GLOBAL GEOSCIENCE TRANSECT 8: QUEBEC-MAINE-GULF OF MAINE TRANSECT, SOUTHEASTERN CANADA, NORTHEASTERN UNITED STATES OF AMERICA

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INTRODUCTION

Transects produced by the North American Continent-Ocean Transect Program (Speed and others, 1982; U.S. Geodynamics Committee, 1989) describe the complex orogen formed along the Late Proterozoic to Early Cambrian rifted margin of the Laurentian craton, including continental rocks of the Grenville province, which themselves had been earlier formed by a Middle Proterozoic continent-continent collision. Continental accretion was followed by continental separation and the formation of a passive continental margin. This 900-km-long transect crosses the entire Appalachian orogen and the passive continental margin from the craton to Atlantic oceanic crust. It is supported by abundant seismic reflection and refraction data that were gathered specifically for this transect, in part, to resolve questions posed by the earlier transects across the Atlantic passive continental margin. Thus, it is a "second-generation" transect. This transect has been accepted in the Global Geoscience Transects Project (CC–7) of the Inter-Union Commission on the Lithosphere, International Council of Scientific Unions (Monger, 1986) as Transect 8.

The Quebec-Maine-Gulf of Maine transect provides excellent data that can be used to understand the processes of continental accretion and separation. An uncertain number, possibly ten or more, of tectonostratigraphic terranes of predominantly continental affinity were accreted to the craton, as shown by the transect. In addition, at least two oceanic terranes also were accreted. Although some terranes had been joined together to form composite terranes before being accreted (Boone and Boudette, 1989, describe an example), in general, these terranes were accreted episodically to the southeastern (current geographic direction) part of the craton by thrust and (or) strike-slip faulting at successively younger times during the Paleozoic. The terranes differ principally in the nature of their Late Proterozoic to middle Paleozoic history and stratigraphy (Keppie, 1989) and paleontology (Neuman and others, 1989). Accretion in the part of the orogen shown on this transect took place during multiple collisional or transpressive episodes in the early and middle Paleozoic.

A very thick continental crust was produced during these episodes. Geologic evidence that indicates the formation of this thick crust includes Early Devonian garnet rhyolites extruded after fractional melting of felsic rocks at depths of greater than 50 km (Stewart, 1989), a hinge zone (Hatch and others, 1983) between a thin and a very thick (~10 km) sedimentary prism deposited in a successor basin (the Central Maine synclinorium) that formed after the late Early and Middle Ordovician (Taconian) collisional episode, and kyanite-bearing assemblages indicating the burial of even the youngest of these sediments to mid-crustal depths. Direct geophysical evidence today indicates that thick continental crust remaining from Paleozoic accretion is lacking. However, strong seismic reflectors at the base of allochthonous sheets can be traced to depths of approximately 25 km, and 10 km or more of additional allochthonous sheets may initially have been present but have since been eroded. It is reasonable to infer that the underlying continental crust must have been depressed by such thick tectonic loading. However, most stratigraphic evidence for a foreland basin in the Ordovician has been buried under later thrust sheets, and evidence for a foreland basin related to the Acadian orogeny is lacking in the region of this transect.

As shown by abundant seismic reflection and refraction data, the present Appalachian crust was shaped principally by late Paleozoic to Early Jurassic extension. This extension began with the formation of Mississippian and Pennsylvanian transtensional rift basins and continued with the development of a 400-km-wide province of rift basins of Triassic to Early Jurassic age that was accompanied across a wider region by the intrusion of numerous dike swarms and widespread normal faulting. Extension reversed the throw on many inclined Paleozoic thrust faults. The new seismic data suggest that during extension the lower crust and Moho concurrently acquired their present shape and internal features by ductile shear and recrystallization. Many of the faults bounding rift basins become listric into the top of the lower crust (Stewart and Luetgert, 1990) and, in some places, are paired with a rise of Moho by as much as 4 km over a distance of 15 km. Numerous but discontinuous subhorizontal reflectors in the lower crust indicate that during extension laminar packets of contrasting lithologies were formed and simultaneously through-going older faults were obliterated. The strong seismic reflections from the allochthons in the upper crust die out at depths of 20 to 25 km into the uppermost newly reformed lower crust. The top of the lower crust is a gradational zone as much as 4 km thick where seismic velocity increases from about 6.4 to 6.8 km/s; such a large increase in velocity must correspond to a decrease in quartz and potassium feldspar contents. The extended region has newly reformed lower crust and a well-defined reflection and refraction Moho, which contrasts with

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the indistinct Moho beneath Grenvillian crust to the northwest. Extension led to rupture of the continental crust and formation of the present Atlantic Ocean, which was followed by deposition of sedimentary rocks upon the passive continental margin.

