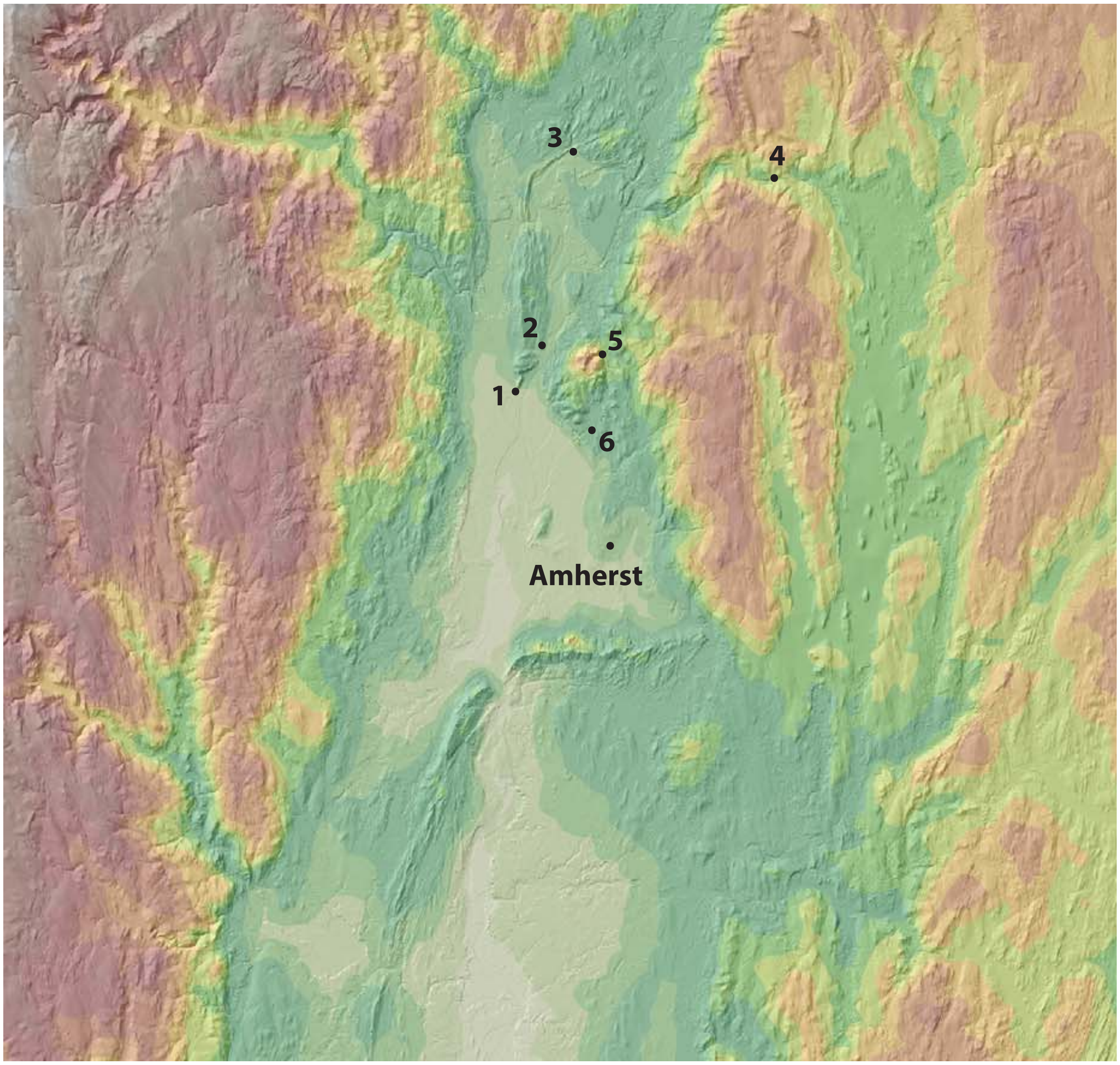
Connecticut Valley Field Trip
Keck Symposium 2006

John Brady and Jack Cheney



Stop 1 – Mt. Sugarloaf

We will ascend Mt. Sugarloaf along the access road, which cannot accommodate our busses. Mt. Sugarloaf is the type locality for the Sugarloaf Arkose, a Triassic sandstone, described by Paul Olsen (1992) as a "crudely bedded, poorly sorted arkosic sandstone and conglomeratic sandstone." The walk to the top will provide plenty of opportunity for you to view this rock. See if you can find bedding. What are the clasts in the conglomeratic sandstones? Can you find cross-bedding? Stevens and Hubert (1980) published a mean transport direction of 288° in this area. Sediment transport from east to west is supported throughout the Mesozoic Basin by both current indicators and by a fining of grain size from east to west. Olsen (1992) suggests that the dearth of sedimentary structures may be due to intensive bioturbation. Stevens and Hubert (1980) argue for braided stream deposition, which is consistent with the red (oxidized iron) coloration of the rocks.

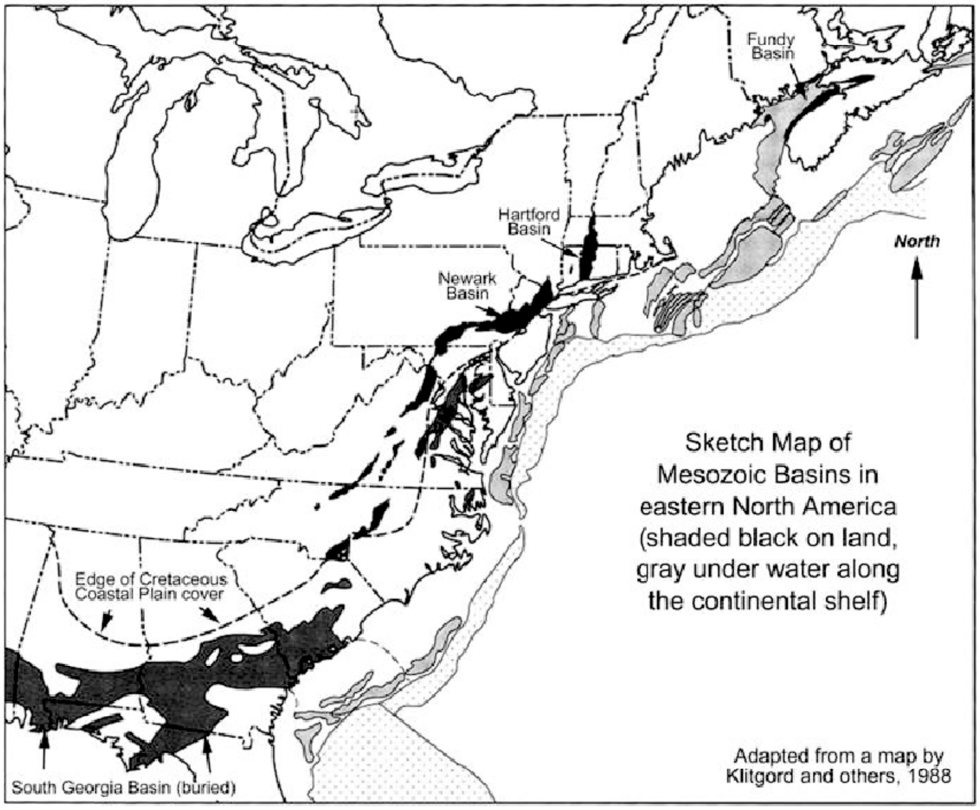
From the top of Mt. Sugarloaf, you can get a panoramic view of the Connecticut River Valley. To the east is Mt. Toby and the Pelham Hills. To the west are the Berkshire Hills. To the south is the Holyoke Range running east-west across this north-south valley. These hills are upheld by a basalt caprock that is more resistant to weathering than the Mesozoic sediments beneath. The visible asymmetry of the Mt. Tom Range displays the eastward dip of the basalt and the bedding of the sedimentary layers it protects. The Holyoke Range is also asymmetric, but its basalt flow dips to the south and the profile is not visible from here. The large fold defined by the two ranges can be seen by comparing the eastern most part of the Holyoke Range with Mt. Tom.

