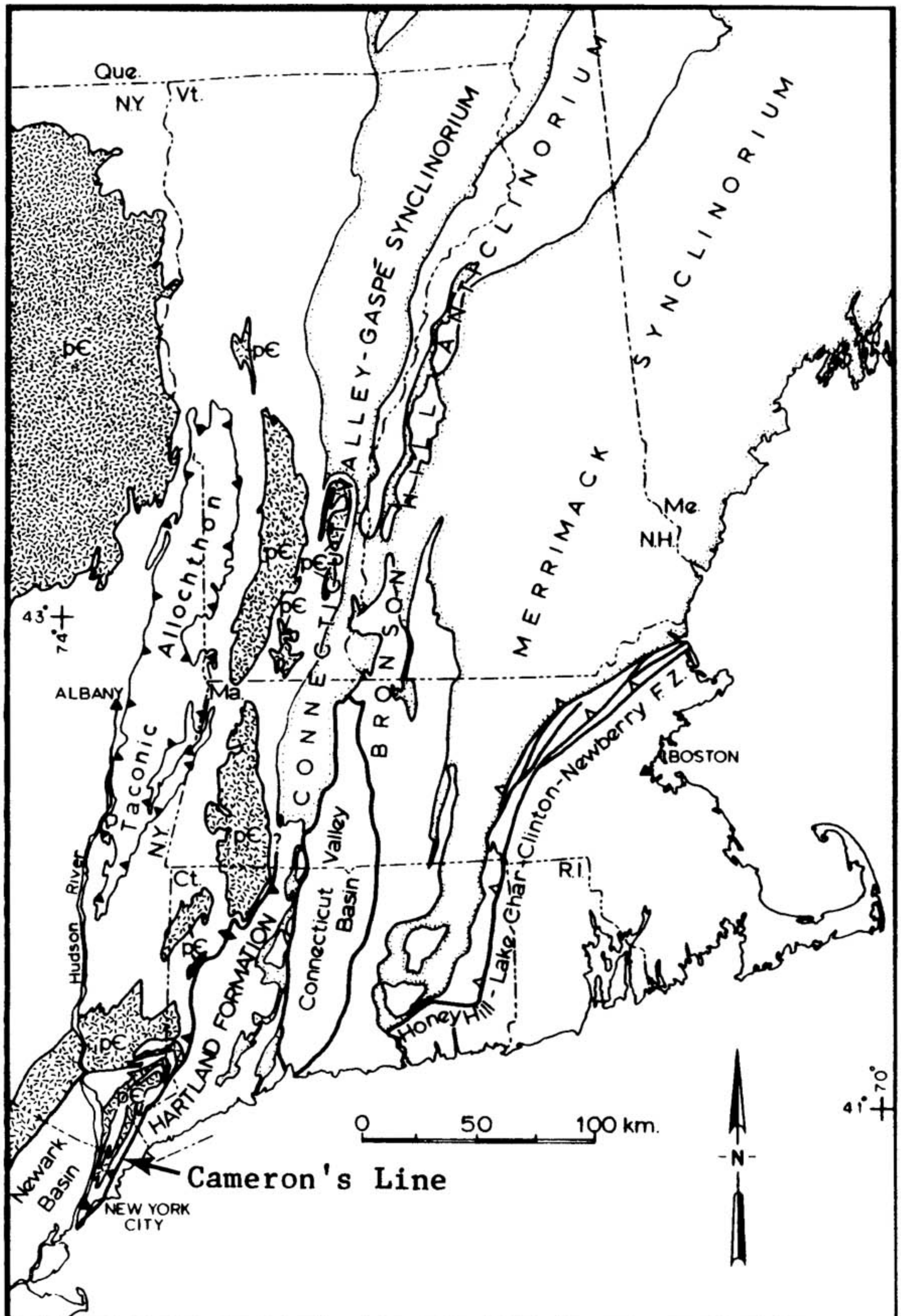


Figure 29. Tectonic sketch map of southern New England showing the major geotectonic provinces. (Charles Merguerian, 1983, fig. 1, p. 343.)

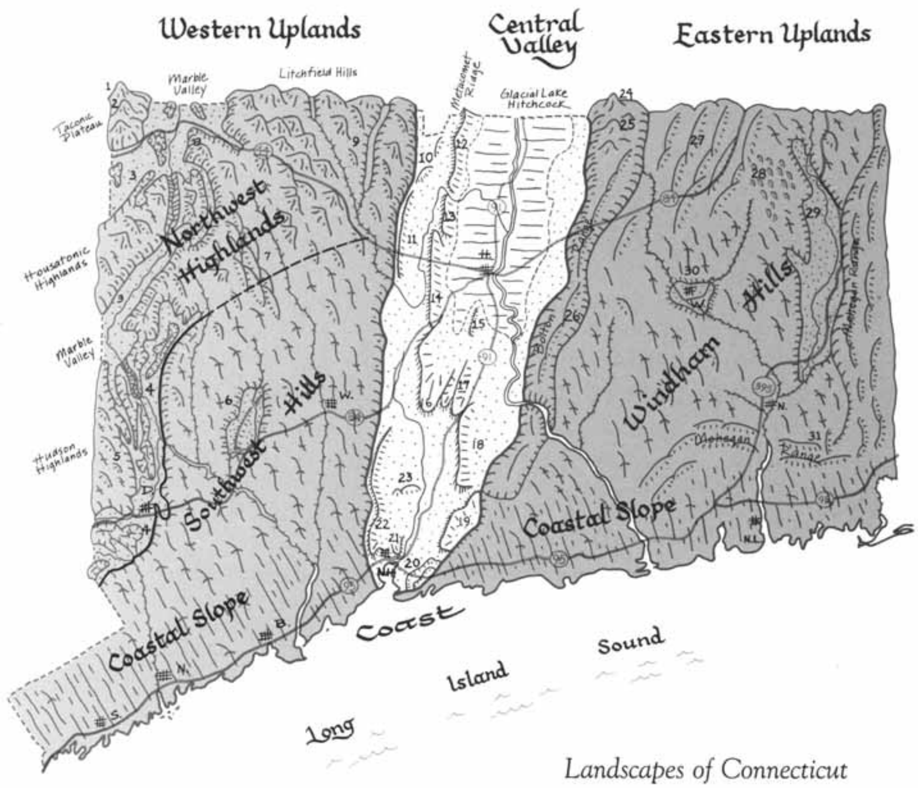


Figure 30. Physiographic map of Connecticut showing the various geological provinces. (Bell, 1985, figure on p. 7.)

Figure 31. Sketch showing the cunning James Gates Percival. (Bell, 1985, figure on p. 135.)

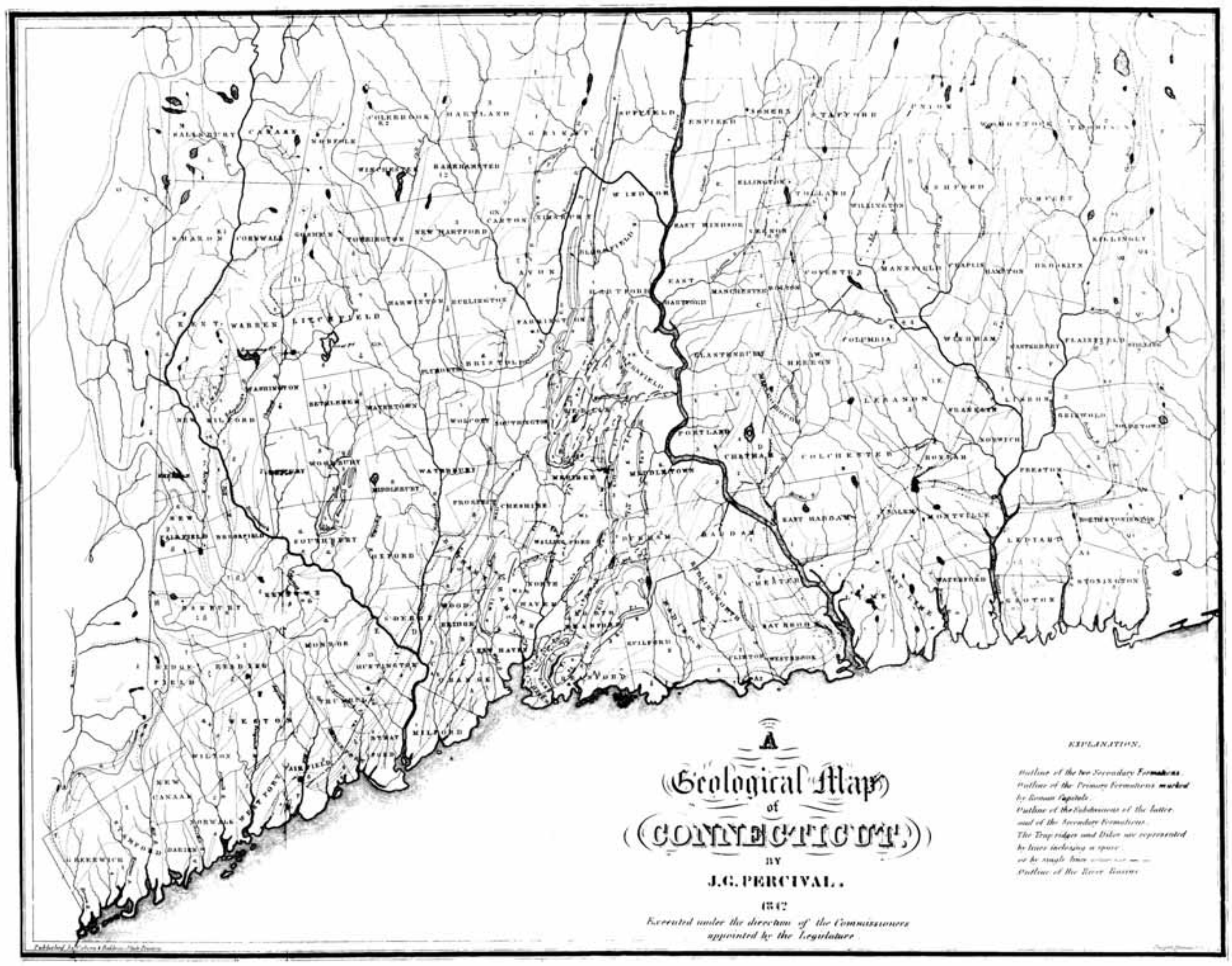


Figure 32. J. G. Percival's geological map of Connecticut (Percival, 1842.)

The Hartland Formation (Cameron, 1951, Gates, 1951, 1952; Merguerian, 1977) consists of aluminous metasedimentary and interlayered metavolcanic rocks. These rocks are bounded on the west by Cameron's Line and to the east, are overlain by metamorphosed rocks of probable Silurian and Devonian age (Hatch and Stanley, 1973) (Figure 33). The Hartland Formation (Layer IIA(E)) constitutes the bulk of the highlands of western Connecticut (See Figure 29.) and is a metamorphosed sequence of eugeosynclinal rocks (formerly deposited in deep water on oceanic crust). Occurring to the west of Cameron's Line is a sequence of massive gneissic rocks known as the Waramaug Formation (Cambrian and ?Ordovician) which is correlative to the north with the Cambrian Hoosac Schist and to the south with unit C-Ot of New York City (Merguerian, 1983). This sequence is interpreted as a continental slope/rise deposit that was situated between the depositional sites of Layers IIA (W) and (E). Thus, on either side of Cameron's Line, notably disparate sequences of equivalent age occur with lower-plate continental-shelf, -slope, and -rise rocks and upper-plate oceanic rocks juxtaposed along a major zone of mylonite (ductile shear zone).

Cameron's Line

According to Eugene Cameron (of Cameron's-Line fame) in a confidential personal communication with CM, the geologic relationship of Cameron's Line was first noted by William Agar who shared same with E. Cameron. According to EC - "I don't know why they called it Cameron's Line, it should have been called Agar's Line!". In any case, Cameron's Line delimits the easternmost exposures of autochthonous Proterozoic Y and Z gneiss and overlying lower Paleozoic quartzite and marble (shallow-water sedimentary strata [Layer IIA(W)] formed originally on continental crust of proto-North America. Together, Layers I and IIA(W) represent the deformed North American craton and overlying shelf deposits.

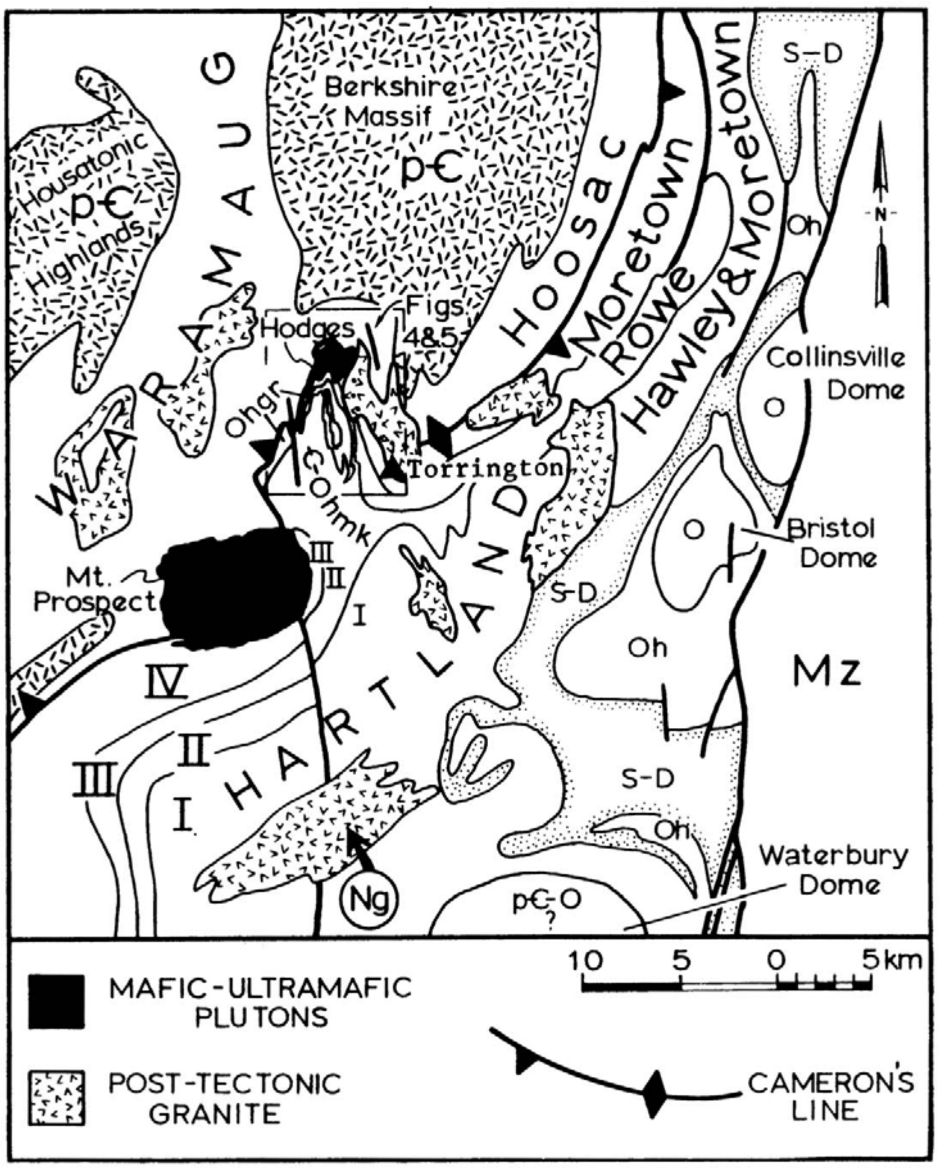


Figure 33. Geological sketch map showing the lithostratigraphic correlations between rocks of the West Torrington area and adjacent regions. Oh = Hawley Fm., Om = Moretown Fm., Ng = Nonewaug granite. (After Hatch and Stanley, 1973; Gates, 1967; Merguerian, 1983; Rodgers and others, 1959; Rodgers, 1985.)

In western Connecticut, the Hartland Formation or Complex of Merguerian (1983) is interpreted as an internally sheared imbricate thrust package that marks the former site of a deep-seated accretionary complex or subduction zone. It consists of a thick sequence of interlayered muscovite schist, micaceous gneiss and granofels, amphibolite, and minor amounts of calc-silicate rock, serpentinite, and manganiferous- to ferruginous garnet-quartz granofels (coticule) (Merguerian, 1981). Hartland rocks are correlative with metamorphosed eugeosynclinal (deep-water deposition) Cambrian to Ordovician rocks found along strike in New England. (See Figure 24.) Note the northwest (eastern New York state) to southeast (central Connecticut) stratigraphic variation from shallow-water shelf to deep-water volcanogenic interpreted protoliths in Figure 34.