The Quebec-Maine-Gulf of Maine transect demonstrates that the record of Paleozoic orogenies in this part of the orogen is preserved almost exclusively in the presently brittle upper crust. Paleozoic terrane boundaries only can be identified to the top of the lower crust, which appears to pass smoothly beneath all terranes. Any genetic relation of the lower crust to the overlying crust becomes increasingly uncertain as the amount of extension increases progressively southeastward toward the locus of oceanic rupture. The lower crust was attenuated and extended southeastward from its original position as it was ductilely deformed. The high proportion of the upper crust that now consists of exposed granitic rocks was formed by partial melting of lower crust in the Paleozoic. Given any reasonable figure for melt extracted, indications are that most of the lower crust was partially melted then. The bulk composition of the lower crust must have been changed by gravitational ascendance of magmas, resulting in depletion of quartz and alkali feldspar. For these reasons we have shown the lower crust as a discrete geologic entity on our interpretive cross section.

This transect is a cartographic product that is an internally consistent generalized representation of the data in an underlying digital geographic information system (GIS) where the map units are shown with the detail present in the original 1:250,000- and 1:500,000-scale maps (Globensky, 1987; Klitgord and others, 1988, plate 2C; Osberg and others, 1985; St.-Julien and Slivitsky, 1985; Weed and others, 1974). Map units were assigned attributes according to a scheme devised by Wright and Stewart (1990). Geologic data preserved in the GIS that cannot be shown on maps at a scale as small as 1:1,000,000 include the specific names, symbolic abbreviations, and lithologies of stratified formations and their members; greater detail about intrusive rock names, compositions, and mineralogies; and more complete details concerning age assignments and their ranges, if any. Another assigned attribute in the GIS is the tectonic setting of original deposition or igneous intrusion. We have used these data to display a tectonic setting map and accompanying interpretive cross section of the whole crust that conforms to the tectonic setting map. Digital optical discs (CD-ROM) will be used to archive the GIS with the different geological and geophysical data sets used to prepare this transect. The data set archived for the seismic reflection profiles will include reflection intensities, and the data set for the refraction data will include lines and fans not shown as profiles because they were parallel to regional strike. Many other combinations of geologic and geophysical data sets can be performed using digital data. Some examples given by Phillips (1990) include a color shaded-relief aeromagnetic map with superimposed geologic contacts and faults and color composites of different wavelengths of bandpass-filtered aeromagnetic or gravity maps (or combinations of both kinds of potential-field data) with superimposed geologic contacts and faults. A three-dimensional model of parts of this transect in western Maine has been prepared also (Unger and others, 1989).

ACKNOWLEDGEMENTS

Starting in 1983, this transect required the efforts of a large team of geologists, geophysicists, and cartographers from the U.S. Geological Survey (USGS), the Geological Survey of Canada, the Maine Geological Survey, the New Hampshire Geological Survey, and numerous universities. Only the principal authors of the transect graphic and pamphlet are listed as compilers. This group was substantially augmented for specific tasks by the scientists listed as contributors. Many of the contributors have or will publish their work in greater detail elsewhere; most published work is cited in this report. Most field data were collected in 1983–1985, although some geophysical data were collected as recently as 1988. In addition, we obtained important data, advice, or other valuable assistance from W.A. Anderson, M.E. Loiselle, and R.G. Marvinney of the Maine Geological Survey; Phyliss Charlesworth, John Glynn, Alan Green, Patrick Morel-a-l'Huissier, W.H. Poole, and M.D. Thomas of the Geological Survey of Canada; G.M. Boone of Syracuse University; R.A. Phinney and Christel Bottcher of Princeton University; W.E. Doll of Colby College; S.L. Klemperer, then at Cornell University; C.V. Guidotti and P.H. Osberg of the University of Maine-Orono; A.M. Hussey II of Bowdoin College; C.E. Mann of Stanford University; and R.A. Ayuso, Richard Goldsmith, and A.M. Trehu of the U.S. Geological Survey. Joshua Comenetz and Sara Minnich of the U.S. Geological Survey helped to prepare the transect graphic.

SOURCES OF DATA USED

Extensive new compilations of regional magnetic and gravity data on land and at sea were recently prepared for much of the transect region (Macnab and others, 1990; D.R. Hutchinson, unpub. data). Included were gravity and magnetic data collected by transect participants in support of this transect. Recently published geologic maps for Maine (Osberg and others, 1985) and southeastern Quebec (St.-Julien and Slivitsky, 1985; Globensky, 1987) were used (see small-scale index map for coverages). The regional compilations were prepared at scales larger than the transect, mostly at 1:500,000 or 1:250,000, so only a minor part of each regional compilation was used. Digitizing all these regional data sets was one of the most time-consuming aspects of transect preparation. All data were converted to the Lambert Conformal Conic map projection having standard parallels at 33° N. and 45° N. and a central meridian at 69° W. Gravity and magnetic data were incorporated in gridded form.

Most of the geologic data were digitized by machine-scanning of stable-base separates of the maps. After editing and conversion to line-graph data, attribute codes were then manually attached to these data. Details of procedures and problems encountered and the attribute codes are given in Wright and Stewart (1990). The minimum size of a geologic object shown on the transect graphic is approximately 1 mm by 1 mm, or 1 km^2 at 1:1,000,000. For clarity at this scale, some similar-aged formations and all members of formations had to be combined on the geologic map within a polygon with one color. In a few places, volcanic units within formations were generalized into map units large enough to display overprinted patterns. Although some of the detail shown on the source maps was lost to make the generalized transect maps legible, all of the attributes of the individual geologic map units were preserved in the GIS. Geologic point data, such as strikes and dips, could not be shown at map scale and were not digitized. Other geologic point data, such as drill holes at sea, ocean-bottom seismometer locations, and sites sampled by the submersible *Alvin* (Hermes and others, 1978), were digitized manually.