The valley is broad and flat, shaped by erosion that is controlled by the contrast of the softer Mesozoic sedimentary rocks of the valley and the harder Paleozoic metamorphic and igneous rocks that surround it. Thousands of feet of erosion have contributed to the development of the elevation contrasts we see today. Indeed, the path of the Connecticut River cuts the basalt layer that was once continuous between Mt. Tom and Mt. Holyoke, a feat for a superposed stream whose origin predates the emergence of the two ranges. Pleistocene glaciation has helped with the erosion, smoothing and grooving the rocks atop Mt. Sugarloaf and throughout the region.

One should ask why a linear basin developed here in the Mesozoic. Indeed, why are there similar Mesozoic basins all along the east coast (see Figure 1.2)? The best explanation is normal faulting associated with the break up of Pangea and the opening of the Atlantic Ocean. The absence of rocks older than about 200 Ma on the Atlantic sea floor provides strong support for this model. If true, we are here today because of the rifting that occurred over 200 Ma ago.

Perhaps the most striking feature of the valley is its flatness. The valley slopes to the south only about 0.5 meter per km, so flat that the postglacial Connecticut river has developed meanders and oxbow lakes. The level valley is one consequence of a post-glacial lake that occupied the valley for many years as the ice sheet melted. 30-40 meters of sediment accumulated in this lake, smoothing the valley and burying the troublesome glacial boulders that characterize the landscape elsewhere in New England. With no boulders to move, there are no stone walls in the valley, also an anomaly for the region. If you look to the east and slightly south, you can see a broad flat field that is substantially above the level of the valley floor. That field is planted on the Sunderland Delta formed along the edge of Glacial Lake Hitchcock, which we will visit at the end of our day.

Figure 1.1. View of Sunderland from Mt. Sugarloaf during the 1936 flood (taken by Smith Geology Professor Robert Collins).



Sketch Map of Mesozoic Basins in eastern North America (shaded black on land, gray under water along the continental shelf)

Adapted from a map by Klitgord and others, 1988

Figure 1.2. The Connecticut Valley is one of many Mesozoic basins with similar features that line the east coast. Figure modified from Olsen et al. (2003).

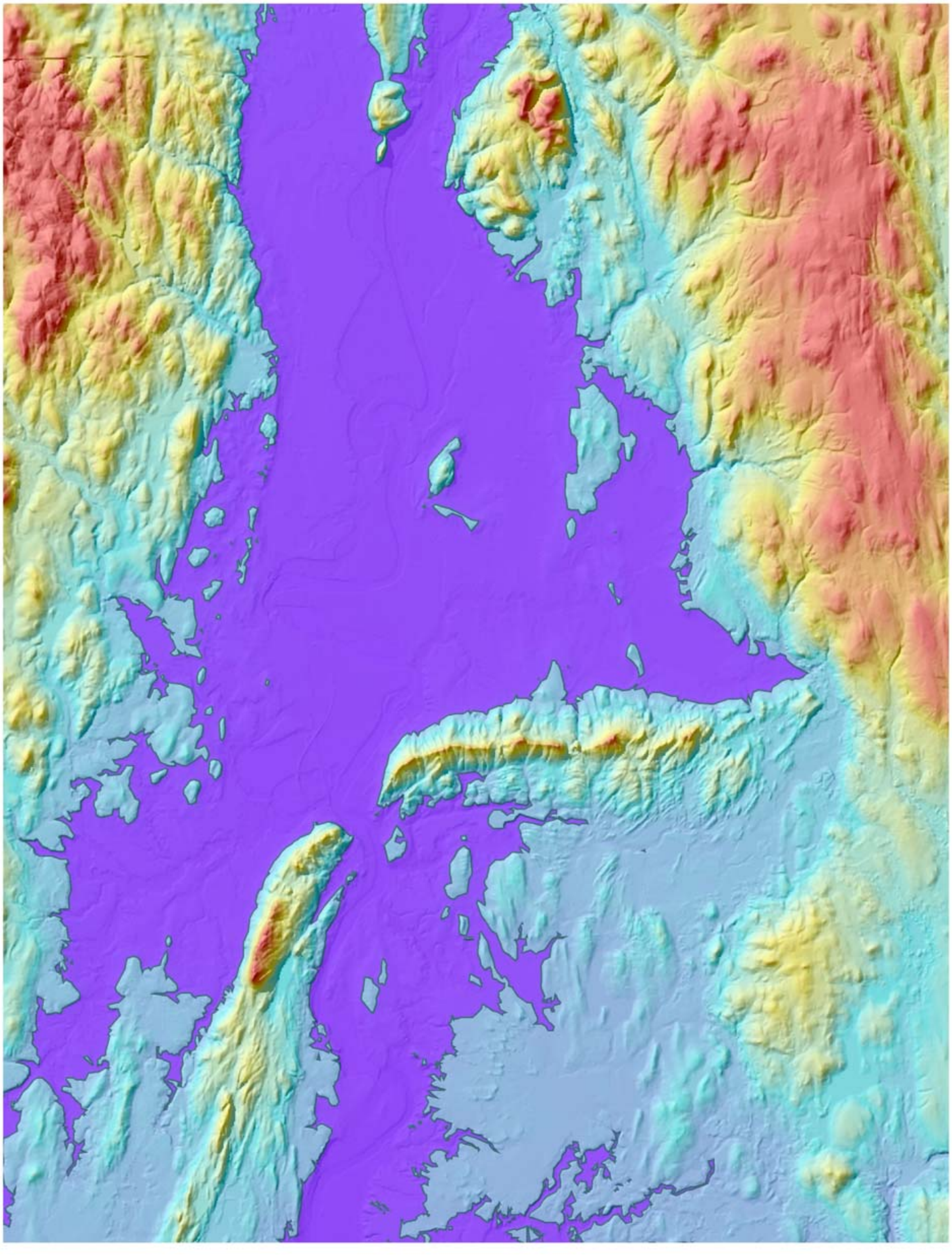


Figure 1.3. Lake Hitchcock based on delta elevations (created by Bob Newton).

Stop 2 – Lake Hitchcock Sediment along River Road

As the last continental glacier retreated north from Massachusetts, approximately 15,000 years ago, a lake formed within the Connecticut Valley (Figure 2.1). At its maximum, the lake extended from just north of Middletown, Connecticut to St. Johnsbury, Vermont. The former presence of a lake was first recognized in 1818 by Edward Hitchcock, a geologist who began teaching at Amherst College in 1825 and became its third president (1845-54). The lake was named Lake Hitchcock by Richard Lougee (1935), who studied the lake features while teaching at Colby College. The lake was confined behind a dam composed of unconsolidated sands and gravels deposited by meltwater streams and was, therefore, weak and unstable. However, a spillway for the lake was cut across bedrock in New Britain Connecticut, so the dam was able to survive for over a thousand years before it failed, catastrophically draining the lake.