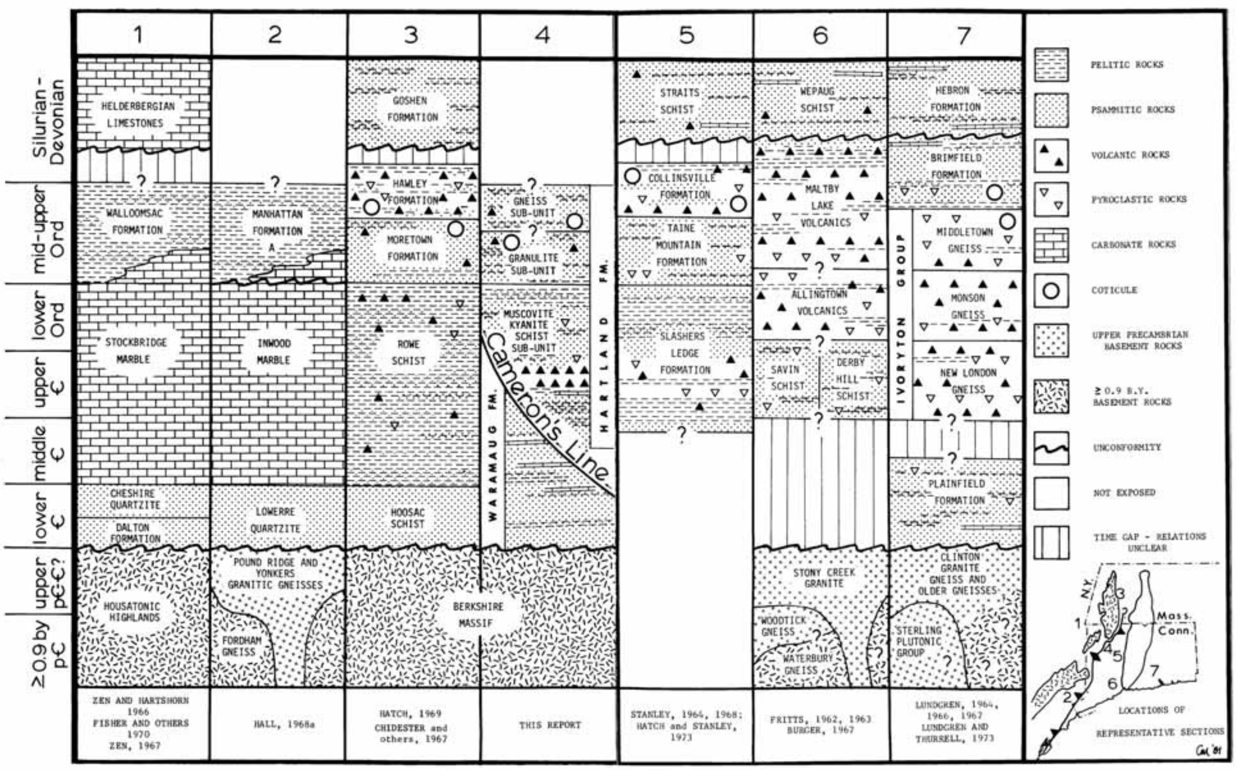
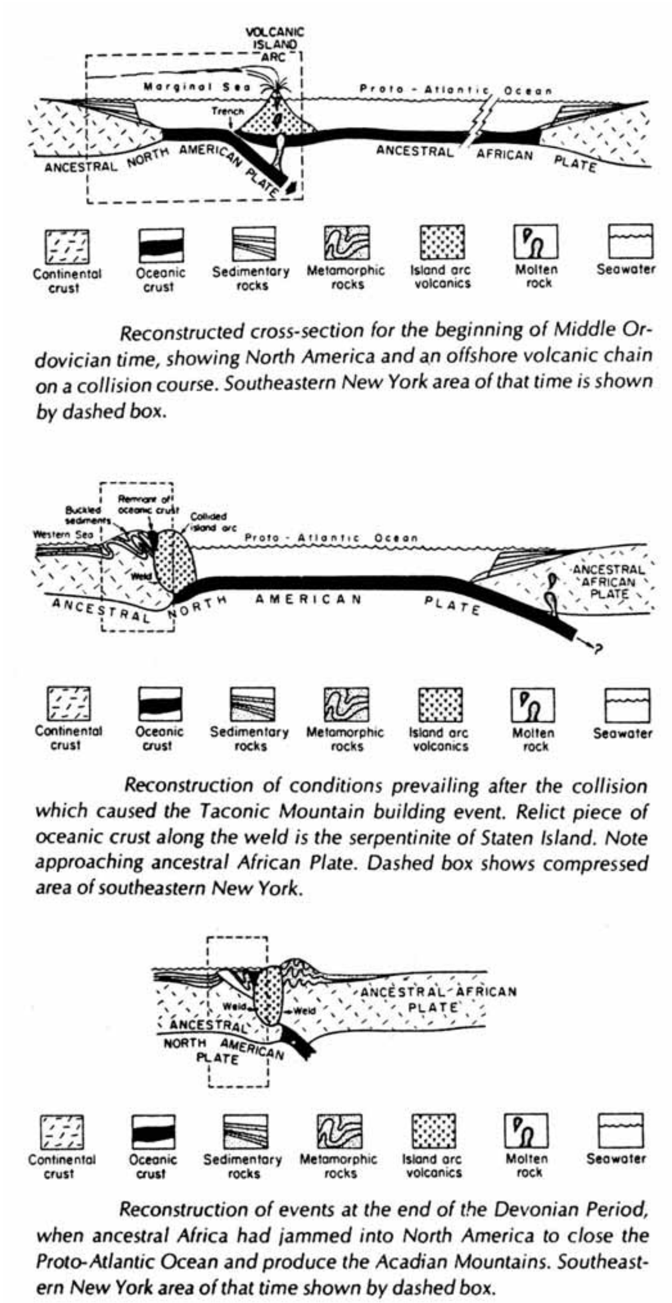


Figure 34. Stratigraphic correlation chart for southern New England showing the interpreted protoliths of the lithologies depicted. Mixed symbols indicate relative, non-quantitative abundances of various pre-metamorphic lithologies. (Charles Merguerian, 1985, fig. 3, p. 414.)

In western Connecticut, numerous lower Paleozoic calc-alkaline plutons are present. Near West Torrington, Connecticut, the Hodges mafic-ultramafic complex and the Tyler Lake Granite were sequentially intruded across Cameron's Line (Merguerian, 1977). Because of their formerly elongate shapes and because the regional metamorphic fabrics related to the development of Cameron's Line in both the bounding Waramaug and Hartland formations display contact metamorphism, these plutons are interpreted as syn-orogenic. The recognition of significant medial Ordovician plutonism across Cameron's Line (Mose, 1982; Mose and Nagel, 1982; Merguerian and others, 1984; Amenta and Mose, 1985) establishes a Taconian or possibly older age for the formation of Cameron's Line and the syntectonic development of regional metamorphic fabrics in western Connecticut (Merguerian, 1985). Judging by metamorphic minerals in the regional fabric, Layers IIA(W) and IIA(E) were juxtaposed at depths of roughly 20-to 25 km along Cameron's Line during early Paleozoic times. The force behind such deep-seated deformation presumably resulted from a collision between a volcanic-arc terrane and the passive continental margin of North America (Figure 35). At present, the arc terrane is exposed in the Bronson Hill Anticlinorium and its extension southward into cental Connecticut. (See Figures 24 and 29.)



Reconstructed cross-section for the beginning of Middle Ordovician time, showing North America and an offshore volcanic chain on a collision course. Southeastern New York area of that time is shown by dashed box.

Reconstruction of conditions prevailing after the collision which caused the Taconic Mountain building event. Relict piece of oceanic crust along the weld is the serpentinite of Staten Island. Note approaching ancestral African Plate. Dashed box shows compressed area of southeastern New York.

Reconstruction of events at the end of the Devonian Period, when ancestral Africa had jammed into North America to close the Proto-Atlantic Ocean and produce the Acadian Mountains. Southeastern New York area of that time shown by dashed box.

Figure 35. Reconstructed cross section for the beginning of mid-Ordovician time, showing North America and an offshore volcanic chain on a collision course. (Yngvar Isachsen, 1980, fig. p. .)

In summary, during a series of Paleozoic mountain-building episodes (the Taconic-, Acadian-, and Appalachian orogenies), the sedimentary protoliths [Layers IIA(W and E) and IIB] of the New England Appalachians were sheared, folded, and metamorphosed during a collision between an exotic volcanic-island chain and the passive continental margin of proto North America. (See Figure 35.) Much of the bedrock in western Connecticut is therefore interpreted to be allochthonous (a fancy term intended to confuse the layman which simply means transported from somewhere else or not deposited where currently found!). In this model, Cameron's Line marks a fundamental plate-tectonic boundary (suture) between continental [Layer IIA(W)] and oceanic realms [Layer IIA(E)] and marks the root zone for much of the Taconic sequence in eastern New York state. The Hartland Formation (Complex) marks the deeply eroded roots of an uplifted accretionary complex marking the former trench into which the raw edge of North America was subducted. In a later section we discuss specifics of the geology of the West Torrington quadrangle.

Layer V: Hartford Basin-filling Strata

Although we shall not be making any stops at localities to examine the Newark Supergroup strata of Connecticut, we mention them for two reasons: (1) some mineral deposits in the pre-Newark rocks may have resulted from the Newark igneous activity and associated hot waters; and (2) the drainage history of western Connecticut includes activities that date back as far as the Late Triassic and Early Jurassic Periods when the Newark Supergroup was being deposited.

In Connecticut, the Newark-age strata are exposed in three areas (See Figures 24 and 29.): (a) the Hartford basin (forms the main region of central Connecticut, extending northward from New Haven Harbor into central Massachusetts); and two smaller areas underlain by strata thought to have been continuous formerly with those of the Hartford basin: (b) the Pomperaug Valley belt; and (c) the Cherry Brook valley belt, Canton Center. In all of these three areas, the regional dip of the strata is toward the east. Because all the strata were initially deposited in horizontal positions, their modern-day dips must be ascribed to the effects of tectonic uplift, with the axis of this feature located to the west of South Britain, Connecticut, where east-dipping Newark-age strata are exposed in the Pomperaug River.

Layer VII: Quaternary Sediments: Glacial Deposits

The glacial deposits of Connecticut include the work of at least two ice sheets that flowed from contrasting directions: (i) from the NNE to the SSW, and (ii) from the NW to the SE, the same two directions discussed in the New York City region (On-The-Rocks Trip 3, Manhattan; and Trip 4, Staten Island). These two directions were noticed under the heading of "Diluvial Scratches" by the great genius of Connecticut geology, J. G. Percival (1842). Two tills having these contrasting directions of flow have been described from central Connecticut: the older, Lake Chamberlain till with flow NW-SE; and the younger, Hamden till, with flow NNE-SSW (Flint, 1961). In addition, evidence for significant end moraines has been described along the

Long Island Sound coast of Connecticut (Flint and Gebert, 1974; 1976). We think that these end moraines are the terminal moraines for the Woodfordian glacier (depositor of the Hamden Till).

Drumlins having long axes oriented NW-SE are abundant in the Newtown-Roxbury area. In this same area, one can also find drumlins having long axes oriented NNE-SSW. Study of the U. S. Geological Survey topographic maps of the West Torrington quadrangle will provide numerous examples of each. The co- existence of such drumlins inferred to have been the work of two different ice sheets is hard for some glacial geologists to accept. They argue that the younger glacier should have wiped out all traces of any older glacier, especially one having a contrasting flow direction (Flint, 1943, 1951). We offer no simple explanation for how this can have happened; we merely cite the topographic maps as proof that it did happen.

Drainage History

The drainage history of Connecticut has resulted from several complex episodes of uplift and valley erosion, some of which antedated the arrival of the first Pleistocene continental glacier and some of which were related to periods of drainage re-arrangement that accompanied the melting of the glaciers.

As in the New York City region, the oldest surviving evidence for ancient drainage is contained in the Newark Supergroup. The provenance data (Krynine, 1950) and the cross strata in the Newark strata of the Southbury outlier show that the Late Triassic-Early Jurassic drainage flowed from E to W, from E of the basin-marginal fault along the E side of the Hartford basin all the way to South Britain (Hubert student on Pomperaug; also Don Schutz, Yale undergrad res. project).

The next possible time for drainage change was in the medial Jurassic Period in connection with the uplift and breakup of the Newark-Hartford basin complex. During this time, the regional arch formed with its axis lying W of the east-dipping strata at South Britain and E of the W-dipping strata of the Newark basin. This period of erosion culminated in the Cretaceous-age Fall Zone peneplain, whose traces in Connecticut have been studied by Flint (1963). A valley that may have formed at this time lies buried beneath New Haven harbor (Sanders, 1965, 1994; Haeni and Sanders, 1974). The age of this valley is not known, but one possible interpretation is that it extends to the WSW from New Haven harbor and disappears by going underneath the Upper Cretaceous strata underlying western Long Island. If so, it would be the same age as the strike valley at the base of the Newark Supergroup that passes beneath the Upper Cretaceous on western Staten Island (Lovegreen, 1974 ms.).

Because the Upper Cretaceous coastal-plain strata probably covered parts of all of Connecticut and Massachusetts (at least all of the Hartford-Deerfield basin-filling strata to keep them out of circulation), we can infer that no drainage systems formed during the time (Late Cretaceous to the end of the Miocene) while western Connecticut was subsiding and receiving sediment during its most-recent passive-margin phase. After this phase, the first time when erosion could have started again is late in the Miocene Epoch (or early in the Pliocene Epoch), when New England was regionally elevated, all traces of any former updip extension of the coastal-plain strata were removed, and the depression now occupied by Long Island Sound was eroded. This depression was in existence when the Pleistocene glaciers arrived. This Pliocene time of elevation and erosion is probably as far back as it is possible to trace the history of most modern valleys.

Geophysical studies of the subbottom sediments in western Long Island Sound have shown that a U-shaped valley, trending more or less E-W, and with its thalweg extending down to about 600 feet below modern sea level has been cut into the coastal-plain strata (Grim, Drake, and Heirtzler, 1970). JES thinks this U-shaped valley was carved by one of the glaciers that flowed from the NW to the SE and was diverted to nearly W-E flow by the escarpment facing the inner cuesta lowland at the eroded edge of the coastal-plain strata.

The relationships between Pleistocene glaciers and drainage were complex; just how complex one is prepared to accept depends on how many glacial episodes may have affected the region and how much rearranging accompanied each glacial advance and -retreat. This statement may seem self evident, but it is included because most of the students of the drainage history of Connecticut have thought in terms of a single Pleistocene glaciation and some of them have been persuaded that the effects of this glacier on the landscape and drainage were minimal.

A continental glacier would tend to deepen any valleys trending parallel to the glacier's flow direction and to fill any valleys trending at a high angle to this direction. Each glacial advance would terminate all previous drainage and each glacial retreat would enable new drainage networks to form. JES thinks that some of the new drainage may have been initiated on the top of the ice itself, so that anomalous cross-axial drainage routes, such as that of the Housatonic River across the Housatonic Highlands in northwestern Connecticut, for example, conceivably could have been established by superposition from the glacier. This is a concept that has not been considered by previous students of drainage history. Many such students have visualized the possibility that modern rivers may have attained their present locations as a result of superposition, but they supposed that the only way this could happen would be from the now-eroded former landward extensions of the coastal-plain strata.

The uniformity of flow directions associated with the glaciers that came from the NW and traveled SE, a rectilinear pattern found on even the highest ridges in today's landscape, suggests to JES that these features were eroded by a thick glacier whose flow direction was determined by the slope on the top of a thick ice sheet. Any rivers that began on the top surface of such an ice sheet would have been afforded numerous possibilities for superposition. How many times such thick glaciers overspread southern New England is not known, but we think the minimum number is 2 and have openly suggested (Sanders and Merguerian, 1991a, 1991b, 1992, 1994b, 1995a, b; Sanders, Merguerian, and Okulewicz, 1995a,b) that four glaciers may have affected the New York City area.

During the melting- and retreat of the latest glacier, large lakes formed in the important lowlands--Long Island Sound and the central lowlands in Connecticut and Massachusetts (Ashley, 1972). In these lakes were deposited varved sediments that Antevs (1922) used to assemble a chronology of glacial retreat (Schove, 1987).

So much for the geologic background. We now turn to the specifics of today's trip, starting with the objectives.