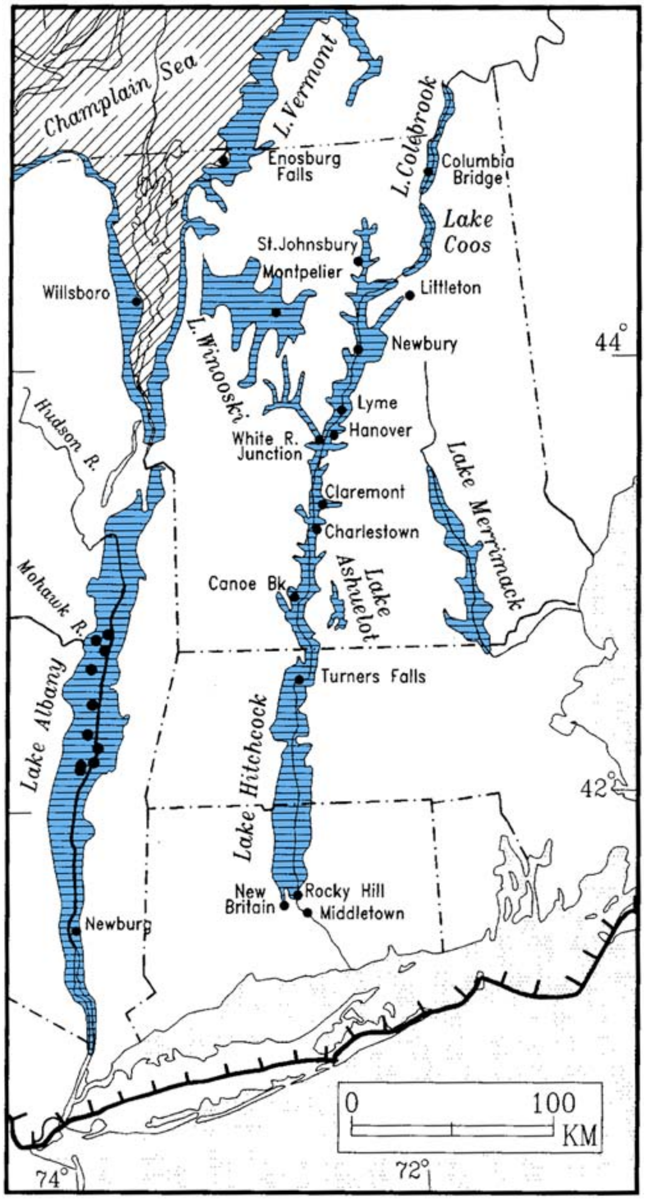


Figure 2.1. Major glacial lakes of New England (modified from Ridge et al., 1999).

During the time the lake occupied the valley tremendous amounts of fine-grained sediment were deposited in the lake bottom. For example, a 32-meter-long core of Lake Hitchcock sediment was retrieved from a drill hole on the athletic fields at UMass in 1997 (Brigham-Grette and Rittenour, 2003). This lake-bottom material originated from sediment-rich meltwater streams draining from the retreating ice sheet. The meltwater streams were most active in the summer, transporting lots of coarser sediment into the lake. During the winter little additional sediment entered the lake so only fine clay-sized material accumulated on the lake bottom. Thus, these deposits are composed of annual alternation of coarse and fine sediment called varves.

Many lakes were created in New England as the ice melted. Ernst Antevs (1922, 1928) studied the varves in many locations and found that there was a pattern to varve thicknesses that could be correlated from place to place. Fitting the overlapping varve patterns from across the northeast, Antevs developed a "New England Varve Chronology" that is still in use today. Modern work, summarized recently by Ridge et al. (1999), has shown the relationship between Antev's Varve Chronology, atmospheric ^{14}C

ages, and U-Th calibrated ages (Figure 2.2). Based on Ridge's chronology, the UMass core represents deposition from 15.8 ka (varve 4638) to 14.4 ka (varve 6027 – a total of 1389 varves).

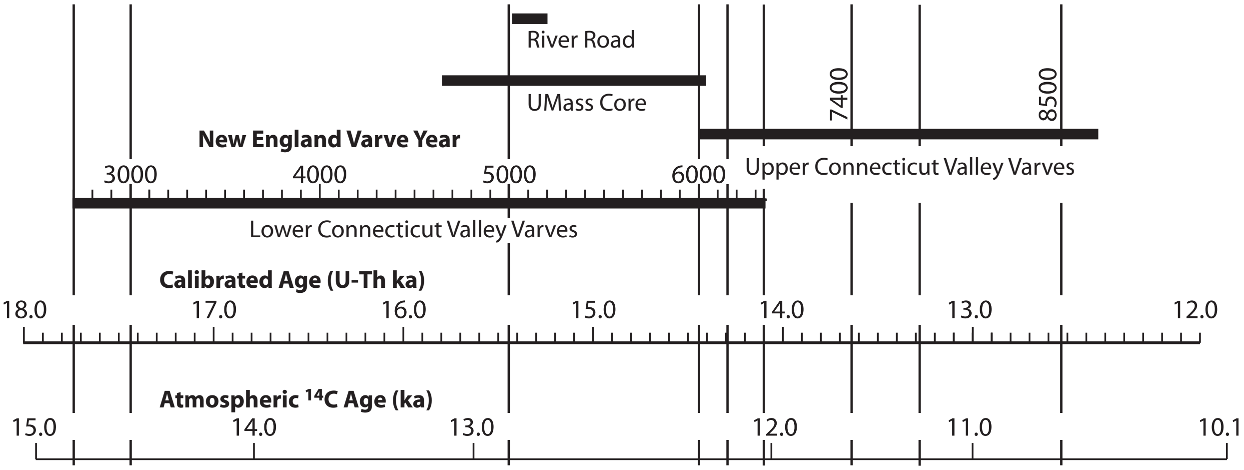


Figure 2.2. Calibrated ages of Antev's New England Varve Chronology (Ridge et al., 1999) along with varve numbers for the UMass core and River Road.

Lake Hitchcock clays are exposed at this stop (due to excavation by geologists) in a narrow gorge cut by a stream flowing eastward from the Pocumtuck Range into the Connecticut River. Note the varves, and the distorted beds with pebbles in them. The one distorted bed here includes 12 varves. Its origin is obscure: it may be a result of floating ice that dropped the pebbles, but then why the distorted varves? The distorted varves may have resulted from earthquakes (similar distorted varves are ascribed to earthquakes in Quebec), but then why the pebbles? The proximity to the Pocumtuck Range may offer a clue. These distorted beds likely resulted from subaqueous slumping, wherein nearshore pebbles and sand were entrained by the slump.

Figure 2.2. Varves at the River Road stop.

Approximately 200 varves have been counted at our stop on River Road. Carbonate concretions from River Road varve locality were recently studied by Laura Levy (1998), whose varve thickness measurements are shown in Figure 2.4. Jack Ridge (personal communication, 2006) drew this figure from Levy's data, showing how these varve thicknesses compare with those measured at two other Massachusetts varve localities.

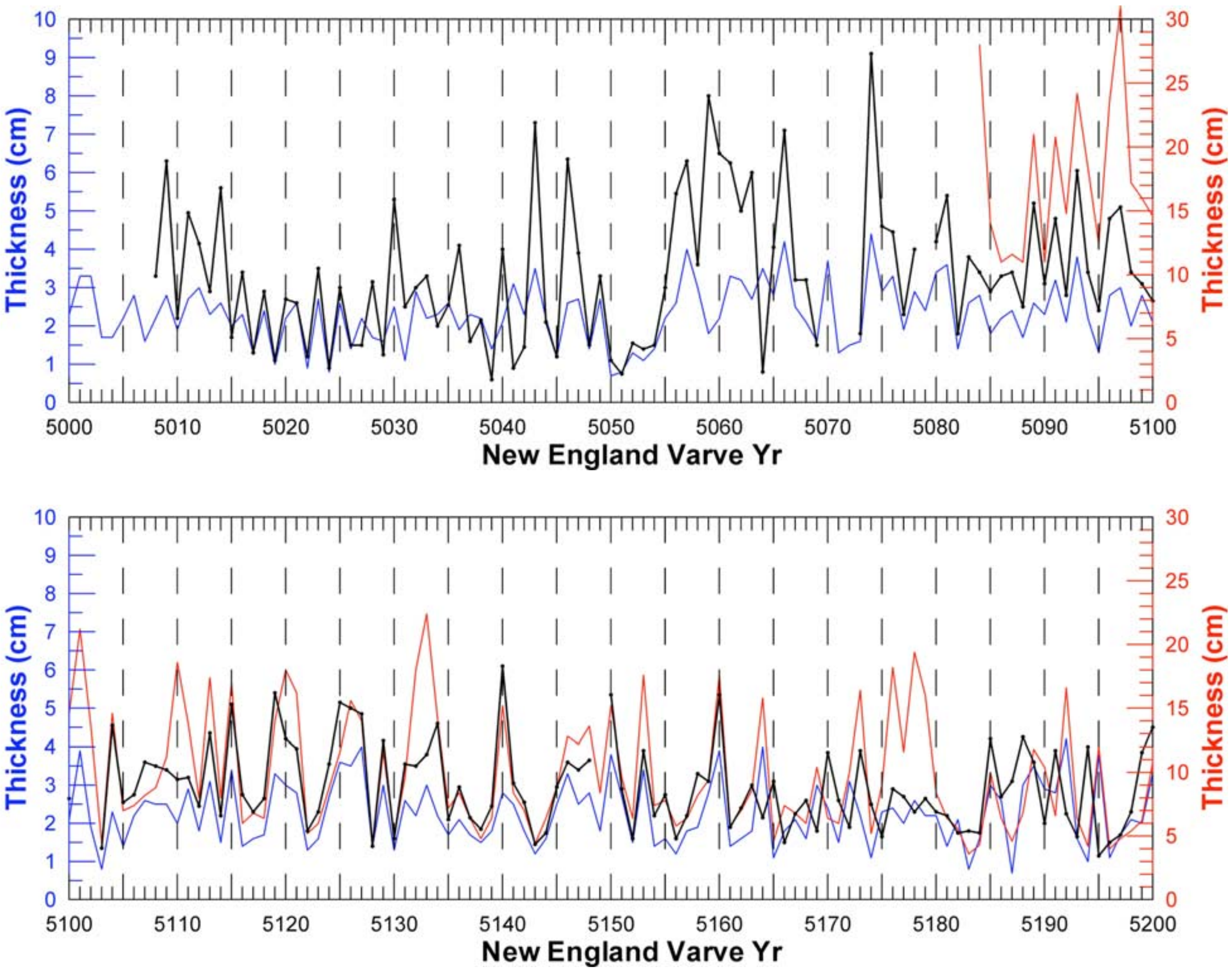


Figure 2.4. Varve thickness at River Road measured by Levy (1998) and correlated with two other Massachusetts varve localities by Ridge (personal communication, 2006). Blue curve data are from MAS37-53.DAT, Varve Years 3767-5222, Connecticut Valley, Mass., MA 4-12 (lower graph) of Antevs (1922) - from distal outcrops in southern and central Massachusetts. Red curve data are from MAS50-55., Varve Years 5084-5500, Connecticut Valley, Mass., MA 11-13 (upper graph), Antevs (1922) -from local ice-proximal outcrops.

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Stop 3 – Turners Falls

This is one of the most visited geologic sites in the Connecticut Valley Region. There is much to see here that reveals the character of this and other Mesozoic Basins in eastern North America. The following text and figures are taken from NEIGC guidebook descriptions by Olsen et al. (1992) and Wise et al. (1992). Paul Olsen and others believe that the stratigraphic sequence at Turners Falls and in other Mesozoic basin rocks can be related to Van Houten cycles, named for the scientist who first described them in the Newark Basin (Van Houten, 1964). Olsen et al. (1992) describe these cyles and the rocks as follows: "By analogy with precisely the same pattern of cycles in the Newark basin (Olsen, et al., 1989) the Van Houten cycles of Turners Falls were produced by the rise and fall of lakes controlled by climate cycles averaging about 20,000 years. The climate changes were controlled, in turn, by the precession of the equinoxes, modulated by the deformation of the orbit of the Earth. Van Houten cycles vary in the magnitude of the deepest water unit, forming larger cycles of ~100,000, 413,000, and ~ 2,000,000 years (Olsen, 1986; Olsen and Kent, 1990). The origin of these cycles is as old and persistent as the solar system itself. The precession of the equinoxes is produced by the gravitational pull of the Moon and Sun on the Earth's equatorial bulge, while the longer cycles are caused by deformation of the Earth's orbit by the attraction of the other planets to the Earth-Moon system. Ultimately, these celestial mechanical cycles influence the distribution of sunlight on the Earth's surface and thus control climate."

"Lake bed 0 is the wettest phase of the first ~ 100,000 year cycle in the Turners Falls Sandstone (Figure 3.1). Lake beds 1, 2 and 3 occur in the wettest phase of next 100,000 year cycle, and Lake bed 4 is the first of (probably) three lake beds marking out the next 100,000 year cycle. These upper two ~100,000 year cycles occur in the wettest phase of a 413,000 year cycle, while the lowest (with lake bed 0) is in the driest phase. The whole cyclical sequence from the Fall River beds though the exposures at the dam (two almost complete 413,000 year cycles) occur in the wettest phase of a ~2,000,000 year cycle. The same pattern occurs throughout the entire thickness of the Turners Falls Sandstone and Mt. Toby Conglomerate."

"The cycles seen at Turners Falls are almost certainly very laterally continuous, as has been demonstrated in other Newark basins (Olsen, et al., 1989). The black shales and gray limestones reported at Sunderland and at various other places along the Connecticut River in the Deerfield basin are almost certainly correlative with those at Turners Falls, although this has yet to be demonstrated. In addition, the climate cycles recorded at Turners Falls are exactly the same as cycles in at least the Hartford and Newark basins. That allows, along with the geochemical signature of the interbedded basalt sequences (Tollo, in Olsen, et al., 1989; Philpotts and Reichenbach, 1985) a precise correlation of individual cycles in the Newark Supergroup over a distance of at least 500 km (Olsen, et al., 1989)."

"Turners Falls is one of the premier fossil localities in the Connecticut Valley. Fish are most abundant in the microlaminated shale beds (preserved whole but flattened), and in the center of calcareous siltstone concretions (somewhat dissociated but more three-dimensional). Better preserved, more robust specimens are occasionally found in siltstone beds. All of the articulated fish found so far come from the dark gray to black portions of lake beds 2, 3, and 4