OBJECTIVES

1) To examine the Hartland and Waramaug formations of western Connecticut.

2) To study mafic- and ultramafic rocks of the Hodges Complex and the younger Tyler Lake Granite.

3) To establish the contact relationships of these plutons to each other and to Cameron's Line.

4) To identify slivers of ophiolite and to compare them to the mafic- to ultramafic igneous rocks of the plutons.

5) To illustrate methods of analyzing geologic structures in rock that have been complexly deformed.

6) To locate and discuss glacial features.

7) Not to get bitten by ticks or mosquitos, and,

8) To visit all of our field trip stops (Fat Chance!).

LIST OF LOCALITIES TO BE VISITED

Stops 1 through 8, in western Connecticut are shown on the topographic map, Figure 36.

Stop 1: Commuter parking lot at end of exit ramp for Route 118 off Conn. Route 8, Hartland (upper member) granofels, schist, and amphibolite.
Stop 2: Highland Avenue, Torrington, 0.5 mi from radio towers, Soapstone Hill, Hartland Formation (lower member), amphibolite and subsidiary D2 shear zone.
Stop 2a: (Optional): Soapstone quarry.
Stop 3: Woods north of Soapstone Hill Road: Hartland Formation (lower member) muscovite-kyanite-staurolite schist.
Stop 4: Dirt road near Patterson Pond: Cameron's Line and dismembered ophiolite.

LUNCH STOP: South of dam, Brass Mill Dam Road: Waramaug Formation.

Stop 5: John Brown Road; Waramaug Formation and contact with Hodges diorite sill.
Stop 6: (Optional): Weed Road; 1320-ft hill: Hodges Complex, mafic- and ultramafic rocks.
Stop 7: Wright Road, ridge near first barn on left: Hodges Complex, mafic- and ultramafic rocks.
Stop 8: Ducci Electrical Contracting Co. parking lot, off Lovers Lane: Tyler Lake Granite.

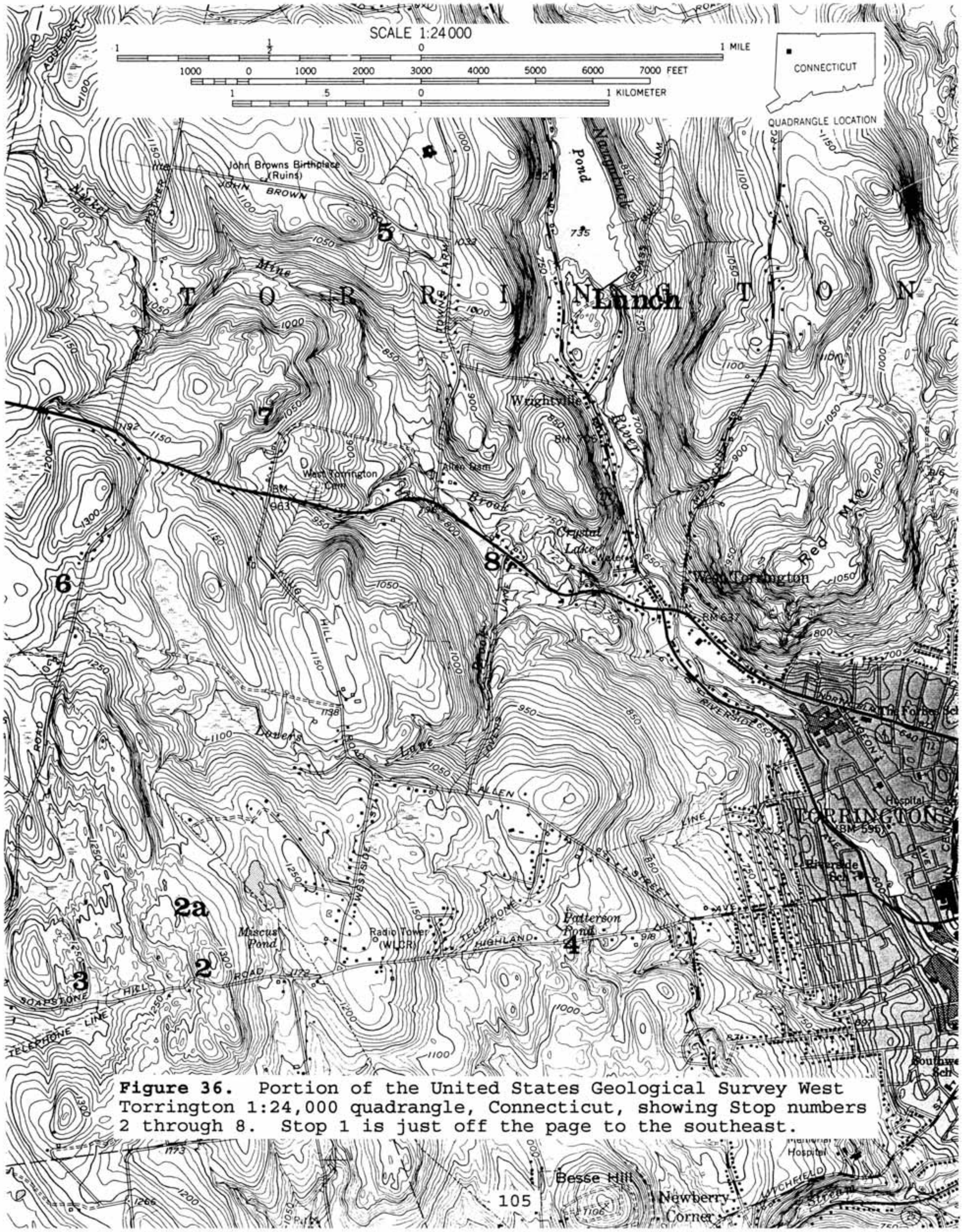


Figure 36. Portion of the United States Geological Survey West Torrington 1:24,000 quadrangle, Connecticut, showing Stop numbers 2 through 8. Stop 1 is just off the page to the southeast.

GEOLOGY OF THE WEST TORRINGTON QUADRANGLE

Modern stratigraphic- and structural studies of lower Paleozoic metamorphic rocks in the New England Appalachians have defined distinctions between miogeosynclinal, transitional, and eugeosynclinal tectonostratigraphic units (Hatch and others, 1968; Cady, 1969; Hall, 1976, 1980; and Robinson and Hall, 1980). Many workers in western Connecticut have noted the abundance of Proterozoic Y gneiss and autochthonous lower Paleozoic migeosynclinal cover rocks to the west of Cameron's Line and the abrupt eugeosynclinal characteristics of the Hartland Formation to the east (Agar, 1927; Cameron, 1951; Rodgers, 1985; Rodgers and others, 1959; Hatch and Stanley, 1973; and Merguerian, 1983, 1985, 1987).

Recent interpretations suggest that Cameron's Line is an important deep-seated ductile fault in the mid-Ordovician Taconic suture zone separating slope/rise rocks (Waramaug Formation) from essentially coeval deep-water eugeosynclinal rocks (Hartland Formation). Included in the western terrane (Belt I in Figure 24) are metamorphosed Cambrian- to Ordovician allochthonous rocks such as the Waramaug Formation and the Hoosac Schist, deposited transitionally between shallow- and deep water realms (discussion below).

The Hartland Formation composes the eastern terrane (Belt II in Figure 24) and occurs to the east of Cameron's Line. The Hartland is a metamorphosed sequence of eugeoclinal rocks formerly deposited on oceanic crust. Judging by metamorphic minerals, the western- and eastern terranes were juxtaposed at depths of ca. 20 km along Cameron's Line during Early Paleozoic time. The force behind such deep-seated deformation presumably was associated with a collision between lithosphere plates. One possible combination is between a volcanic-arc terrane and the passive margin of North America. (See Figure 35.) At present, the inferred remnants of the arc terrane are exposed in the Bronson Hill Anticlinorium and its extension southward into central Connecticut. (See Figure 29.) Note the northwest-to-southeast stratigraphic variation from miogeoclinal- to eugeosynclinal rocks from eastern New York State to central Connecticut. (See Figure 34.)

Numerous Lower Paleozoic calc-alkaline plutons are present both in the western- and eastern terranes in southern New England. Near West Torrington, the Hodges mafic-ultramafic complex and then the Tyler Lake Granite were sequentially intruded across Cameron's Line (Merguerian, 1977). These plutons are considered to be late synorogenic because: (1) their former shapes were elongate; (2) because they are surrounded by contact-metamorphic aureoles that affect regional-metamorphic fabrics related to the development of Cameron's Line in both the bounding Waramaug and Hartland formations. The recognition that midlle Ordovician plutons cut across Cameron's Line (Mose, 1982; Mose and Nagel, 1982; Merguerian and others, 1984; Amenta and Mose, 1985) establishes a Taconian- or possibly older age for the formation of Cameron's Line and the syntectonic development of regional metamorphic fabrics in western Connecticut.

Stratigraphy

The Waramaug Formation of Gates (1952) forms a belt up to 10 km wide from Torrington southward to New Milford, Connecticut (Figure 13) where Clarke (1958) correlated the Waramaug and Manhattan formations. In the vicinity of West Torrington, the Waramaug (p€-Owg) crops out west, north, and northeast of the Hodges Complex (Figure 37) and consist of a heterogeneous assemblage of -rusty-, -gray-, and locally maroon-weathering gneiss, mica schist, and granofels with subordinate amphibolite gneiss, amphibolite, and calc-silicate rocks (Stops 4, lunch stop and 5). Outcrops are massive and indistinctly layered with a nubby weathered surface resulting from resistant clusters of quartz and aluminosilicate minerals that may have been segregated from their surroundings during metamorphism.

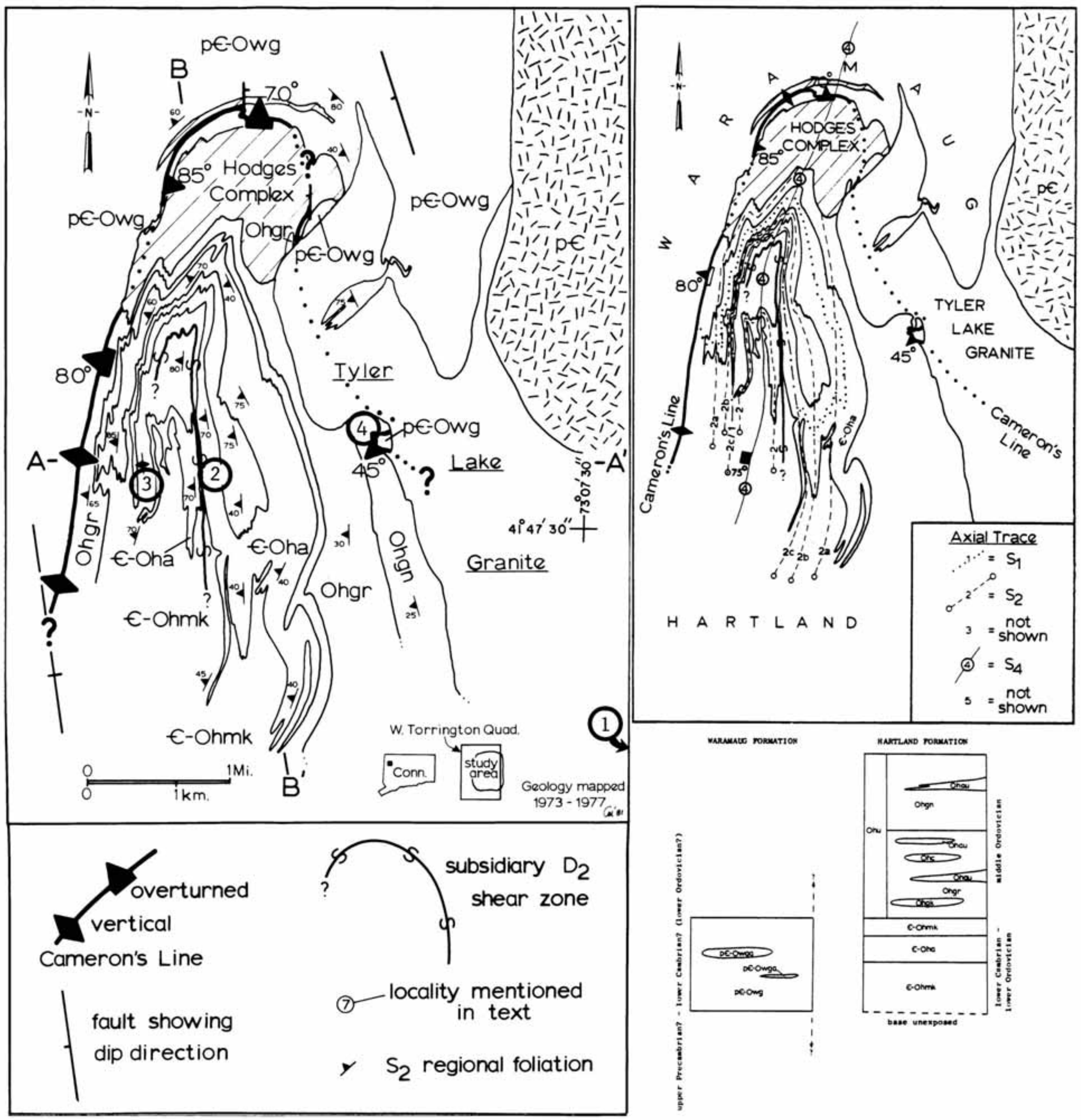


Figure 37. Simplified geological map (showing Stops 2-4), axial- surface map, and stratigraphic column of a part of the West Torrington quadrangle. (Charles Merguerian, 1985, fig. 4, p. 415.)

Mapping by Jackson (1980), and Jackson and Hall (1982) near Kent, Connecticut; by Alavi (1975) near Bedford, New York; by Hall (1968a, b) in White Plains, New York; and by Merguerian and Baskerville (1987) and Merguerian (unpublished data) in New York City support this correlation. Figure 34 shows regional correlations. In northwestern Connecticut, where the Waramaug is not as extensive it is where it was originally defined (Gates 1952), Dana (1977) and Hall (pers. comm., 1981) has redefined it. The Waramaug Formation is correlative and physically continuous between Connecticut and Massachusetts with the Late Proterozoic (?) to Cambrian Hoosac Schist (Hall, 1971, 1976; Hatch and Stanley, 1973; Merguerian 1977).

In western Connecticut, the Hartland Formation consists of a thick sequence of interlayered muscovite schist, micaceous gneiss and granofels, amphibolite, and minor amounts of calc-silicate rock, serpentinite, and manganiferous garnet-quartz granofels (coticule). Hartland rocks are correlative with metamorphosed eugeoclinal Cambrian to Ordovician rocks found along strike in New England. (See Figure 14.) Because the Hartland is overlain by Silurian and Devonian metamorphic rocks and the dominant regional foliation in the Hartland is truncated by the 383 +/- 5 Ma Nonewaug granite (Mose and Nagel, 1982), a pre-Silurian minimum depositional age for the sequence is indicated. Rocks mapped as Hartland extend from New York City (Seyfert and Leveson, 1969; Baskerville, 1982; Merguerian and Baskerville, 1987) through southeastern New York (Hall, 1968a; Pelligrini, 1977) and western Connecticut, northward to the Massachusetts state line (Hatch and Stanley, 1973).

In the vicinity of Torrington, based on stratigraphic positions, the Hartland has been subdivided into upper- and lower members. (See Figures 34 and 37.) The lower member (Ꞓ-Ohmk) consists of lustrous gray-weathering muscovitic schist typically containing large (up to 10 cm) porphyroblasts of garnet, biotite, staurolite, and kyanite (Stop 3). A texturally- and mineralogically diverse assemblage of thick, laterally variable amphibolites (Ꞓ-Oha) is interlayered within the lower member (Stop 2). The lower member grades, with some lensing, into the upper member.

The upper member (Stops 1, 4) consists of lustrous pin-striped muscovitic gneiss (Ohgn), well-layered quartzofeldspathic granofels and -schist (Ohgr), amphibolite (Ohau), and subordinate quartzite, coticule, and calc-silicate rocks (Ohc) and lenses of muscovite-kyanite schist (Ohmk). (See Figures 14 and 17.) The upper- and lower members of the Hartland Formation are correlative with the Rowe-Moretown-Hawley eugeoclinal sequence of western Masschusetts. (See Figure 14.)