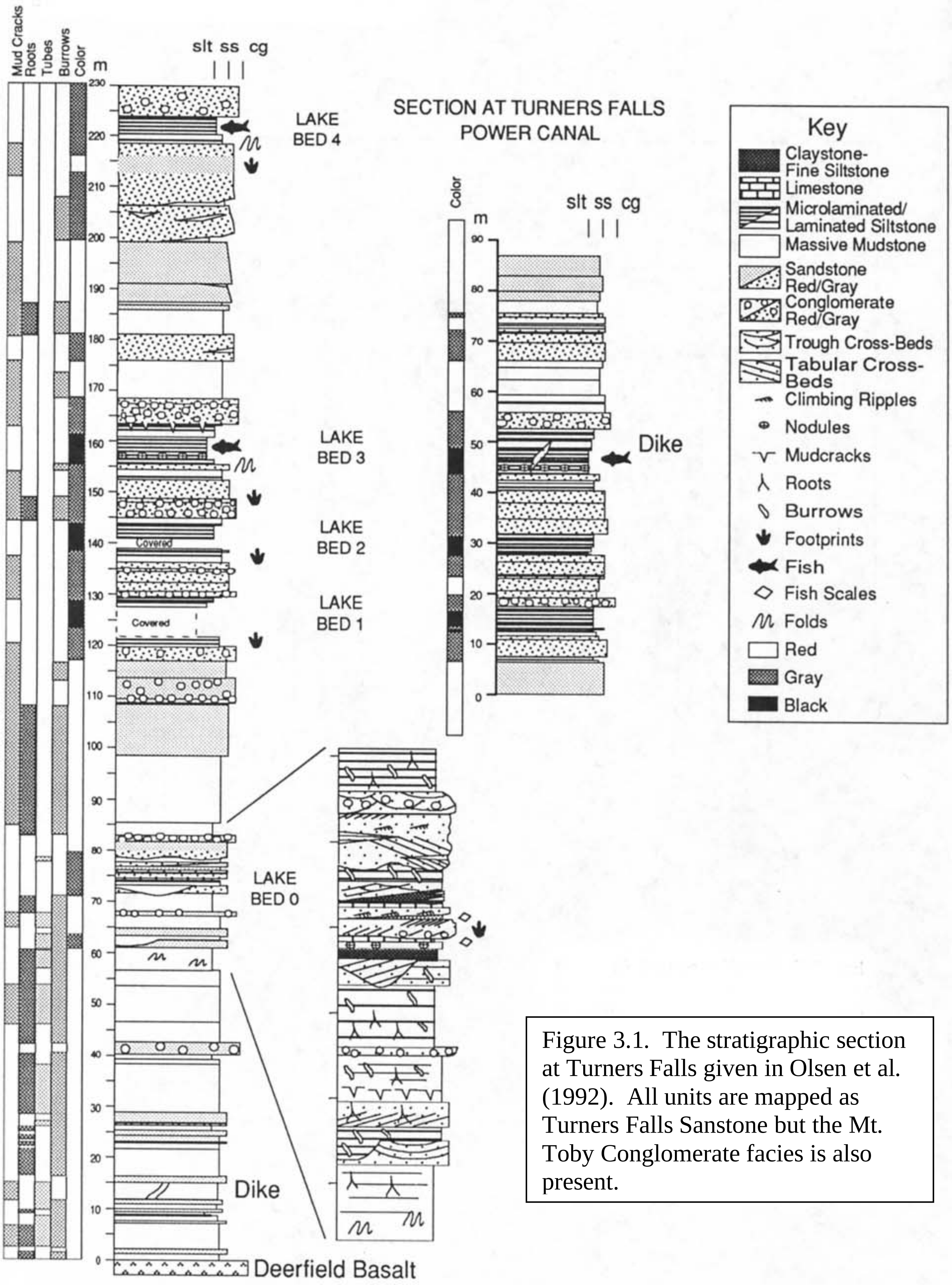


Figure 3.1. The stratigraphic section at Turners Falls given in Olsen et al. (1992). All units are mapped as Turners Falls Sanstone but the Mt. Toby Conglomerate facies is also present.

(Figure 3.1). By far the most common fish are semiontoids of the “Semionotus tenuiceps” and “small scale” groups of Olsen et al. (1982). The fish average 7-15 cm in length but can attain sizes up to 40 cm. Less common are the subholostean Redfieldius and the coelacanth Diplurus.”

“In the mid 1800’s Turners Falls was one of the most productive footprint localities in the Connecticut Valley, and was favorite of Edward Hitchcock and fellow collectors James Deane, Dexture Marsh, Roswell Field, and Timothy Stoughton. Tracks are now uncommon on the mainland, but the islands in the river occasionally yield fine specimens. In situ footprints are most common in transgressive portions of Van Houten cycles; less distinct examples are also present in the red beds. The most common ichnotaxa are Grallator (Eubrontes) spp. and Grallator (Anchisauripus) spp., but Anomoepus, Batrachopus and Otozoum have been reported (Hitchcock, 1858). Unfortunately, even Hitchcock was sometimes not specific about the precise localities from which the tracks came. He often used the term Turners Falls for the entire stretch of exposures from Fall River to the present French King Bridge (Hitchcock, 1858).”

We will descend to the riverbank along a low ridge of basalt. You can identify the top of the basalt flow by the presence of vesicules that are reemerging as holes due to weathering of amygdaloidal fillings of zeolites and quartz. Truncation of a vesicular zone may help you locate a fault, mapped by Wise et al. (1992) in Figure 3.2. Cracks and other openings in the top of the lava flow are filled with sand and mud, as sedimentation continued after the lava cooled. The basalt is 55 m thick and, based on rusty, vesicular zones, is believed to consist of two flows. The lower flow has very clear pillow structures that can be observed in outcrops nearby that we will not visit. The best age for of all basalt flows dated in the Connecticut Valley is 200±1 Ma (Philpotts and McHone, 2003).

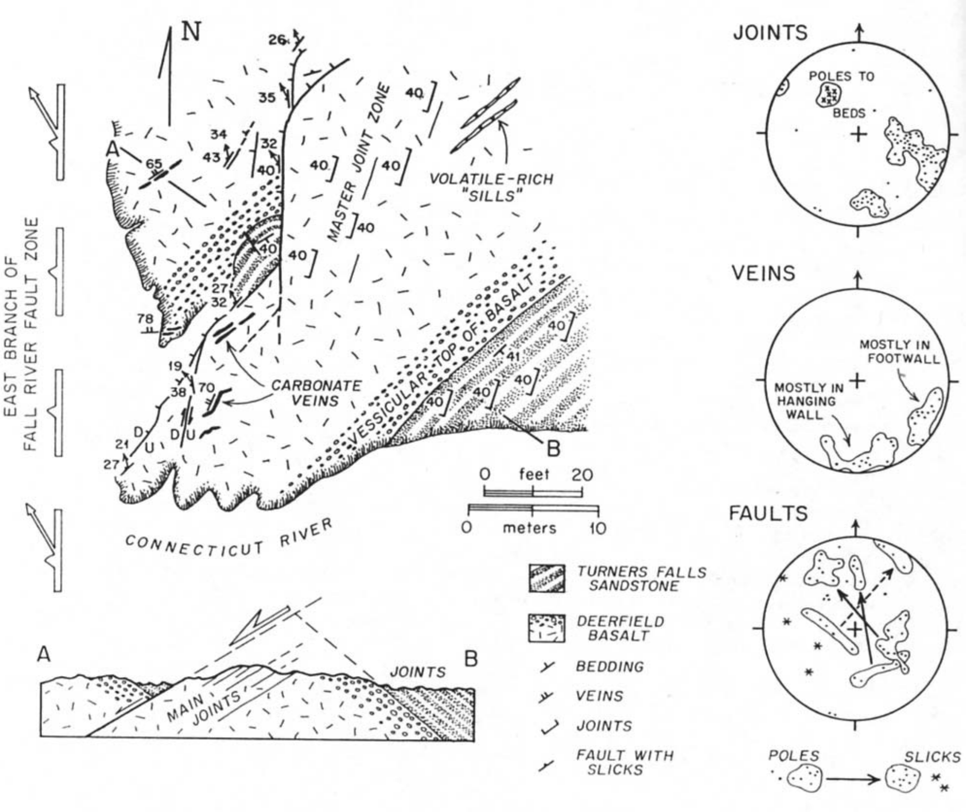


Figure 3.2. A sketch map of the basalt outcrop from Wise et al. (1992).

Above the basalt are outcrops that expose a 250 meter section of the Jurassic Turners Falls Formation. According to Wise et al. (1992): "The sequence of redbed sandstones and mudstones that lie on the Deerfield Basalt are interpreted as the record of a playa that succeeded the lava flows of the Deerfield Basalt. The initial playa muds and sands filled fissures and irregularly-shaped openings in the upper surface of the lava, visible as we cross the contact between the basalt and Turners Falls Formation. These are "neptunian dikes" injected from above."

"Here, the evidence for accumulation of the redbeds in the topographically closed basin of a playa includes the following. 1) The bedding planes are smooth and level without plant root traces. 2) Graded beds that record flood events, thicker examples of which proceed from plane-bedded sandstone with chips of red mudstone -> climbing ripple cross-lamination of the stoss erosional type -> laminated mudstone -> mudcracks. 3) Superabundant burrows, perhaps made by insects. 4) Superabundant mudcracks. 5) Abundant layers of ripple marks. 6) Occasional raindrop impressions. 7) Dinosaur and other reptile tracks were once common, but have been mostly removed. 8) Groove marks made by tools (plant fragments and pebbles?) that were dragged over the muds by flood waters. Over time, the playa surface was built up by floods that varied from major events to small-scale mud-laden ponding of runoff. These flood waters coursed to the southwest across the playa surface from an alluvial fan located to the northeast. The nearly flat surface would be covered by a playa-lake for a few hours to weeks after each flood event, to be followed by drying and desiccation of the muds."

"In the middle of these playa redbeds is a 2-m thick fluvial channel body of pebbly arkose with some clasts up to 4 cm in size. The sandstones are in part cross-bedded, and there is minor channeling into the underlying mudstone. A river flowed a substantial distance across the playa surface, fed by rivers that flowed down from highlands to the N and NE."

From this channel northward along the shore of the Connecticut River to lake sequence 0, faint, spaced cleavages are visible as subtle parallel lines on bedding. Also visible are many depositional grain lineations in horizontally laminated sandstones. A change in strike of about 10 degrees occurs in this area, but it is uncertain whether this is the result of later faulting or of rapid subsidence during sedimentation. The former seems more likely. About 6-8 m below lake 0, a zone of nested scours (?) occurs at the base of a sandstone which has fragments of the underlying mudstone imbricated by flow to the SW."

There are many interesting geologic features that can be observed at this locality. The following is a list of features that you can find, in order, in a traverse to the east. Can you find them all? Check them off as you go.

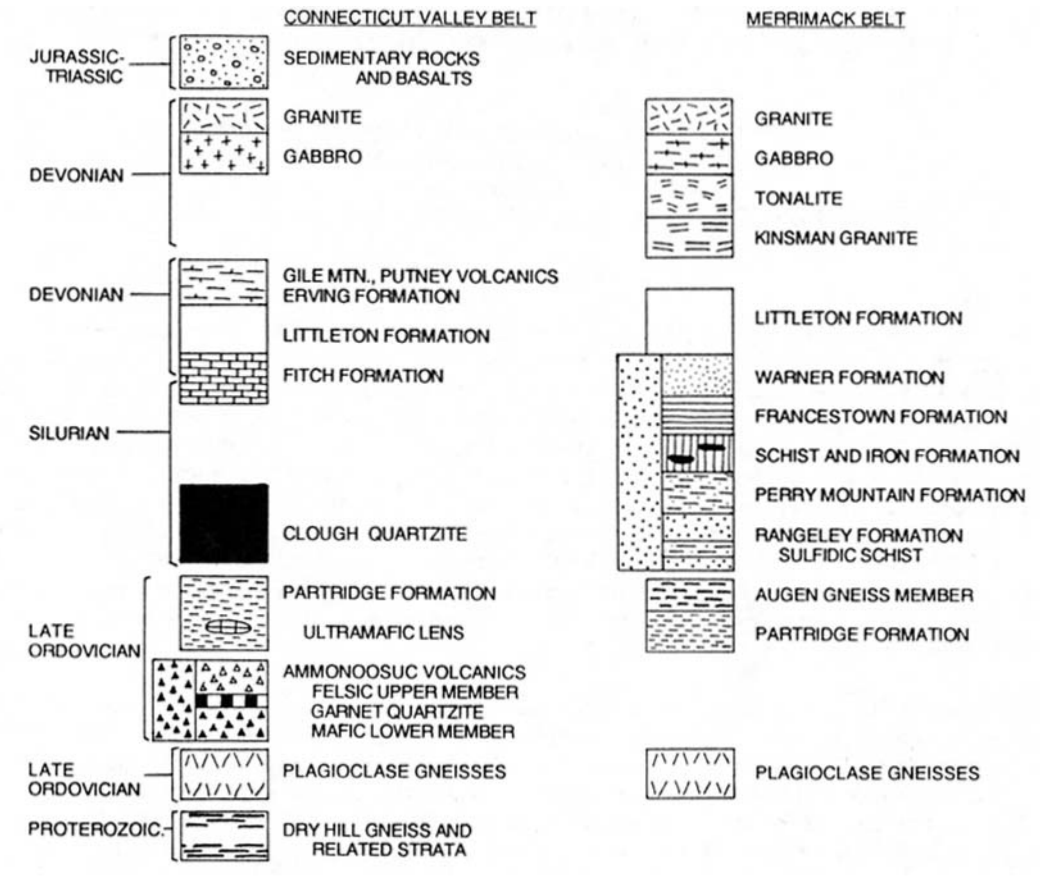
☐ vesicles in the top portion of the basalt
☐ fault offset of the basalt
☐ contact between the basalt (below) and the sedimentary layers (above)
☐ sandstone filling cracks in the basalt
☐ interbedding of red colored mudrock and sandstone layers
☐ small scale cross-bedding in sandstone

☐ start of extensive bioturbation (burrows) in mudrocks (why aren't the layers deposited immediately above the basalt bioturbated?)
☐ mudcracks in mudrocks
☐ ripples in sandstones
☐ the first pebbly arkose unit (check out the shape of the sandstone layer beneath, and look for any evidence for sorting and stratification in the pebbly arkose to help you interpret its origin)
☐ rip-up clasts of mudrocks in this pebbly arkose
☐ soft-sediment deformation features (within 1 to 2 meters above this pebbly arkose)
☐ large-scale cross-bedding in sandstone
☐ the first (and subsequent) gray to black mudrocks/shale
☐ coalified plant fossils in large sandstone blocks/boulders just passed the bridge
☐ neat fractures in the shale with fibrous dolomite crystals
☐ carbonate (dolomicritic) nodules, lenses and layers (many weathered brownish)
☐ more faults!
☐ a large conglomerate block/boulder with interbedded sandstone with beautiful cross-bedding (very close to the dam)
☐ black shales with fish fossils
☐ deep mudcracks in the uppermost dark mudrock unit near the dam (what are these mudcracks filled with?)

Stop 4 – New Erving Roadcut

The Paleozoic history of New England is complex and includes three major orogenic events: the Taconic Orogeny (460-450 Ma), the Acadian Orogeny (420-360 Ma), and the Alleghenian Orogeny (330-290 Ma). This new roadcut is a good place to view some of the consequences of those events as recorded in rocks exhumed by millions of years of erosion from deep within the Himalayan-scale mountains that once existed here.

Here you will see rocks of the Ordovician Partridge Formation (sulfidic and graphitic mica schist, amphibolite, and calc-silicate) and Ordovician or older? Fourmile Gniess (layered to massive, biotite-feldspar gneiss and amphibolite). These rocks have been metamorphosed to the kyanite-staurolite zone, but you are not likely to see either of those minerals because of the bulk composition of the rocks. You will see numerous garnet crystals, whose pale color indicates an Mg-rich composition due to the preference of sulfides for iron. You will also see numerous small folds in the rock that provide hints to the complex structure of the region. The stop is located on the west limb of a syncline between the Kempfield anticline to the east and the Pelham Dome to the west (Figure 4.1)