Significant tectonic intercalation at Cameron's Line and intense regional isoclinal folding under amphibolite-grade metamorphic conditions create uncertainties in distinguishing between the Waramaug and Hartland formations near Cameron's Line. Elsewhere, their unique lithologic characteristics make identification simple. Waramaug rocks are generally rusty- to gray-weathering, coarse- to medium-textured, gneissic, and granular to foliated; quartz, biotite, and plagioclase are the dominant minerals. Muscovite is typically present but not as abundant as in the Hartland. The Waramaug contains thin layers of amphibolite and amphibolitic gneiss, which typically are granular and gray-green to black in color, and rare tremolite-quartz calc-silicate layers.

In contrast, the Hartland rocks weather gray. They are well layered, fine- to coarse-textured, and typically schistose with interlayers of granofels, amphibolite, and rare coticule. The rocks are very rich in muscovite and quartz and, to a lesser extent, plagioclase; they contain thin- to thick layers of greenish amphibolite.

The absolute ages of the Waramaug and Hartland formations are not known. Nevertheless, most workers (Hall, 1976) consider them to be basically time-stratigraphic equivalents. These two formations are inferred to be products of depositional settings that were predominantly transitional slope-rise (Waramaug) and adjacent deep-sea floor (Hartland). They were deposited in the Early Paleozoic seaward of the North American shelf. Subsequently, they were converted into metamorphic rocks and, during deformation, were juxtaposed at depth along Cameron's Line.

The Hodges Complex and the Tyler Lake Granite

Metamorphosed mafic- and ultramafic rocks of the Hodges Complex underlie an area of 2.5 square kilometers. (See Figures 37 and 38.) The complex is a steep-walled, folded, mushroom-shaped pluton. The core consists of hornblende gabbro core and the chilled margin, of dioritic rock. A stock-like central intrusive and many smaller separated masses of pyroxenite and hornblendite crosscut not only the main gabbro-diorite pluton but also the foliated amphibolites of the Hartland Formation that extend to the south. The pluton is in direct contact with both the Waramaug and Hartland formations and is surrounded by a narrow contact aureole.

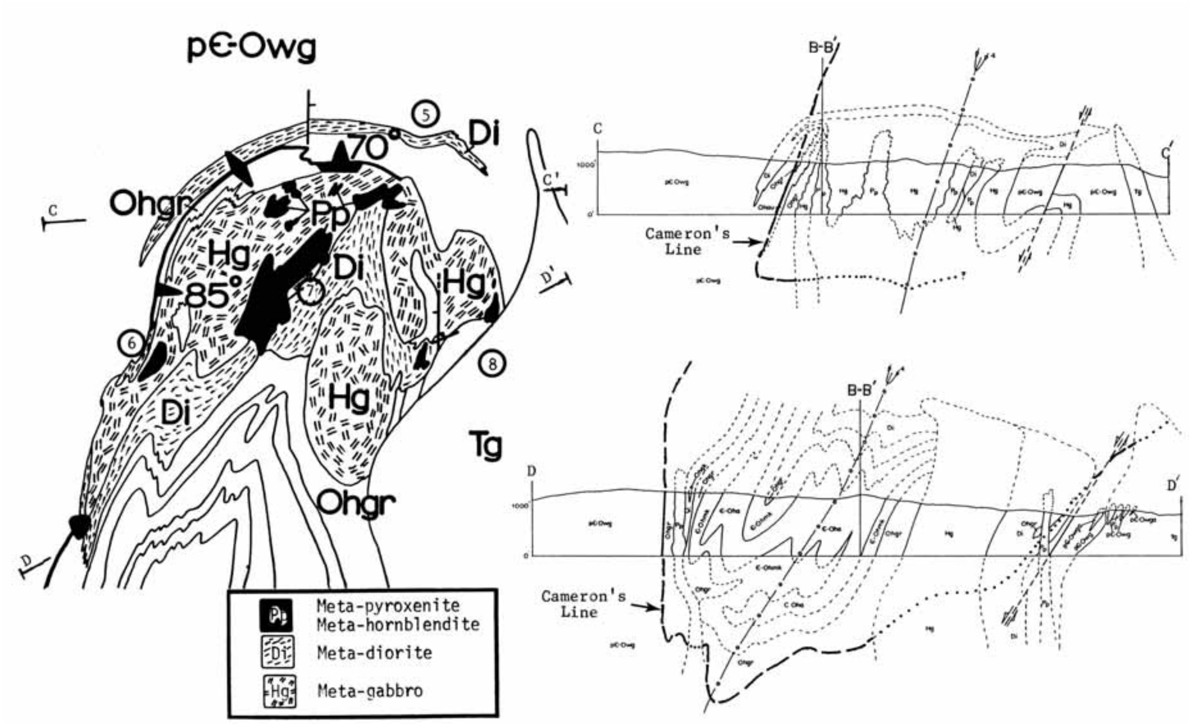


Figure 38. Geologic map and sections of the Hodges Complex showing Stops 5-8. (Charles Merguerian, 1985, fig. 5, p. 418.)

All rocks of the Hodges Complex have been metamorphosed. Their original igneous textures have been overprinted and recrystallized. Relict olivine, enstatite, hypersthene, augite, and hornblende have been corroded and replaced by tremolite, actinolite, anthophyllite, cummingtonite, hornblende, magnesian chlorite, calcite, talc, and serpentine minerals. Gabbros and diorites remain relatively nonaltered but some changes are evident. These include recrystallization of plagioclase and replacement of hornblende by biotite and by chlorite.

The Hodges is subdivided into three mappable metaigneous rock units. These are indicated on the map by the symbols: Pp = pyroxenite, hornblendite; Hg = gabbro; and Di = diorite. The ultramafic rocks (Pp) are typically highly altered, medium-textured to pegmatitic, dense and deeply iron-stained, silver-green to dark-green to black hornblende orthopyroxenite, biotite-tremolite-orthopyroxenite, orthopyroxene hornblendite, hornblendite, and biotite hornblendite (Stops 6, 7).

The main gabbroic mass of the Hodges Complex (Hg) is composed of medium- to very coarse-textured, dark-gray-weathering, hornblende-plagioclase-biotiteñquartz gabbro (Stop 6). Labradorite (An50-55), generally clouded displaying oscillatory zoning, is about equal to the content of hornblende. The hornblende contains pyroxene ghosts defined by opaques suggesting that prior to metamorphism, pyroxene was an important mineral phase.

The dioritic rocks (Di) are by far the most variable in texture and mineralogic composition (Stops 5-7). They are greenish to black and white, poor- to well-foliated, fine- to medium-textured, banded hornblende-plagioclase-biotiteñquartz diorites. Alternating layers of subhedral hornblende together with euhedral- to subhedral laths of plagioclase together define an igneous flow layering. The diorites form the flow-layered chilled margin of the main gabbroic mass of the Hodges.

The Tyler Lake Granite (Stop 8) is an elongate pluton initially described by Gates and Christensen (1965) as the eastern mass of the Tyler Lake Granite. It had been intruded across Cameron's Line and includes xenoliths of the Hodges rocks at their contact zone.

The $S_1 + S_2$ regional foliation exerted a strong control on the original geometric form of the Hodges and Tyler Lake intrusives. As a result, their shapes are sheetlike, rather than equidimensional. As discussed later, minerals of the Hodges contact aureole postdate the S_2 regional foliation in the Waramaug and Hartland wall rocks. Post-intrusive folding has deformed both Cameron's Line and the plutons; it brought about abundant metamorphic alteration of original igneous textures (Table 3).

Structural Geology, Intrusive Relationships, and Metamorphism

The wall rocks of the Hodges Complex and the Tyler Lake Granite have experienced a complicated Phanerozoic structural history that began with two phases of isoclinal folding (F_1 and F_2) yielding two subparallel regional foliations (S_1 and S_2). F1 folds are rare and usually developed in amphibolites which were less ductile than the surrounding schistose rocks during subsequent deformation (Figure 39a). However, in both the Waramaug and Hartland formations,

F_2 folds have commonly deformed an S_1 foliation and parallel compositional layering. The orientations and styles of the D_1 and D_2 events were similar. So, also, is the grade of metamorphism (amphibolite facies) which is considered to be progressive. They mark the initial prograde metamorphic pulse (M_1 in Table 3) that culminated during the formation of Cameron's Line.

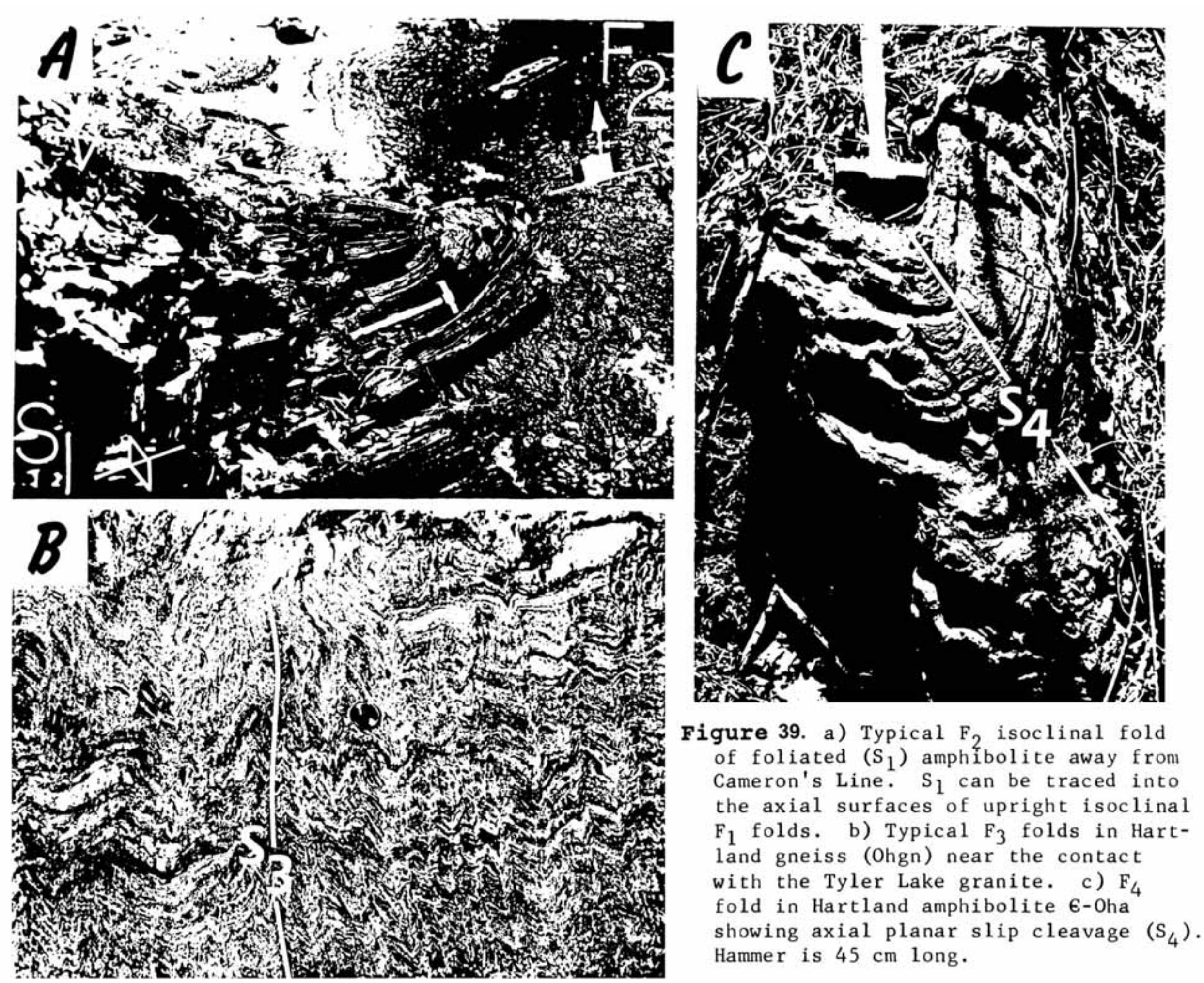


Figure 39. A) Typical F_2 isoclinal fold of foliated (S_1) amphibolite away from Cameron's Line. S1 can be traced into the axial surfaces of upright isoclinal F1 folds. B) Typical F_3 folds in Hartland gneiss (Ohgn) near the contact with the Tyler Lake granite. C) F_4 fold in Hartland amphibolite Ꞓ-Oha showing axial-planar slip cleavage (S_4). Hammer is 45 cm long. (Charles Merguerian, 1985, fig. 10, p. 430.)

Cameron's Line is a zone, 15 to 90 m wide, of intense localized isoclinal F_2 folds, having limbs sheared parallel to S_2, with S_1 fabrics transposed, and that regionally truncate Hartland subunits. The synmetamorphic shear zone includes layers of mylonitic amphibolite intercalated with both Waramaug and Hartland rocks and, locally, deformed slivers of serpentinite (Stop 4). Away from Cameron's Line, D_2 created a penetrative regional foliation (S_2) in the Waramaug and Hartland Formations. Although it is not clear whether motion along Cameron's Line, initiated during D_1, has anything to do with the fact that S_2 axial surfaces and the trace of Cameron's Line in West Torrington (See Figure 17.) are regionally parallel. However, this parallelism strongly suggests that Cameron's Line formed at essentially at the same time as the S_2 in the wall rocks. A subsidiary D_2 shear zone, marked by mylonitic amphibolite (Stop 2) and a soapstone-talc body (optional Stop 2a) are also are present within the Hartland.

A secondary regional metamorphic pulse (M_2 in Table 3) occurred after the Waramaug and Hartland formations had been juxtaposed. This is demonstrated by the fact that porphyroblasts of garnet, staurolite, and kyanite overgrow the S_2 foliation. Because the S_4 cleavage deforms the M_2 porphyroblasts, the peak of the M_2 metamorphic pulse was reached after D_2 but before D_4. The contact effects of the Hodges Complex also overprint the S_2 foliation. Therefore, it is likely that the Hodges was intruded synchronously with the regional M_2 event as shown in Table 3. Based on a Rb/Sr age on the Tyler Lake Granite of 466 +/- 12 Ma, reported by Merguerian and others (1984), both the M_1 metamorphic pulse and the development of the Hodges contact aureole are of pre-medddle Ordovician age. It is possible that the M_2 event is also of Ordovician vintage despite the fact that most workers in Western Connecticut attribute the growth of large post-regional foliation porphyroblasts to Acadian (middle Devonian) Barrovian metamorphism that has been documented in Massachusetts (Hatch, 1975; Stanley, 1975; Robinson and Hall, 1980).

Open- to tight, crenulate F_3 folds are present, but predominantly in the vicinity of the plutons (Figure 39b). Their axial-surface cleavages commonly are parallel to the margins of the plutons; their regional effect on the map pattern of the plutons is trifling. These folds are interpreted as being syn-intrusive; their axial surfaces are not shown in Figure 37.

Near West Torrington the $S_1 + S_2$ foliations, Cameron's Line, and the Hodges and Tyler Lake plutons were strongly deformed by dextral F4 folds (Figure 39c) and cut by an associated axial-planar spaced schistosity. S_4 is characterized by the growth of idioblastic biotite and -hornblende and recrystallized quartz; by parting in M_2 garnet-, staurolite-, and kyanite porphyroblasts; and by brittle deformation of plagioclase twin lamellae. The S_4 schistosity crosscuts the large M_2 staurolite-kyanite-garnet porphyroblasts as well as the $S_1 + S_2 = M_1$ regional foliation (Table 3). Metamorphism (M_3) during the D_4 event fostered retrograde biotite and recrystallization of amphibole. In addition, the D_4 event caused widespread metamorphic recrystallization, serpentinization, and chloritization in the Hodges Complex and recrystallization and domainal shearing in the Tyler Lake Granite (Stops 5-8).

A fifth-, and possibly sixth, deformation is suggested by the warping of the S4 axial-surface trace (See Figure 37.) and by local open- to crenulate folds with variable plunges and shallow NE- and NW- to W-trending axial surfaces. This deformation is low grade and is marked by recrystallized quartz and chlorite +/- white mica. These crosscutting structural-, metamorphic-, and intrusive relationships are summarized in Table 3 and discussed in greater detail in Merguerian (1977, 1983, 1985). The D_x, F_x, S_x, and M_x nomenclature (to denote deformational event, fold generation, axial-surface fabric, and metamorphic event, respectively) will be utilized in later field descriptions.

Stereograms and Structure Sections

Stereograms of the major structural features described above are shown in Figure 40. Stereogram 1 shows poles to S_2 in both the Waramaug and Hartland formations. The wide scatter of poles distributed about a NW-SE girdle indicates the presence of features resulting from post-D_2 deformation. Poles to S_4 (Stereogram 2) and F_4 fold axes and L_4 intersection lineations (Stereogram 3) show a consistent trend for S_4 of about N19°E, 72° NW and F_4 about S50°W at 60°. Clearly, the girdle distribution of S_2 poles is largely the result of F_4 folding. Some scatter resulting from local F_3 and F_5+ folds may have also occurred.