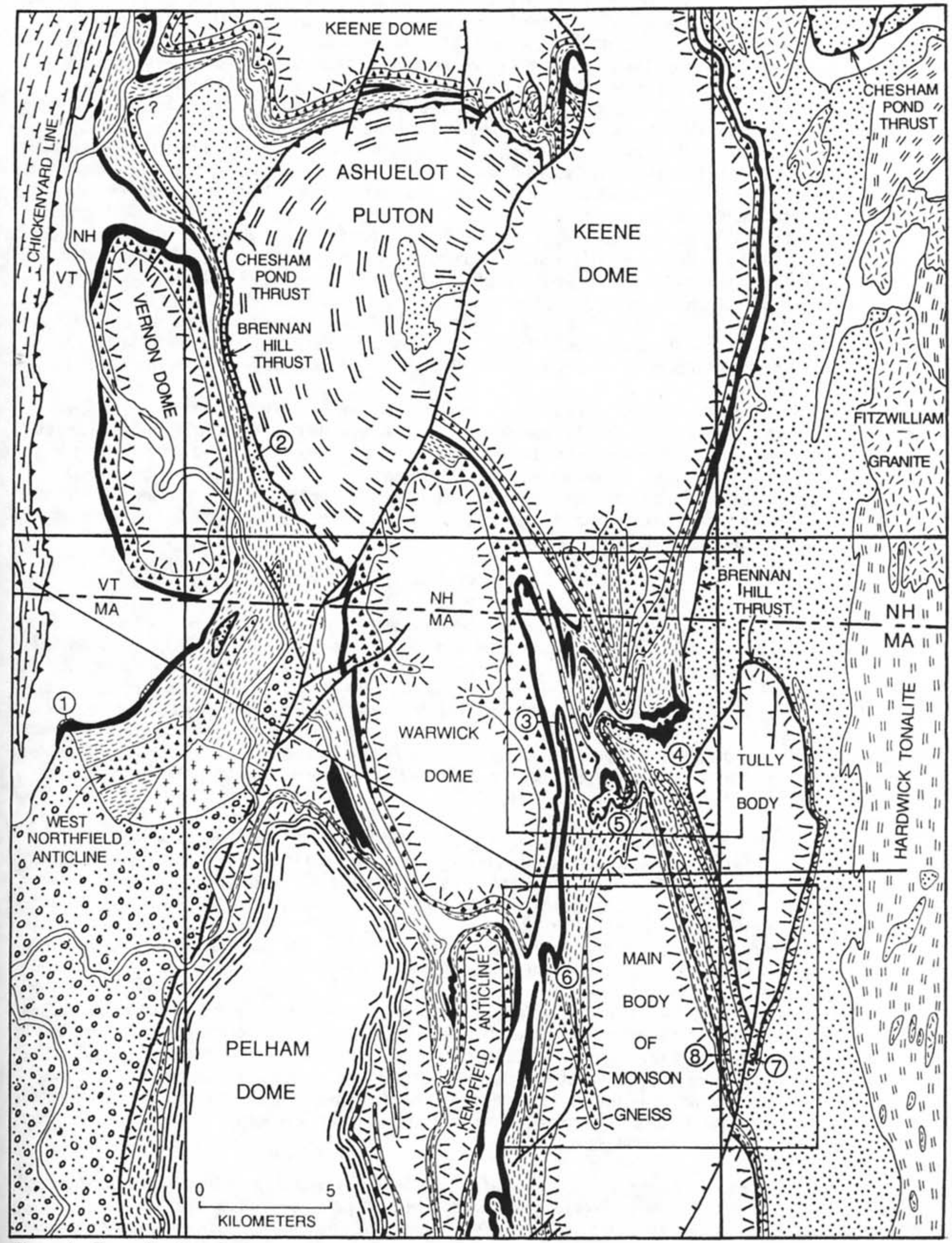


Figure 4.1. Geologic map of north-central Massachusetts (Robinson and Elbert, 1992). Stop 4 is located in the bottom center between the Pelham Dome and the Kempfield Anticline.

Some of the geological secrets concerning the history of this area are being revealed by zircon and monazite dating. The following abstract by Spear et al. (2006) is for a paper that was presented at the Northeastern Section meeting of the Geological Society of America a few weeks ago. The attached Figure 4.2 will help Jack Cheney explain their arguments.

THE SUTURING OF NEW ENGLAND

Spear, F. S., Cheney, J. T., Pyle, J. M.

A long standing conundrum of New England geology is the enigmatic relationship between the contemporaneous Barrovian metamorphism of Vermont, with its clockwise P-T paths culminating in maximum burial depths of ca 30-35 km (indicative of continental collision) and the Buchan metamorphism of New Hampshire, with counterclockwise P-T paths and regional high-T, low-P metamorphism (indicative of an extensional terrane). Thermal and tectonic considerations do not readily permit these distinctively different terranes to coexist in their present proximity.

We have examined a belt of Devonian Littleton Formation that crops out from northern New Hampshire to central Connecticut along the Bronson Hill anticlinorium in central New England, which we believe holds the answer to this paradox. The metamorphism in this unit is distinct from both overlying nappes and underlying schists and is characterized by nearly ubiquitous, large (several cm) staurolite crystals that postdate the dominant fabric (the “Big Staurolite nappe” or BSN).

Monazite from this unit is typically only weakly zoned with high Th cores, suggesting a single episode of growth. No multi-generation monazite has been observed. New electron probe dates on metamorphic monazite from this unit reveal no ages younger than ca 340 Ma. Monazite ages from regions along the strike of the unit are: 280±10 from the Salmon Hole Brook syncline (northern NH); 300±10 from near Mascoma, NH (west-central NH); 320 ±10 from around Fall Mountain (west-central NH); 270±10 from the near Bolton, CT. Similar ages have been obtained by SIMS, and a single TIMS analysis of monazite from the Northfield syncline in north-central MA of ca 295 Ma (R. Tucker).

In our model, the BSN represents the trace of a mid-crustal shear zone that was responsible for the westward transport of the rocks of the Central Maine terrane onto the Bronson Hill anticlinorium, in juxtaposition with the Barrovian terrane of Vermont during the early stages of the Alleghanian orogeny. Transport took place over as much as tens of millions of years through the Carboniferous culminating in the Early to Mid Permian. Metamorphic pressures of 5-7 kbar suggest the shear zone was active at depths of ca 20 km depth. Rocks of the BSN are folded by the domes of the Bronson Hill anticlinorium, and it is suggested that dome formation is a consequence of thrust ramping of Oliverian plus Avalon basement onto the Laurentian margin.

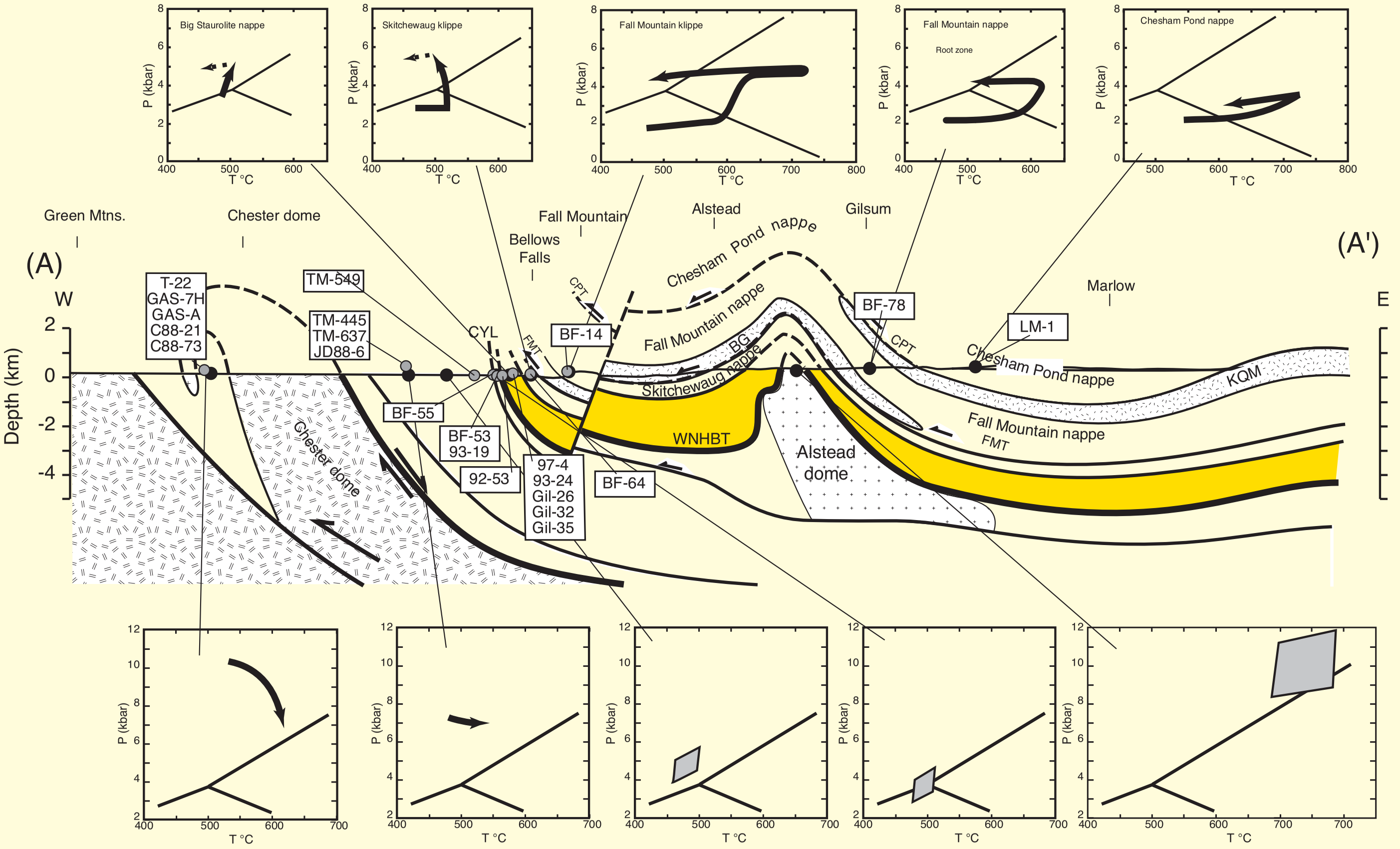


Figure 4.2. Summary of metamorphic PT paths and a cross-section across Central New England. Modified from Spear et al. (2002).