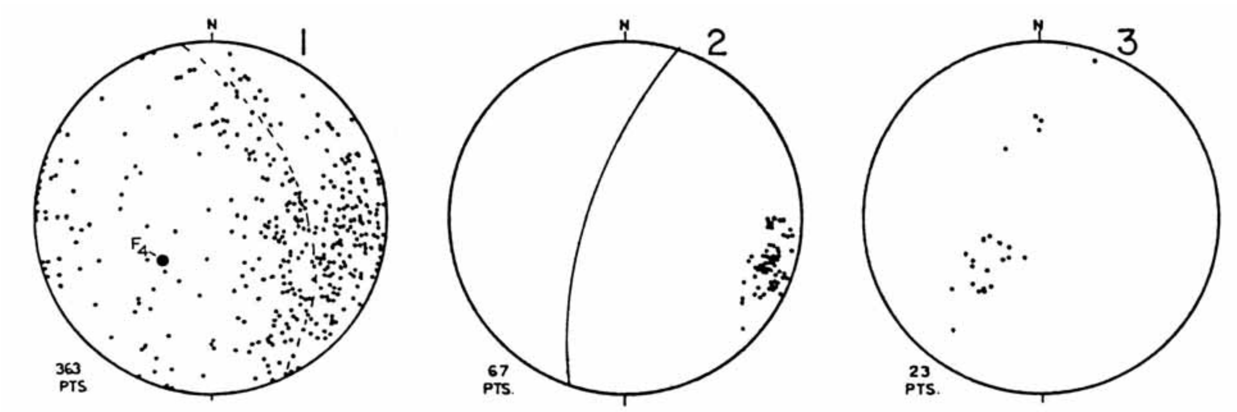


Figure 40. Stereograms of structural elements. (Merguerian, 1985, fig. 7, p. 422.)

Sections in Figure 41 have been drawn from map data and from axial-surface traces shown in Figure 37 but the exact configuration of F_1 closures in the subsurface is hypothetical. This has been caused by extensive D_2 transposition. In section A-A', the major obvious structure is a dextral F_4 synform with a steep western limb (vertical to locally overturned toward the east) and a shallow west-dipping eastern limb. The interference of F_1 and F_2 folds yields a complex interdigitating map pattern of Hartland subunits. The section shows that F_4 folds have been superimposed on the older structures, that Cameron's Line and the subsidiary D_2 shear zone have been folded, that Cameron's Line truncates Hartland subunit Ohgn, and that the Tyler Lake Granite is a discordant pluton.

Section B-B' shows a north-south view roughly parallel to the trace of S_4. Again, the complicated pattern of folds of the Hartland subunits, truncation of Hartland subunit Ohgn, and crosscutting relationships of the Hodges Complex are indicated.

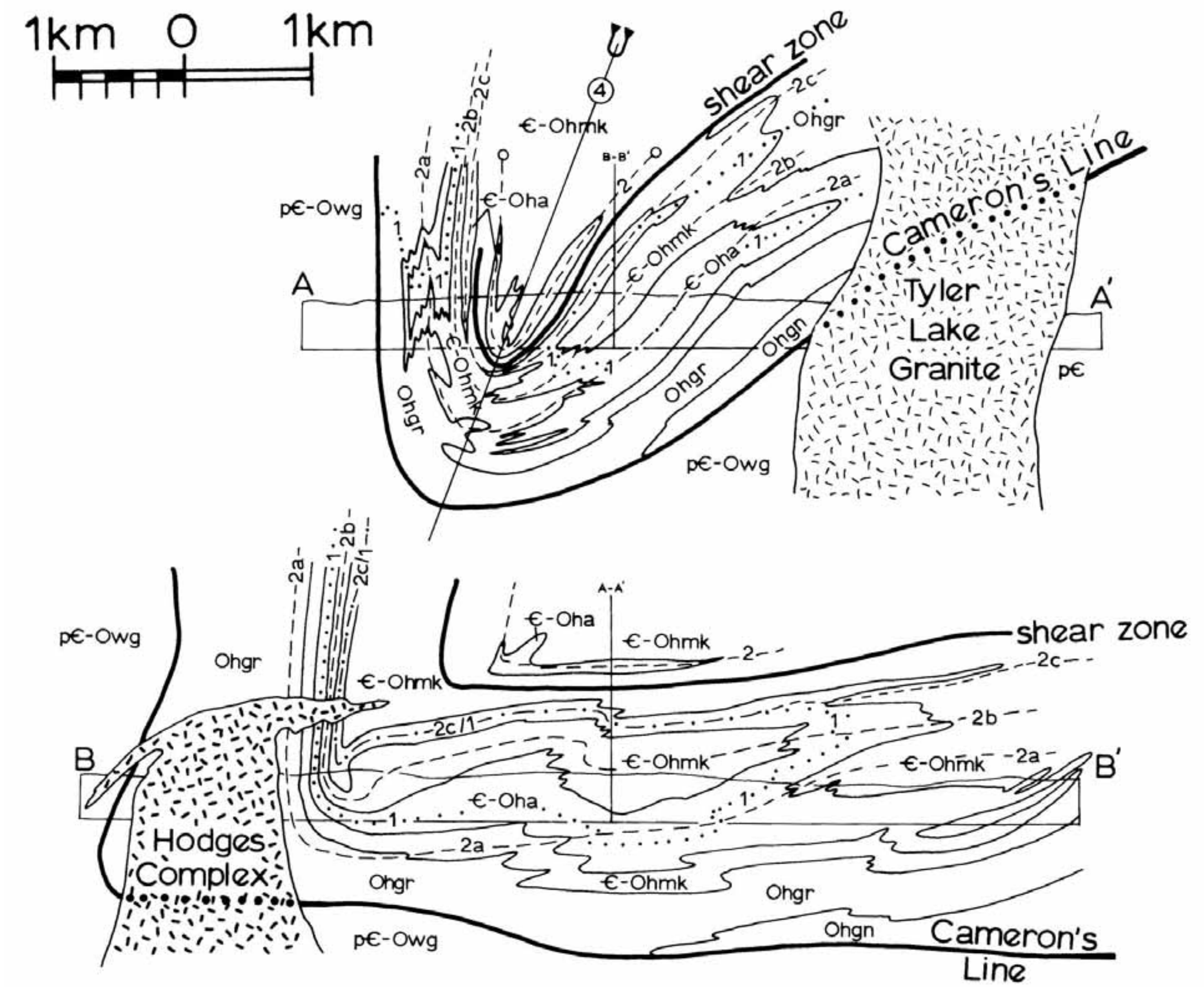


Figure 41. Geologic structure sections. Section lines are shown in figure 37. No vertical exaggeration. (Charles Merguerian, 1983, fig. 6, p. 358.)

DRIVING DIRECTIONS

NYAS to the western Connecticut via the Major Deegan Expressway, Routes 684, 84, and 8).

Turn L on 5th Ave. and drive south to 62nd Street. Turn L toward the East Side Drive and take the Drive north to the Third Avenue Bridge over to the Major Deegan Expressway (I-87N). Past Ardsley, New York, take the Saw Mill River Parkway northward toward Bedford Hills, New York. Just past Bedford Hills, switch to I-684(N) toward Brewster, New York and then switch to I-84(E) to Connecticut. Travel east to Exit 8 (Route 6) and at the end of the exit ramp turn R to shopping center entrance on R. This will be a brief rest stop (Friendly's); those of you who forgot to bring lunch can pick some up here (Pathmark).

To Stop 1: Turn L out of shopping center and follow signs to I-84(E). Continue east on I-84 until intersection with Connecticut Route 8. Take Route 8 northward toward Torrington, Connecticut, through the heart of the crystalline bedrock of western Connecticut. Note the interesting roadcuts on both sides of the road. Continue north to Exit 42 (CT Routes 118 and 8). Bear R (east) on Route 118 at the end of the exit ramp and park in the commuter parking lot. Outcrops of the Hartland occur across from the parking lot on the north side of Route 118. A detailed road log is presented later under a separate section heading entitled Description of Individual Localities (Stops).

DESCRIPTIONS OF INDIVIDUAL LOCALITIES ("STOPS")

STOP 1 - Hartland Formation (upper member) granofels, schist, and amphibolite. [UTM Coordinates: 656.80E / 4624.88N, Torrington quadrangle.]

A convenient place to initiate today's field trip, the rocks exposed in the roadcut across from the commuter lot were originally described by Martin (1970). Here, 2- to 15-cm-scale very well-layered muscovite-biotite-plagioclase-quartz-(hornblende)-(garnet) granofels are interlayered with schist having similar mineral assemblages. The major minerals are listed in order of decreasing abundance, those in parentheses are not found in all exposures. The abundance of muscovite in the granofels and schist creates a lustrous sheen from foliation surfaces that reflects sunlight. A layer of hornblende-plagioclase-biotite-epidote-quartz-(garnet) amphibolite 2 m thick is exposed on the south-facing portion of the exposure. The pervasive interlayering of granofels and schist, high content of muscovite and plagioclase, and presence of amphibolite suggest that the protoliths of these rocks were volcaniclastic graywackes and interlayered shale with subordinate basalt flows. The Hartland upper member is similar to and correlative with the Moretown Formation of western Massachusetts. (See Figure 34.)

The dominant layering is parallel to the composite $S_1 + S_2$ regional foliation, all striking roughly N65°E and dipping 60°NW. The $S_1 + S_2$ foliation has been deformed by crenulate F3 folds with axial surfaces oriented N30°E, 26°SE. The F_3 hinge lines are expressed as L_3 crinkle-axis lineations in highly micaceous layers and as L_3 intersection lineations in more-massive granofels. The F_3 and L_3 elements trend N55°E and plunge 9°. Note the upright warping of S_3 axial surface traces and the decrease in wavelength of F_3 folds in mica-rich interlayers. A pegmatite 2 m thick intrudes across the S_3 axial surfaces, locally rotating F_3 folds and older fabrics. Note the F_4 "z" folds with N20°E, 84°NW axial-planar slip cleavage.

Near the eastern end of the roadcut, L_2 lineations are deformed by subhorizonal F_3 folds and overprinted by L_3 lineations. The associated S_3 axial surfaces (N40°E, 20°SE) are warped by late F_4 crenulations with axial surfaces oriented roughly N25°E, 70°NW. Broad arching by later F_5+ is also evident.

On the west-facing portion of the roadcut adjacent to the northbound entrance ramp for Route 8, F_2 intrafolial folds are present in thinly layered granofels. The S_2 axial surface strikes N55°E and dips 56°NW and F_2 hingelines are subhorizontal, trending N55°E-S55°W. The F_2 folds deform a pre-existing S_1 mica foliation. To the north along the roadcut, many amphibolite layers are exposed.

MILEAGE TOTAL INTERVAL

0.0 0.0 Exit the commuter lot and take the northbound ramp for Route 8 toward Torrington. In the exposure 0.6 miles from the starting point, note the upright F_2 folds. At exit 44 (Routes 4 and 202, Downtown Torrington) follow the exit ramp to the traffic signal.
3.3 3.3 Make a left traveling westward on Route 202 (East Main Street) past three traffic lights.
3.9 0.6 At the fourth traffic light, bear right (across Main Street) up the hill onto Water Street. Follow Water Street past the railroad tracks to the traffic light.
4.3 0.4 Turn left onto Church Street. Drive over the Naugatuck River that separates Proterozoic Y gneiss of the Berkshire massif on the east from the Tyler Lake Granite on the west. Follow Church Street to the small traffic triangle.
4.6 0.3 Turn left (west) driving uphill onto Highland Avenue. For the next 0.9 miles, the Tyler Lake Granite crops out in wooded areas away from the road. Pass Allen Road on the right (5.2 mi.) and Stop 4 near Patterson Pond (5.6 mi.). Continue west, now driving on Ohgr and pass the radio towers to the right (6.3 mi.), which essentially mark the contact between the upper- and lower members of the Hartland.
6.4 1.8 Pass Westside Road on the right and Rossi Road on the left and continue uphill to the massive outcrops on either side of the road (here known as Soapstone Hill Road!).
6.8 0.4 Park in the bend of the road just past the exposures.

STOP 2 - Hartland Formation (lower member) amphibolite and subsidiary D_2 shear zone. [UTM Coordinates: 651.75E / 4629.13N, West Torrington quadrangle.]

The roadside exposures consist of fine- to medium-textured, dark-green hornblende-plagioclase-biotite-(quartz)-(epidote)-(chlorite)-(garnet) amphibolite with lineated prismatic hornblende. Elliptical quartz segregations up to 4 cm thick lie within the S_2 foliation. Elsewhere, felsic granofelsñhornblende, +/- biotite, +/- chlorite in layers 1 to 2 m thick, are interlayered with the amphibolite and the muscovitic schist.

The S_2 foliation strikes N15°W and dips 67°SW and a prominent L_2 hornblende lineation trends N80°W and plunges 65°. Because of intense transposition and overprinting during D_2, S_1 is essentially coplanar with S_2. S_2 and L_2 parallel the axial surfaces and hingelines, respectively, of rootless F_2 isoclinal folds exposed on gently northeast-dipping joint faces. The S_1 foliation, composed of hornblende and plagioclase, is locally preserved in F_2 hinges. Because of the coplanar $S_1 + S_2$ foliation and an oblique S_4 spaced cleavage striking N-S and dipping 75°W, the amphibolite tends to break into wedge-shaped pieces. Oriented samples show that S_4 is defined by idioblastic biotite, the product of M3 metamorphism.

Walk 110 m west on Soapstone Hill Road where amphibolite exposures exhibit S_2 mylonitic layering (N14°W, 82°SW). Between these exposures, the muscovite schist is phyllonitic and thin (Figure 41; See also Figure 37.). The mylonitic textures may mark a subsidiary D_2 shear zone that imbricates the Hartland amphibolite. Alternatively, shearing could simply have been the result of ductility contrasts developed across the amphibolite-schist contact.

STOP 2a (Optional) - Soapstone Quarry. [UTM Coordinates: 653.63E / 4629.55N, West Torrington quadrangle.]

Walk north on a dirt trail immediately west of the parking area for Stop 2. Along the way, ridges are composed of amphibolite and the intervening valleys are underlain by muscovitic schist. Roughly 700 m north, the trail ends at a pit, of a former soapstone quarry, that is 90 m long by 20 m wide. (They don't call it Soapstone Hill Road for nothing, you know!). The excavation, which is oriented parallel to S_2 in the bounding muscovite-chlorite schist, produced commercial quantities of soapstone. Blocks from the tailings pile include talc-tremolite schist, chlorite schist, and very coarse amphibolite rich in opaque minerals. The quarry is on strike with mylonitic amphibolite to the south. The elongate shape parallel to S_2 and foliated nature of the altered serpentinite body suggests that the soapstone-talc body represents ultramafic rock deformed during D_2 and possibly D_1. It may mark a syntectonic ultramafic intrusive (Gates and Christensen, 1965) or a small sliver of ophiolite (Merguerian, 1979).

6.8 0.0 Continue west on Soapstone Hill Road and pull into a large clearing to the left.
7.3 0.5 Exposures for Stop 3 are in the woods north of the road.

STOP 3 - Hartland Formation (lower member) muscovite-kyanite-staurolite schist. [UTM Coordinates: 651.10E / 4629.12N, West Torrington quadrangle.]

The lower-member Hartland schist crops out less than 50 m north of the road. The rocks are highly lustrous, gray-weathering, medium- to coarse quartz-muscovite-plagioclase-biotite-opaque-(garnet)-(chlorite)-(apatite) schists often containing 1-to 10-cm porphyroblasts of kyanite, staurolite, and garnet, and more rarely, plagioclase and biotite. The proportions of quartz and muscovite are roughly equal. Together, these two minerals constitute more than half the rock. The appearance of deeply eroded exposures is knotted. This results from differential weathering of porphyroblasts. Granular, clear- to smoky-gray quartz pods are conspicuous and have been flattened into S_2. Near contacts with amphibolite, the content of hornblende, chlorite, and/or biotite in the lower member increases markedly. The muscovite schist, amphibolite, and rare felsic granofels of the lower member are probably derived from metamorphosed pelitic sediments within which units of basalt and rare volcaniclastic layers (ash-fall tuffs?) were interbedded. The rocks are lithically correlative with the Rowe Schist of western Massachusetts. (See Figure 34.) The large, non-oriented porphyroblasts of kyanite, staurolite, and garnet overgrow the S_2 foliation and represent the M_2 metamorphism (Table 3). Kyanite tends to occur mimetically within S_2. Staurolite tends to form spongy porphyroblasts, sometimes twinned, protruding randomly from the schist.

7.3 0.0 Backtrack east on Soapstone Hill Road.
9.0 1.7 Make a sharp right onto a partly hidden dirt road just past Patterson Pond. Pull up as far as possible.

STOP 4 - Cameron's Line and dismembered ophiolite. [UTM Coordinates: 653.70E / 4629.30N, West Torrington quadrangle.]

The mylonitic amphibolite (Ohau) exposure on the dirt road occurs within Cameron's Line, a zone 90 m thick consisting of highly sheared, tectonically intercalated units of the Hartland (upper member) and the Waramaug formations (Figure 42). In the amphibolite, an S_2 mylonitic foliation (N85°W, 70°NE) is parallel to the axial surfaces of F_2 folds with sheared-out limbs plunging 40° into N75°W (Figure 43a). In the hinge areas of F_2 folds, one finds an S_1 foliation composed of aligned hornblende. A specimen collected from this exposure after blasting in 1973 shows an F_1 isocline refolded by F_2 with significant shearing and recrystallization parallel to S_2 (Figure 43b). F_3 folds with subhorizontal axial surfaces warp S_2. The mylonitic amphibolite is interlayered with lustrous muscovitic gneiss (Ohgn).

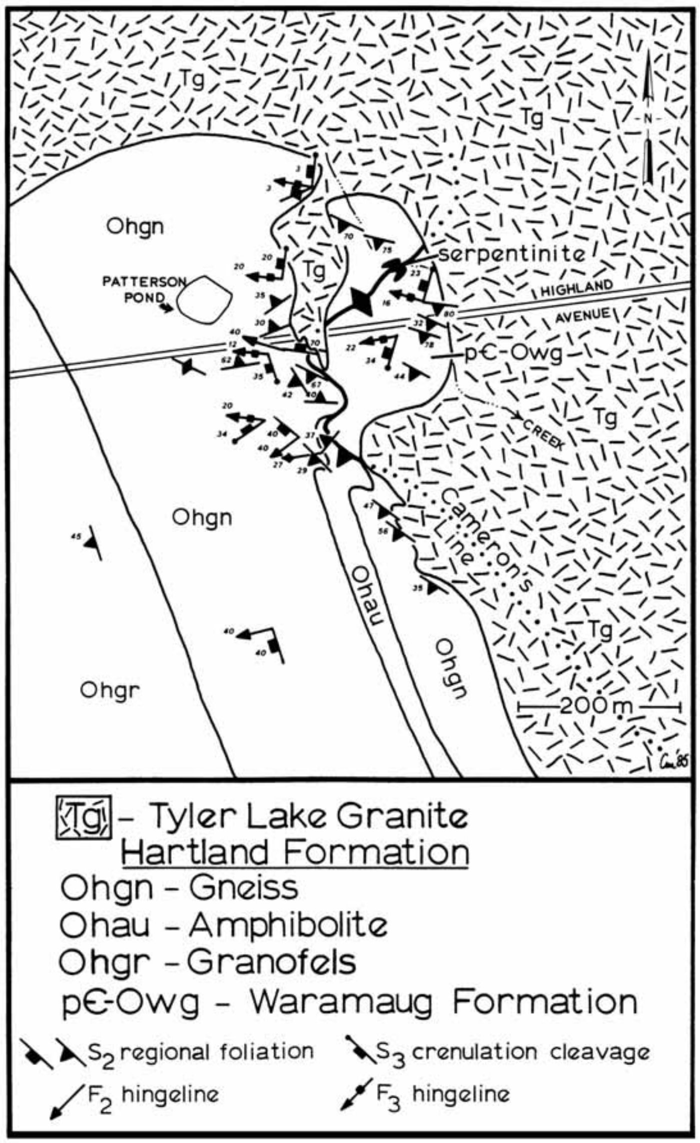


Figure 42. Geological sketch map in the vicinity of Stop 4. (Charles Merguerian, 1987, fig. 6, p. 163.)

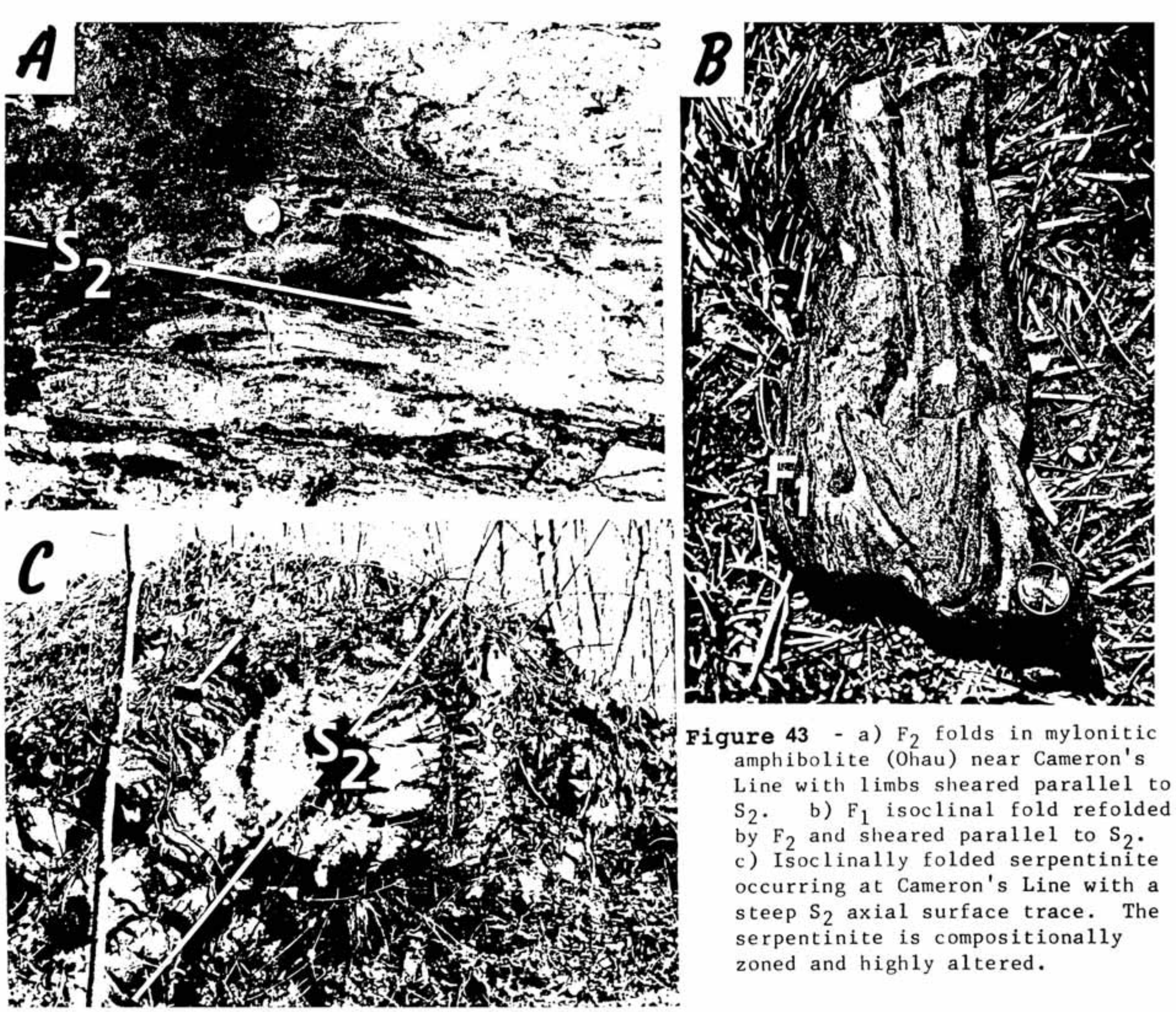


Figure 43. Folds in mylonitic amphibolite near Cameron's Line. A) F_2 folds in mylonitic amphibolite (Ohau) near Cameron's Line with limbs sheared parallel to S_2. B) F_1 isoclinal fold refolded by F_2 and sheared parallel to S_2. C) Isoclinally folded serpentinite occurring at Cameron's Line with a steep S_2 axial surface trace. The serpentinite is compositionally zoned and highly altered. (Charles Merguerian, 1985, fig. 10, p. 430.)

The Hartland crops out to the north, west, and south. (See Figures 37 and 42.) Follow the dirt road south, stopping first to examine the Hartland gneiss exposures on the subdued knob to the east. Here, because of F_3 folding, S_2 dips gently toward the west. About 120 m to the south, note mylonitic amphibolite in the woods to the east. S_2 strikes N40°E and dips 37°NW; L_2 trends S80°W and plunges 27°. Farther south and east, the Tyler Lake Granite crops out. (See Figure 42.) Walk back to Highland Avenue and walk east roughly 100 m. The first exposure on the right (before the creek) consists of massive, rusty-weathering quartz-plagioclase-biotite-sillimanite-muscovite-garnet-chlorite-tourmaline gneiss of the Waramaug Formation. The laminated S_2 mylonitic foliation strikes N65°W and dips 32°SW. The Waramaug crops out to the south and north. As such, Cameron's Line is situated between these exposures and the cars.

As we walk north, adjacent to the small creek, note the overturned F_3 fold with pegmatite intruded along S_3. The small hill to the west is underlain by southwest-dipping Waramaug gneiss. Trace the creek to where a 10-m isoclinally folded serpentinite body separates Waramaug rocks to the southeast from Hartland rocks to the west and northwest. Folded by F_2 folds (See Figure 40c.), the serpentinite is zoned and highly altered and contains relict olivine and orthopyroxene. The zoning (compositional? or tectonic?) has resulted from relative enrichment of greenish, intergrown cummingtonite and tremolite in the upper part of the body compared to the dense, black serpentine- and anthophyllite-enriched lower part.

In mineral composition and texture, this body differs from ultramafic rocks of the Hodges Complex. The overall eugeoclinal nature of the Hartland and the features resulting from D_2 (and possibly D_1) deformation in the serpentinite suggest that the body represents dismembered ophiolite (Merguerian, 1979).

9.0 0.0 Drive east on Highland Avenue back to the turnoff from Church Street.
10.0 1.0 At the stop sign, turn left onto Riverside Avenue. At the entrance to Charlene Susan Besse Park (10.9 mi.), note the exposures of Waramaug amphibolite (p-Ꞓ-Owga). On the hilltop roughly 500 m due south, Waramaug rocks are exposed. Continue north on Riverside Avenue to the intersection with State Route 4.
11.3 1.3 Turn right onto Route 4 and drive 0.25 miles east to Scarpelli's Drive-In. Those without lunch can pick something up here but we'll lunch at a nearby picturesque spot. With supplies in hand, drive west on Route 4 to the traffic light.
11.8 0.5 Turn right onto Route 272 toward Wrightville.
12.6 0.8 Turn right onto Brass Mill Dam Road.
12.9 0.3 LUNCH STOP. Park in the wide area south of the dam. There are nice places to lunch to the northwest.

Post Lunch Stop

In the woods to the west, the Waramaug crops out. In the spillway north of the cars, are other excellent exposures. Here, the Waramaug and interlayered amphibolite have been deformed into 15-m amplitude F_2 folds with N42°E, 57°NW axial surfaces and hingelines trending S40°W at 15°. F_5 crenulate "z" folds trend W at 37°; the attitude of the S_5 axial-surface cleavage is N15°W, 52°SW. Because abundant closely spaced fractures oriented N20°W, 75°SW cut the exposed Waramaug rocks hereabouts, Stillwater Pond may have been localized by faults. If so, then it would be referred to as being fault controlled.

12.9 0.0 Backtrack on Brass Mill Dam Road.
13.2 0.3 Drive across Route 272 uphill onto Hodges Hill Road.
13.7 0.5 At the stop sign, turn right onto University Drive (Town Farm Road).
14.1 0.4 Turn left onto John Brown Road and continue to dirt-trail turnoff to the left.
14.3 0.2 Pull in as close as possible; we're at Stop 5!

STOP 5- Waramaug Formation and contact relationships of Hodges diorite sill. [UTM Coordinates: 652.38E / 4633.12N, West Torrington quadrangle.]

Walk roughly 60 m south on the dirt trail to an exposure of massive but internally laminated, gray-weathering quartz-plagioclase-biotite-sillimanite-muscovite-garnet gneiss. The rocks at this exposure are similar to those of the Waramaug at Stop 4. Because of differential erosion of quartz and sillimanite, the weathered surface is nubby. A 30-cm layer of garnet amphibolite has been isoclinally folded by F_2. Here, S_2 strikes N80°W and dips 83°SW. L_2 lineations trend N85°W at 26°. A spaced S_4 slip cleavage oriented N26°E, 84°NW deforms S_2.

To the northeast and southwest, the Waramaug crops out (Figure 44). The exposures on the northeast show S_1 metamorphic layering trending N20°E, 17°NW that has been folded by upright antiformal F_2 "m" folds having vertical axial surfaces oriented N75°W, and hingelines trending N75°W at 17°. To the southwest, the exposures illustrate the typical massive, nubby-weathered appearance of the Waramaug Formation. S_2 strikes N80°W and dips 83°NE; a prominent L_2 lineation trends N70°W at 36°. L_2 has been produced by intersection with S_1 that is locally preserved at a small angle to S_2. Commonly, S_1 has been transposed into parallelism with S_2 but here, because of F_2 isoclinal folding, S_1 dips 70° to 75° NE. The axial surface traces of F_2 isoclines are indicated in Figure 44.

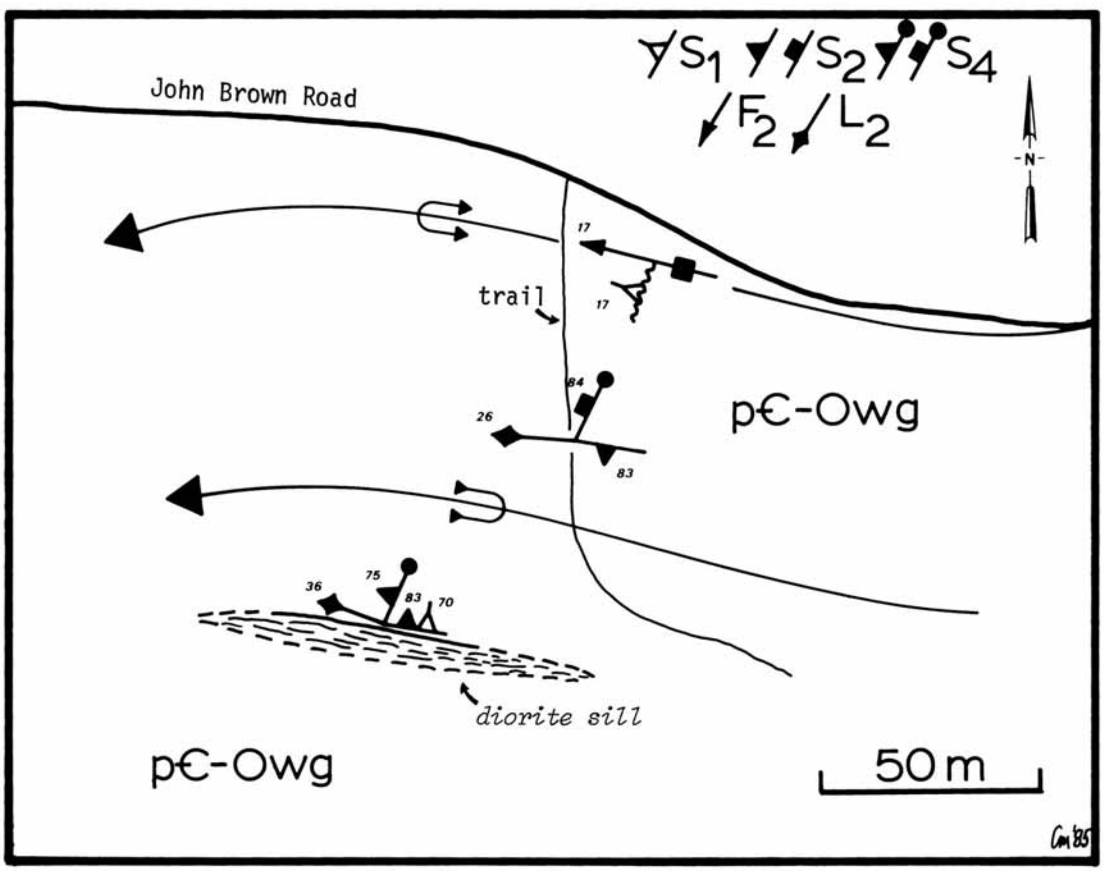


Figure 44. Geological sketch map of Stop 5 showing outcrops described in text and trace of major subvertical S_2 axial surfaces. (Merguerian, 1985, fig. 11, p. 433.)