Stop 5 – Eastern Border Fault of the Mesozoic Basin

Walk west along the wide path to the right (north) of the Ashram parking lot. Exposed in the woods and especially along the powerline cut, are low outcrops of metamorphic rocks filled with pegmatite intrusions. These rocks are mapped as part of the same Ordovician Partridge Formation that we saw at the last stop in Erving. Continue to the west, crossing a railroad track, to where Roaring Brook has formed a gorge and waterfall at the base of Mt. Toby. There you will find outcrops of Mt. Toby Conglomerate, a Jurassic unit that is stratigraphically above and interfingers with the Turners Falls Sandstone we saw earlier. The Mt. Toby Conglomerate forms cliffs that expose unusual sedimentary textures. The clasts are very angular and can be quite large, some approaching a meter across.

The juxtaposition of Ordovician and Jurassic rocks here is evidence for the presence of a large normal fault, the “Eastern Border Fault” of the Mesozoic rift basin. Although there are faults along both the east and west sides of the Hartford and perhaps the Deerfield Basins, the throw on the Eastern Border Fault is much greater, and has been estimated to be as much as 5 miles near Hartford. The dip of the fault there is also believed to flatten at depth towards the west, a feature common in the listric style of normal faults. In the Mt. Toby area, and south to the Holyoke Range, fault steps are found which bring the basement up close to the ground level (Chandler, 1979). The coarse clast size here is the result of rock slide and talus deposition adjacent to a fault scarp, similar to deposits being built today in Death Valley.

Look for bedding in the conglomerate. See if you can identify the rocks that are the clasts. Are they igneous and metamorphic rocks only? Sketch a cross section that runs from the Partridge Formation across the fault and into the Mt. Toby conglomerate. Does this story make sense to you? Enjoy the waterfall. Be careful not to fall yourself.

Figure 5.1. Mt. Toby Conglomerate. The field of view is four feet wide.

Stop 6 – Sunderland Delta

Glacial meltwater streams flowing into Lake Hitchcock deposited the coarser part of their sediment load as they entered the lake to form a delta. In cross-section, nearly horizontal coarse gravelly sand topset beds overlie dipping foreset beds of medium to fine sand which in turn overlie bottomset beds of varved clay (Figure 6.1). The water level within the lake can be determined by measuring the elevation of the topset/foreset contact.

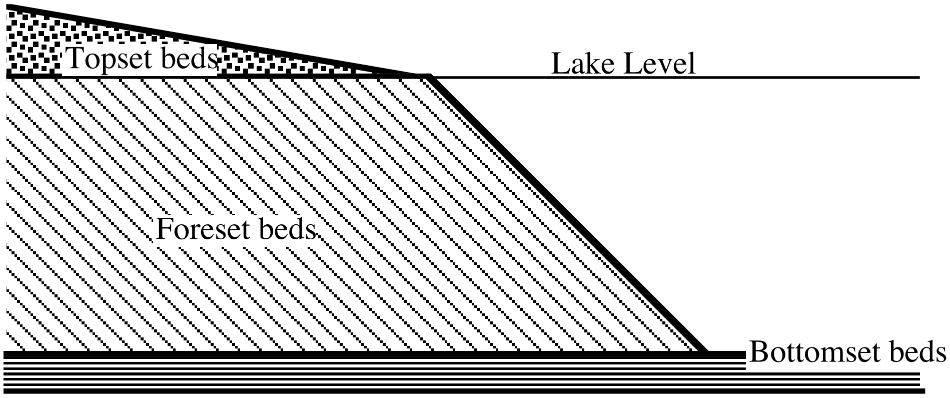


Figure 6.1. Diagram showing a cross-section through a delta.

Once geologists realized that there was once a glacial lake in the Connecticut Valley, they looked for deltas to determine the depth of the water. The top of the Sunderland Delta is 96 meters above sea level. The bottom of the Connecticut Valley is 35-45 meters, making the lake as much as 60 meters deep. To the surprise of some, delta elevations are not constant up and down the valley. For example,150 km to the north, near Hanover, NH, deltas have elevations over 200 meters. Because lake surfaces must be horizontal, this observation requires differential uplift, with the Hanover, NH area rising over 100 meters more than the Sunderland, MA area. The explanation for this uplift is glacial rebound: rising of the crust due to the removal of the mass of a two-mile thick ice sheet. It is believed that the greater rebound in the north is the results of the ice having been thicker there on average. For more details on the postglacial uplift, see Koteff and Larsen (1989).

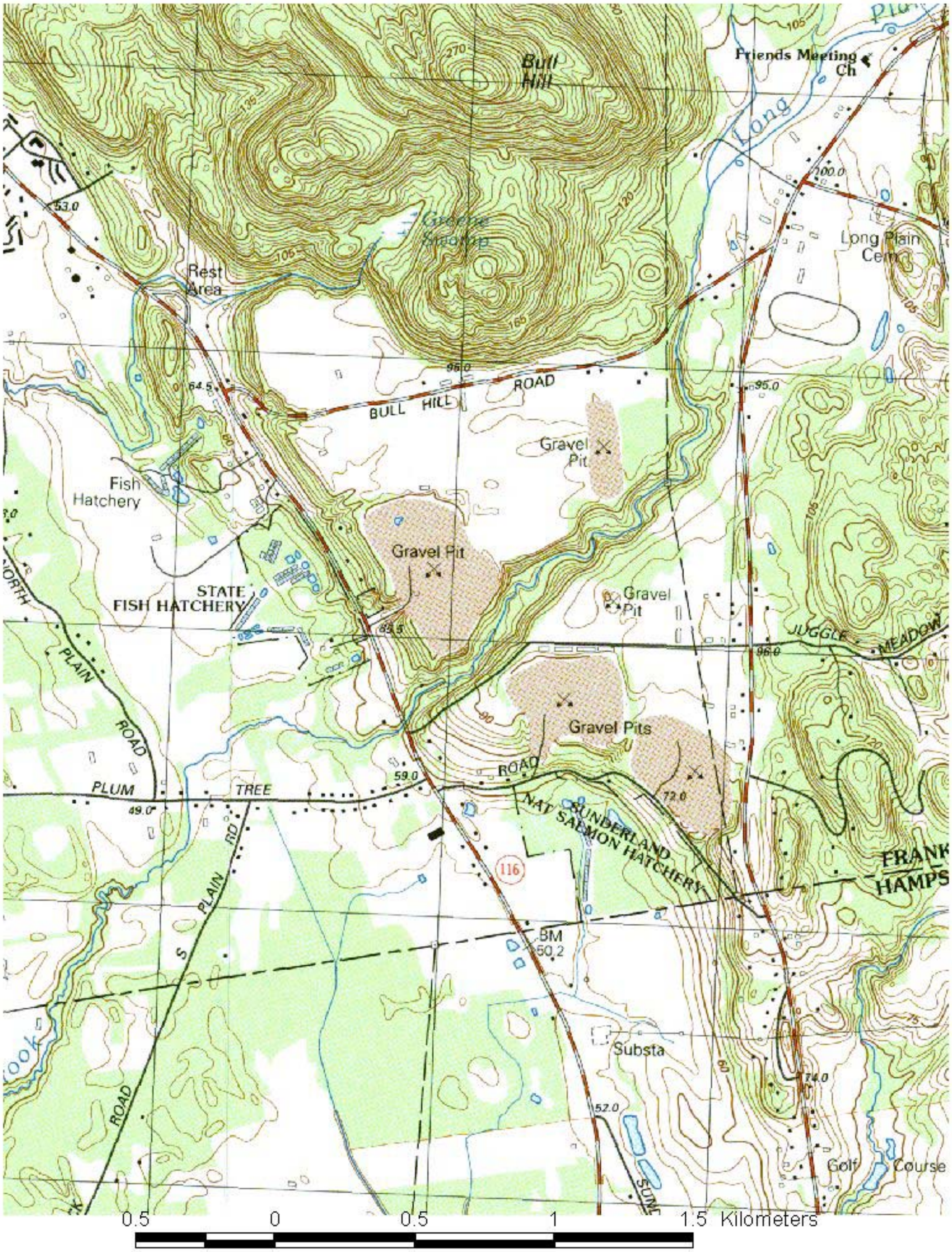


Figure 6.2. Topographic Map of the Sunderland Delta.

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