At the southern margin of the gneiss, note the abundance of garnet. The enrichment marks the contact effects of adjacent flow-layered diorite of the Hodges Complex. The diorite forms a small sill intruded parallel to S_2 in the Waramaug. (See Figure 44.) The microscope shows that in the Waramaug wall rocks, the contact garnets (up to 1 cm) have grown across S_2. Garnet has also been enriched in the diorite. This enrichment suggests that limited alumina metasomatism from the wallrocks took place.

In the woods to the east and west, Hodges diorites were intruded as sills and lit-par-lit injection bodies (typically less than 10 m thick) along S_2 in the Waramaug. Flow layering, defined by oriented hornblende and biotite set in a matrix composed of plagioclase, is regionally parallel to S_2. In addition, garnetiferous diorite sills to the east are folded by SW-plunging F_4 folds and cut by a spaced S_4 biotite schistosity (developed during M_3) oriented N25°E, 75°NW. The near east-west trend of S_2 in this area has resulted from the effects of the major dextral F_4 fold. (See Figure 37.) On the flanks of the F_4 fold (Stops 2-4), S_2 trends approximately NNE, but in the F_4 hinge areas (Stops 1, 5), the trends are nearly east-west.

The post-S_2 field- and textural evidence on the contact relationships of the diorites fixes the time of intrusion as being late syn- to post-D_2. Crosscutting by features made during D_4 deformation and the overprinting effects of M_3 place an upper relative age limit on the Hodges intrusive episode.

14.3 0.0 Backtrack east on John Brown Road and turn right (south) on University Drive toward Route 4.
15.4 1.1 Turn right (west) on Route 4 and continue past Klug Hill Road and Wright Road to Weed Road.
16.4 1.0 Turn left (south) onto Weed Road and park about 0.1 miles south along the side of the road.
16.5 0.1 The 1320' hill to the west is Stop 6.

STOP 6 (Optional) - Mafic and ultramafic rocks of the Hodges Complex. [UTM Coordinates: 652.15E / 4632.64N, West Torrington quadrangle.]

As we walk up the overgrown trail westward from Weed Road, notice that the hill to the west is primarily composed of hornblende gabbro. Locally, this rock is melanocratic and its texture, porphyritic. Concentrations of mafic minerals and oriented hornblendes define a west-dipping flow layering. Near the top of the hill and in a small pod to the south, coarse pyroxenite and hornblendite crop out. (See Figure 38.) Here, the intrusives mask Cameron's Line. But based on detailed tracing of screens and xenoliths, CM infers that Cameron's Line crosses the top of the hill in a S20°W direction. The Hodges rocks are in contact with both the Waramaug and Hartland to the west and east, respectively.

To the east, many xenoliths and screens of the Hartland (upper member) display the effects of contact metamorphism. New minerals include cordierite, kyanite, sillimanite, staurolite, and garnet. Muscovite has been eliminated. In the Hartland amphibolite, garnets up to 3 cm across have overgrow the S_2 foliation. Float blocks draping the slopes to the east of Weed Road include pieces showing hornblende porphyroblasts overgrown on S_2 in Hartland amphibolite. Contact-metamorphic assemblages are fully discussed at the next stop (7).

To the west, near exposures of the Waramaug, flow-layered diorite trends N25°E; the dips are vertical to steep easterly. The Waramaug consists of a dense hornfels peppered with garnet. Despite these contact-mineral changes, the characteristic nubby weathering is still preserved. Along the western slope of the 1320' hill, the Waramaug contains white tremolitic calc-silicate layers.

16.5 0.0 Return to Route 4 and turn right. Drive slow and prepare to turn left at Wright Road.
17.1 0.6 Follow Wright Road north and park near the first barn on the left.
17.2 0.1 The ridge to the northwest is Stop 7.

STOP 7 - Mafic- and ultramafic rocks of the Hodges Complex. [UTM Coordinates: 651.80E / 4632.11N, West Torrington quadrangle.]

Walk from the barn northwestward onto a ridge where flow-layered diorites grade westward into gabbroic rocks. The diorites were multiply intruded as thin sill-like masses parallel to S_2 in the Harland Formation. In the contact zone, both the Hartland and the diorites have been strongly enriched in garnet. Flow layering in the diorites is oriented N65°E, 70°NW. Near the top of the ridge, massive gabbroic rocks are present; they extend toward the northeast. (See Figure 38.) To the west, beyond the ridgecrest, a large NE-trending mass of pyroxenite and hornblendite crops out. (See Figure 38.) The ultramafic mass, which crosscuts the diorite-gabbro contact and truncates flow layering in the diorite, is interpreted as the youngest intrusive in the Hodges Complex. Locally, the ultramafic rocks have been strongly sheared and transformed into laminated serpentinite. The shear zone, oriented N36°E, 75°NW, cuts across the central part of the Hodges. It has resulted from shearing along the axial surface of the major dextral F_4 fold which deformed Cameron's Line, the Hodges Complex, and the Tyler Lake Granite into a broad dextral flexure. In fact, the plutons may have acted as immobile plugs localizing the F_4 hinge area.

To the southwest, the S_4 shear zone is marked by zones that have been affected by serpentinization and chloritization. In between these zones, relatively nonaltered pyroxenite and hornblendite are preserved. Contact metamorphism of the Hodges wall rocks. The contact minerals of the Hodges Complex overprint S_2 in both the Waramaug and Hartland formations. In the contact aureole, typical foliated textures in the wallrocks have been replaced by dense, finer-textured, garnet-rich hornfels (Figure 45a). In the contact aureole, amphibolite contains post-S_2 garnet and hornblende as discussed at Stop 6 (Figure 45b). Where gabbroic rocks have been intruded into the Hartland Ohc subunit, a randomly oriented colorless amphibole of the cummingtonite-grunerite series forms. To the northeast of Klug Hill, gabbroic rocks intrude the Hartland granofels (Ohgr). They have produced a unique cordierite-kyanite-staurolite-garnet-biotite-plagioclase (An23) assemblage (Figure 45c). In the contact aureole exposed at Stop 6, staurolite is abundant and the garnet contains microlites of sillimanite.

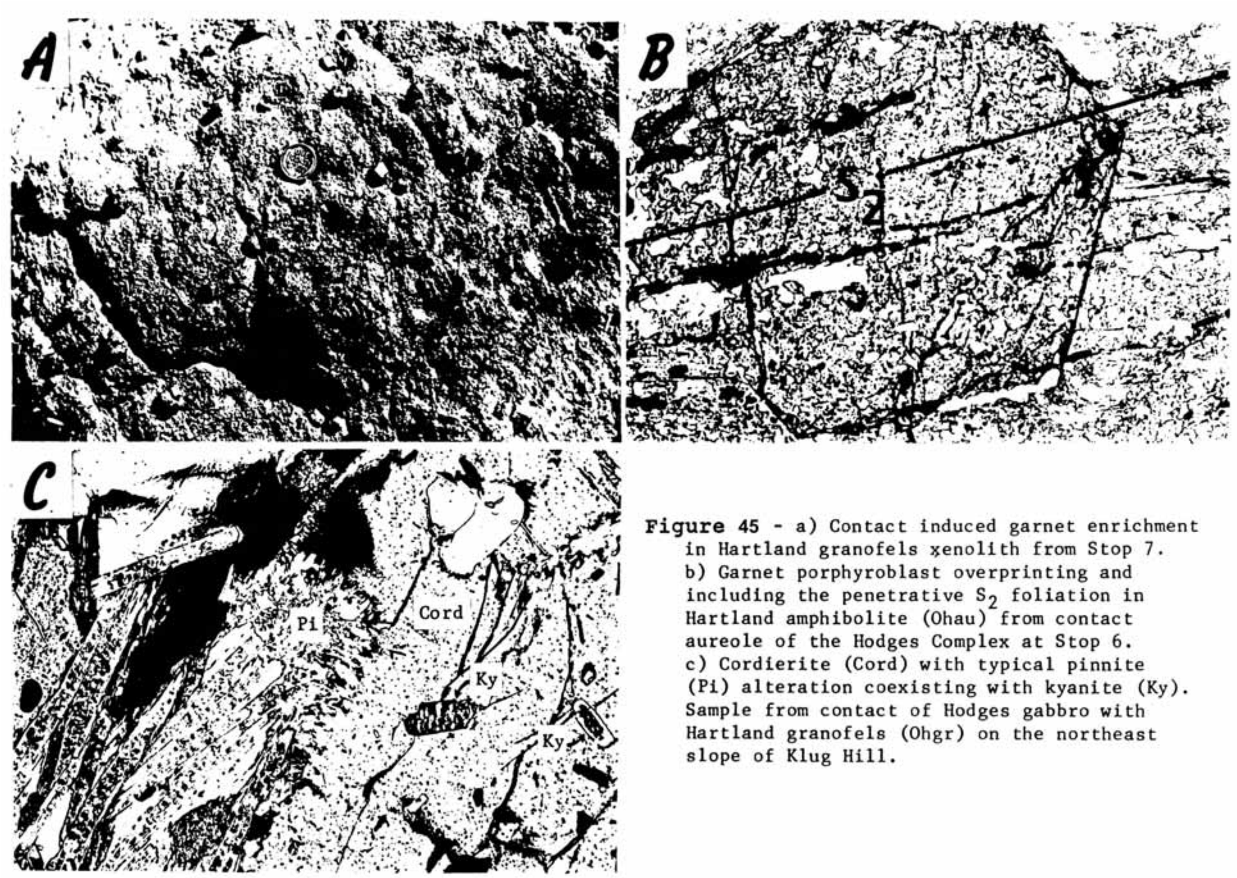


Figure 45. Contact-metamorphic minerals in the wall rocks of the Hodges Complex. A) Result of contact-induced garnet enrichment in Hartland granofels xenolith from Stop 7. B) Garnet porphyroblast overprinting and including the penetrative S2 foliation in Hartland amphibolite (Ohau) from contact aureole of the Hodges Complex at Stop 6. C) Cordierite (Cord) with typical pinnite (Pi) alteration coexisting with kyanite (Ky). Sample from contact of Hodges Gabbro with Hartland Granofels (Ohgr) on the northeast slope of Klug Hill. (Charles Merguerian, 1985, fig. 12, p. 436.)

Table 4 lists the contact assemblages in the wall rocks of the Hodges Complex compared to regional assemblages outside the aureole. Randomly oriented contact phases, the assemblage cordierite-kyanite-staurolite-garnet, and the absence of muscovite are characteristics of the Hodges aureole. These traits indicate that the Hodges was statically intruded at pressures lying between 5 and 8 kb (corresponding to a depth of burial of between 20 and 25 km) and at temperatures in the range of 675° to 700°C, near the Al_2SiO_5 triple point (Merguerian, 1977).

There is no clear overprinting of contact minerals by M_2 kyanite, staurolite, or garnet porphyroblasts. Rather, identical minerals formed in the contact aureole suggesting that intrusion of the Hodges and regional M_2 metamorphism were coeval (Table 3). Local temperature increases adjacent to the Hodges Complex produced cordierite +/- sillimanite and fostered the breakdown of muscovite.

17.2 0.0 Drive back to Route 4 and turn left. Follow Route 4 east to Lovers Lane.
18.2 1.0 Turn right onto Lovers Lane and make another immediate right into Ducci Electrical Contracting Co. parking lot.
18.3 0.1 The Tyler Lake granite crops out in the creek bed along the south edge of the lot.

STOP 8 - Tyler Lake Granite. [UTM Coordinates: 653.00E / 4631.34N, West Torrington quadrangle.]

Near the creek bed is exposed tan-weathering, medium-grained, foliated quartz-microcline-plagioclase-muscovite-biotite-garnet-(chlorite)-(apatite) granite. An X-ray-fluorescence analysis by Dr. D. Radcliffe of Hofstra University produced the following result: SiO_2 = 73.0, Al_2O3 = 14.2, Fe_2O3 = 1.4, MgO = 0.6, CaO = 0.8, K_2O = 5.5, Na_2O = 3.1, TiO_2 = 0.2, MnO = 0.1, loss on ignition = 0.7 (total = 99.6). The granite has been foliated by cm-spaced micaceous layering (S_4) oriented N36°E, 60°NW. The S_4 foliation is cut by a faint slip cleavage (S_6?) oriented N10°E, 42°SW.

The Tyler Lake Granite contains xenoliths of the Hodges rocks and is in direct contact with all major metamorphic units in the area (except Є-Ohmk and Є-Oha). This suggests a young intrusive age. Widely separated sample suites from the granite yield a well defined 466 +/- 12 Ma Rb-Sr isochron with initial Sr 87/86 = 0.7082 +/- 0.0011 (Merguerian and others, 1984). Because the Hodges was intruded following or nearly synchronously with D_2, this mid-Ordovician age is proof of a Taconian or possibly older age for Cameron's Line. The Sr 87/86 data imply that the Tyler Lake Granite was derived either from anatectic melting of the continental crust or materials derived from the crust (i.e., Waramaug and Hartland sequences). Assimilation may have also been an important process during intrusion of the granite. The Hodges Complex may have been emplaced during the late stages of the Taconic orogeny when oversteepened subduction beneath a thick orogenic welt tapped mafic- and ultramafic magmas that ascended along Cameron's Line.

ACKNOWLEDGMENTS

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TABLES

Table 01 - GEOLOGIC TIME CHART

(with selected major geologic events from southeastern New York and vicinity)

ERA Periods (Epochs)	**Years (Ma)**	**Selected Major Events**
CENOZOIC		
(Holocene)	0.1	Rising sea forms Hudson Estuary, Long Island Sound, and other bays. Barrier islands form and migrate.
(Pleistocene)	1.6	Melting of last glaciers forms large lakes. Drainage from Great Lakes overflows into Hudson Valley. Dam at The Narrows suddenly breached and flood waters erode Hudson shelf valley. Repeated continental glaciation with five? glaciers flowing from NW and NE form moraine ridges on Long Island.
(Pliocene)	6.2	Regional uplift, tilting and erosion of coastal-plain strata; sea level drops. Depression eroded that later becomes Long Island Sound.
(Miocene)	26.2	Fans spread E and SE from Appalachians and push back sea. Last widespread marine unit in coastal-plain strata.
MESOZOIC	66.5	
(Cretaceous)		
	96	Passive eastern margin of North American plate subsides and sediments (the coastal-plain strata) accumulate.
	131	(**Begin Atlantic Passive-Margin Stage II**).
(Jurassic)		Baltimore Canyon Trough forms and fills with 8,000 feet of pre-Cretaceous sediments.
		Atlantic Ocean starts to open. Newark basins deformed, arched, eroded.
(Triassic)	190	Continued filling of subsiding Newark basins and mafic igneous activity both extrusive and intrusive. Newark basins form and fill with non-marine sediments.

PALEOZOIC	245	
(Permian)		Pre-Newark erosion surface formed.
	260	**Appalachian orogeny.** (Terminal stage.) Folding, overthrusting, and metamorphism of Rhode Island coal basins; granites intruded.
(Carboniferous)		Faulting, folding, and metamorphism in New York City area. Southeastern New York undergoes continued uplift and erosion.
(Devonian)	365	**Acadian orogeny.** Deep burial of sedimentary strata. Faulting, folding, and metamorphism in New York City area. Peekskill Granite and Acadian granites intruded.
(Silurian)		
	440	**Taconic orogeny.** Intense deformation and metamorphism.
	450	Cortlandt Complex and related rocks intrude Taconian suture zone. (Cameron's Line and St. Nicholas Thrusr Zone). Arc-continent collision. Great overthrusting from ocean toward continent. Taconic deep-water strata thrust above shallow-water strata.
(Ordovician)		Ultramafic rocks (oceanic lithosphere) sliced off and transported structurally above deposits of continental shelf. Shallow-water clastics and carbonates accumulate in west of basin. (= **Sauk Sequence**; protoliths of the Lowerre Quartzite, Inwood
(Cambrian)		Marble, Walloomsac Schist). Transitional slope/rise sequence (Manhattan Schist) and deep-water terrigenous strata (Hartland Formation) form to east. (= **Taconic Sequence**).
PROTEROZOIC		
	570	Period of uplift, rifting,and erosion followed by subsidence of margin and development of **Iapetan Passive-Margin Stage I**.
(Z)	600	Rifting with rift sediments, volcanism, and intrusive activity. (Ned Mountain, Pound Ridge, and Yonkers gneiss protoliths).
(Y)	1100	**Grenville orogeny.** Sediments and volcanics deposited, compressive deformation, intrusive activity, and granulite facies metamorphism. (Fordham Gneiss, Hudson Highlands and related rocks).
ARCHEOZOIC		
	2600	No record in New York.
	4600	Solar system (including Earth) forms.

Table 02 - Generalized Descriptions of Major Geologic "Layers", SE New York State and Vicinity

This geologic table is a tangible result of the On-The-Rocks Field Trip Program conducted by Drs. John E. Sanders and Charles Merguerian between 1988 and 1998. In Stenoan and Huttonian delight, we here present the seven layer cake model that has proved so effective in simplifying the complex geology of the region. Under continual scrutiny and improvement, we provide this updated web-based information as a public service to all students and educators of geology. We encourage any comments, additions, or corrections. References cited can be sought by following this link.

LAYER VII - QUATERNARY SEDIMENTS

A blanket of irregular thickness [up to 50 m or more] overlying and more or less covering all older bedrock units. Includes four or five tills of several ages each of which was deposited by a continental glacier that flowed across the region from one of two contrasting directions: (1) from N10°E to S10°W (direction from Labrador center and down the Hudson Valley), or (2) from N20°W to S20°E (direction from Keewatin center in Hudson's Bay region of Canada and across the Hudson Valley). The inferred relationship of the five tills is as follows from youngest [I] to oldest [V]. [I] - Yellow-brown to gray till from NNE to SSW, [II] - red-brown till from NW to SE, [III] - red-brown till from NW to SE, and [IV] - yellow-brown to gray till from NNE to SSW, and [V] - red-brown till from NW to SE containing decayed stones (Sanders and Merguerian, 1991a,b, 1992, 1994a, b; Sanders, Merguerian, and Mills, 1993; Sanders and others, 1997; Merguerian and Sanders, 1996). Quaternary sediments consist chiefly of till and outwash. On Long Island, outwash (sand and gravel) and glacial lake sediment predominates and till is minor and local. By contrast, on Staten Island, tills and interstratified lake sediments predominate and sandy outwash appears only locally, near Great Kills beach.

[Pliocene episode of extensive and rapid epeirogenic uplift of New England and deep erosion of major river valleys, including the excavation of the prominent inner lowland alongside the coastal-plain cuesta; a part of the modern landscape in New Jersey, but submerged in part to form Long Island Sound].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER VI - COASTAL-PLAIN STRATA (L. Cretaceous to U. Miocene; products of Passive Continental Margin II - Atlantic).

Marine- and nonmarine sands and clays, present beneath the Quaternary sediments on Long Island (but exposed locally in NW Long Island and on SW Staten Island) and forming a wide outcrop belt in NE New Jersey. These strata underlie the submerged continental terrace. The basal unit (L. Cretaceous from Maryland southward, but U. Cretaceous in vicinity of New York City) overlaps deformed- and eroded Newark strata and older formations. Also includes thick (2000 m) L. Cretaceous sands and shales filling the offshore Baltimore Canyon Trough. At the top are Miocene marine- and coastal units that are coarser than lower strata and in many localities SW of New Jersey, overstep farther inland than older coastal-plain strata. Capping unit is a thin (<50 m) sheet of yellow gravel (U. Miocene or L. Pliocene?) that was prograded as SE-directed fans from the Appalachians pushed back the sea. Eroded Newark debris is present in L. Cretaceous sands, but in U. Cretaceous through Miocene units, Newark-age redbed debris is conspicuously absent. This relationship is considered to be proof that the coastal-plain formations previously buried the Newark basins so that no Newark-age debris was available until after the Pliocene period of great regional uplift and erosion. The presence of resistant heavy minerals derived from the Proterozoic highlands part of the Appalachians within all coastal-plain sands indicates that the coastal-plain strata did not cover the central highlands of the Appalachians.

[Mid-Jurassic to Late Jurassic episode of regional arching of Newark basin-filling strata and end of sediment accumulation in Newark basin; multiple episodes of deformation including oroclinal "bending" of entire Appalachian chain in NE Pennsylvania (Carey, 1955), and one or more episodes of intrusion of mafic igneous rocks, of folding, of normal faulting, and of strike-slip faulting (Merguerian and Sanders, 1994b). Great uplift and erosion, ending with formation of Fall-Zone planation surface].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER V - NEWARK BASIN-FILLING STRATA (Upper Triassic and Lower Jurassic)

Newark-age strata unconformably overlie folded- and metamorphosed Paleozoic strata of Layer II and some of the Proterozoic formations of Layer I; are in fault contact with other Proterozoic formations of the Highlands complex. Cobbles and boulders in basin-marginal rudites near Ramapo Fault include mostly rocks from Layers III, IIB, and IIA(W), which formerly blanketed the Proterozoic now at the surface on the much-elevated Ramapo Mountains block. The thick (possibly 8 or 9 km) strata filling the Newark basin are nonmarine.

In addition to the basin-marginal rudites, the sediments include fluvial- and varied deposits of large lakes whose levels shifted cyclically in response to climate cycles evidently related to astronomic forcing. A notable lake deposit includes the Lockatong Formation, with its analcime-rich black argillites, which attains a maximum thickness of about 450 m in the Delaware River valley area. Interbedded with the Jurassic part of the Newark strata are three extrusive complexes, each 100 to 300 m thick, whose resistant tilted edges now underlie the curvilinear ridges of the Watchung Mountains in north-central New Jersey. Boulders of vesicular basalt in basin-marginal rudites prove that locally, the lava flows extended northwestward across one or more of the basin-marginal faults and onto a block that was later elevated and eroded. The thick (ca. 300 m) Palisades intrusive sheet is concordant in its central parts, where it intrudes the Lockatong at a level about 400 m above the base of the Newark strata. To the NE and SW, however, the sheet is discordant and cuts higher strata (Merguerian and Sanders, 1995a). Contact relationships and the discovery of clastic dikes at the base of the Palisades in Fort Lee, New Jersey, suggest that the mafic magma responsible for the Palisades was originally intruded at relatively shallow depths (roughly 3 to 4 km) according to Merguerian and Sanders (1995b).

Xenoliths and screens of both Stockton Arkose and Lockatong Argillite are present near the base of the sill. Locally, marginal zones of some xenoliths were melted to form granitic rocks (examples: the trondhjemite formed from the Lockatong Argillite at the Graniteville quarry, Staten Island, described by Benimoff and Sclar, 1984; and a "re-composed" augite granite associated with pieces of Stockton Arkose at Weehawken and Jersey City, described by J. V. Lewis, 1908, p. 135-137).

[**Appalachian terminal orogeny**; large-scale overthrusts of strata over strata (as in the bedding thrusts of the "Little Mountains east of the Catskills" and in the strata underlying the NW side of the Appalachian Great Valley), of basement over strata (in the outliers NW of the Hudson Highlands, and possibly also in many parts of the Highlands themselves), and presumably also of basement over basement (localities not yet identified). High-grade metamorphism of Coal Measures and intrusion of granites in Rhode Island dated at 270 Ma. Extensive uplift and erosion, ending with the formation of the pre-Newark peneplain].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER IV - COAL MEASURES AND RELATED STRATA (Carboniferous)

Mostly nonmarine coarse strata, about 6 km thick, including thick coals altered to anthracite grade, now preserved only in tight synclines in the Anthracite district, near Scranton, NE Pennsylvania; inferred to have formerly extended NE far enough to have buried the Catskills and vicinity in eastern New York State (Friedman and Sanders, 1982, 1983).

[**Acadian orogeny**; great thermal activity and folding, including metamorphism on a regional scale, ductile deformation, and intrusion of granites; dated at ~360 Ma].

LAYER III - MOSTLY MARINE STRATA OF APPALACHIAN BASIN AND CATSKILLS (Carbonates and terrigenous strata of Devonian and Silurian age)

(Western Facies)	**(Eastern Facies)**
Catskill Plateau, Delaware Valley monocline, and "Little Mountains" NW of Hudson-Great Valley lowland.	SE of Hudson-Great Valley lowland in Schunnemunk-Bellvale graben.
Kaaterskill redbeds and cgls.	Schunnemunk Cgl.
Ashokan Flags (large cross strata)	Bellvale Fm., upper unit
Mount Marion Fm. (graded layers, marine)	Bellvale Fm., lower unit (graded layers, marine)
Bakoven Black Shale	Cornwall Black Shale
Onondaga Limestone	

Schoharie buff siltstone	Pine Hill Formation
Esopus Formation	Esopus Formation
Glenerie Chert	
Connelly Conglomerate	Connelly Conglomerate
Central Valley Sandstone	
Carbonates of Helderberg Group	Carbonates of Helderberg Group
Manlius Limestone	
Rondout Formation	Rondout Formation
Decker Formation	
Binnewater Sandstone	Poxono Island Formation
High Falls Shale	Longwood Red Shale
Shawangunk Formation	Green Pond Conglomerate

[**Taconic orogeny**; 480 Ma deep-seated folding, dynamothermal metamorphism and mafic- to ultramafic (alkalic) igneous intrusive activity (dated in the range of 470 to 430 Ma) across suture zone (Cameron's Line-St. Nicholas thrust zones). Underthrusting of shallow-water western carbonates of Sauk Sequence below supracrustal deep-water eastern Taconic strata and imbrication of former Sauk-Tippecanoe margin. Long-distance transport of strata over strata has been demonstrated; less certain locally is proof of basement thrust over strata and of basement shifted over basement. In Newfoundland, a full ophiolite sequence, 10 km thick, has been thrust over shelf-type sedimentary strata]. In NYC and throughout New England, dismembered ophiolite is common (Merguerian 1979; Merguerian and Moss 2005, 2007).

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER II - CAMBRO-ORDOVICIAN CONTINENTAL-MARGIN COVER (Products of Passive Continental Margin I - Iapetus). Subdivided into two sub layers, IIB and IIA. Layer IIA is further subdivided into western- and eastern facies.

LAYER IIB - TIPPECANOE SEQUENCE - Middle Ordovician flysch with basal limestone (Balmville, Jacksonburg limestones).

Not metamorphosed / Metamorphosed
Martinsburg Fm. / Walloomsac Schist
Normanskill Fm. / Annsville Phyllite

Subaerial exposure; karst features form on Sauk (Layer IIA[W]) platform.

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER IIA [W] - SAUK SEQUENCE	LAYER IIA [E] - TACONIC SEQUENCE
Western shallow-water platform (L. Cambrian - M. Ordovician)	**Eastern deep-water zone** (L. Cambrian-M. Ordovician)
Copake Limestone (Stockbridge,	
Rochdale Limestone (Inwood Marble)	
Halcyon Lake Fm.	
Briarcliff Dolostone	(Ꞓ-Oh) Hartland Fm.
Pine Plains Fm.	(Ꞓ-Om) Manhattan Fm.
Stissing Dolostone	
Poughquag Quartzite	
Lowerre Quartzite	
Ned Mtn Fm.	

[**Pre-Iapetus Rifting Event**; extensional tectonics, volcanism, rift-facies sedimentation, and plutonic igneous activity precedes development of Iapetus ocean basin. Extensional interval yields protoliths of Pound Ridge Gneiss, Yonkers granitoid gneisses, and the Ned Mountain Formation (Brock, 1989, 1993). In New Jersey, metamorphosed rift facies rocks are mapped as the Chestnut Hill Formation of A. A. Drake, Jr. (1984)].

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

LAYER I - PROTEROZOIC BASEMENT ROCKS

Many individual lithologic units including Proterozoic Z and Y ortho- and paragneiss, granitoid rocks, metavolcanic- and metasedimentary rocks identified, but only a few attempts have been made to decipher the stratigraphic relationships; hence, the three-dimensional structural relationships remain obscure. Followed by a period of uplift and erosion.

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

[**Grenville orogeny**; deformation, metamorphism, and plutonism dated about 1,100 Ma. After the orogeny, an extensive period of uplift and erosion begins. Grenville-aged (Proterozoic Y) basement rocks include the Fordham Gneiss of Westchester County, the Bronx, and the subsurface of western Long Island (Queens and Brooklyn Sections, NYC Water Tunnel #3), the Hudson Highland-Reading Prong terrane, the Franklin Marble Belt and associated rocks, and the New Milford, Housatonic, Berkshire, and Green Mountain Massifs.]

~~~~~~~~~~~~~~~~~~~~Surface of unconformity~~~~~~~~~~~~~~~~~~~~

In New Jersey and Pennsylvania rocks older than the Franklin Marble Belt and associated rocks include the Losee Metamorphic Suite. Unconformably beneath the Losee, in Pennsylvania, Proterozoic X rocks of the Hexenkopf Complex crop out.

Table 03 - Linear- and planar structural features and chronology of folding, igneous activity, and metamorphism in the vicinity of Torrington, Connecticut (Merguerian, 1985, Table 1, p. 420.)

DEFORMATIONAL EVENT	LINEAR FEATURES	PLANAR FEATURES	IGNEOUS ACTIVITY	METAMORPHISM
D_1	F_1 isoclinal folds of compositional layering. L_1 quartz ribbing in gneisses and schists. Hornblende lineation in amphibolites.	S_1 gneissic layering in gneisses or hornblende-plagioclase foliation in amphibolites. Generally not recognized in schists.		? Amphibolite-grade M_1
D_2	F_2 penetrative isoclinal folds of early S_1 structures and compositional layering. L_2 mineral streaking in schists and gneisses.	S_2 regional foliation composed of oriented phyllosilicates±kyanite or sillimanite developed axial planar to F_2 folds.	? HODGES COMPLEX TYLER LAKE GRANITE *466±12 m.y.*	Amphibolite-grade M_2
D_3	F_3 shallow SW to NW plunging, open to tight, crenulate folds of the S_2 regional foliation. L_3 intersection lineation in massive rocks; crinkle axis in micaceous rocks.	S_3 crenulation or slip cleavage developed axial planar to F_3 folds. Oriented NW to WSW with shallow dips.	PEGMATITES	?
D_4	F_4 steep SW plunging dextral synformal folds of the S_2 regional foliation	S_4 crenulation cleavage, slip cleavage, or spaced schistosity developed axial planar to F_4 folds Orientation - N20°E, 75°NW.		Biotite-grade M_3 (retrograde)
D_5	F_5 open folds and warps with variable hingelines. L_5 intersection lineation.	S_5 slip cleavage and rock cleavage axial planar to F_5 folds oriented NW to W with variable dip.	?	continued retrograde

Table 04 - Contact-metamorphic mineral assemblages in the country rocks near the Hodges Complex compared with the regional assemblage outside the contact aureole. (Merguerian, 1977 ms., Table 3, p. 77.)

SAMPLE	UNIT	CONTACT ASSEMBLAGE	REGIONAL ASSEMBLAGE AWAY FROM CONTACT	REMARKS
H-169	Ohgr	cord-ky-st-gt-bi-plag-qtz-chl	bi-musc-qtz-plag±ky	Contact with gabbro on Klug Hill. Sample was 5 feet from contact.
H-31	Ohgk	cord-ky-st-bi-qtz	musc-bi-qtz-plag-ky±st	Contact of gabbroic and ultra-mafic rocks with a screen of Hartland rocks. From the SW of the intersection of Weed Road and Route 4.
H-68	Ohc	st-gt-qtz-bi-sill	qtz-musc-plag	
H-69	Ohgk	cord-ky-gt-st-sill-bi-qtz-chl	musc-bi-qtz-plag-ky-±st	
H-30b	Ohau	hyp-hb-plag-bi-op	hb-plag-bi-op	
H-43	p€-Owg	gt-plag-qtz-bi	gt-plag-qtz-bi	From contact of Waramaug and dio-rite NNW of the Hodges Complex.
H-52A	p€-Owg	ky-bi-qtz-plag-gt	musc-gt-ky-qtz-plag	
H-116	Ohc	grun-qtz	musc-bi-qtz-plag	From contact with diorite N of Route 4 and Klug Hill Road.
H-56	p€-Owga	hb-plag-bi-gt-op	hb-plag-bi-op	Direct from contact with diorite N of the Hodges Complex.
H-36	Ohau	hb-plag-bi-gt-op	hb-plag-bi-op	Direct from contact with ultra-mafic rocks N of the Hodges Nickel Prospect.

KEY:

cord = cordierite
ky = kyanite
st = staurolite
gt = garnet
bi = biotite
plag = plagioclase
qtz = quartz
chl = chlorite
sill = sillimanite
hyp = hypersthene
hb = hornblende
op = opaques
grun = grunerite

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