In many places, the geologic data sets from the original sources were modified by the transect compilers so as to make geologic units continuous across political boundaries, to accommodate new geophysical, geochronologic, and geologic mapping data, and to reflect new understanding of regional geologic processes as deduced from the transect. Surficial and glacial deposits are not shown although all the land area was glaciated in the Quaternary. Coastal and south-central Maine were isostatically depressed by ice loading in excess of 100 m relative to present sea level, so surficial deposits of marine character are found as far inland as Skowhegan. Widespread surficial and glacial deposits include sand, gravel, marine clay, and till (Thompson and Borns, 1985); together these reach 100 m in thickness locally.

Seismic reflection and refraction profiles and lines, fans, and related shotpoints shown on the gravity map are the loci of vertical planes of seismic data that traverse the entire crust. Only those vertical planes that lie close to the line of the interpretive cross section are shown here. Data for approximately 1000 km of deep (15 s) seismic reflection lines and approximately 2000 km of deep seismic refraction data were available; nearly all were collected specifically to support this transect (Stewart and others, 1986; Klemperer and Luetgert, 1987; Unger and others, 1987; Luetgert and others, 1987; Hutchinson and others, 1988; Spencer and others, 1989; Luetgert and Mann, 1990; Hennet and others, 1991; and Luetgert and Doll, in press).

DISPLAY OF GEOPHYSICAL DATA AND ITS PROCESSING

Several kinds of geophysical data are displayed as maps or vertical profiles through the crust at a common scale and with no vertical exaggeration. The coverage for potential-field data is complete for the transect and overall is of high quality. Aeromagnetic and gravity profiles for the line of cross section are displayed to facilitate direct comparison with each other and with geologic structures on the interpretive cross section. Migrated normal-incidence seismic reflection data are displayed as projected as much as 10 km onto the trace of the composite seismic reflection profile, except for line 3B, which is projected as much as 30 km along the regional strike. All of the available data on land appear on the composite profile. Only the marine data for USGS lines 1 and 1A (Hutchinson and others, 1988), USGS line 8 (Gyre 85–11–8) (D.R. Hutchinson, unpub. data), and USGS line 19 (Klitgord and others, 1982) were used. Many other deep seismic reflection profiles for Georges Bank show that the data obtained for line 19 is typical for this region (Schlee and Klitgord, 1988). Seismic refraction data are displayed from a 300-km-long profile on land nearly normal to regional strike (Spencer and others, 1989; J.H. Luetgert, unpub. data). This profile has been extended approximately 100 km seaward because, according to J.H. Luetgert (written commun., 1990), traveltimes computed for the coastal part of the profile match well with traveltimes collected at Matinicus Rock during the 1961 Maine cooperative onshore-offshore experiment in crustal seismology. This experiment was the first fixed array-moving shotpoint study (Meyer and others, 1962; Steinhart and others, 1962a, b; Nakamura and Howell, 1964; Suzuki, 1965). As only preliminary velocity models are available for the marine ocean-bottom seismometer profiles (Sattel and Trehu, 1989), and the results contain large uncertainty in the lower crustal thicknesses and velocities, we have displayed only their locations on the map of seismic profiles. Velocity-depth profiles of along-strike refraction profiles (lines 4–8) are plotted on the seismic reflection profile at the point where each along-strike refraction profile crossed the section. Prominent wide-angle reflections from seismic refraction fans (shots fired broadside into refraction deployments) also are shown on the seismic reflection profile where the line connecting midpoints of the refraction fan profile intersected the reflection profile.

The limited available heat flow data are mostly in Maine (Decker, 1987) but have not been plotted. They indicate that a normal reduced flux characterizes the entire region, which implies that the lithosphere beneath Maine has equilibrated since it was perturbed by Paleozoic and early Mesozoic orogenic epochs. Decker concluded from his data that there are no significant differences in heat flow between terranes having different basements and that extension beneath Maine was brief or did not warm large vertical thicknesses of the lithosphere.

Gravity values were corrected for terrain effects onshore (density 2.67 g/cc), earth tide and instrument drift, and are referenced to the International Gravity Standardization Net of 1971 (Morelli, 1974). Anomaly values were calculated by using the 1967 formula of the Geodetic Reference System (International Association of Geodesy, 1971). Complete Bouguer anomalies were used onshore, and free-air anomalies were used offshore. The two types of anomalies are equivalent at the coastline. The data set was gridded at 0.8-km intervals and contoured at 10-milligal intervals. Estimated error is ± 2 milligals. Nearly 2200 new gravity measurements to support this transect were made in 1983 and 1984 along the lines surveyed and vibrated during seismic reflection data collection programs (Snyder and Phillips, 1986; Phillips and others, 1988). Density and magnetic susceptibility measurements were performed on about 150 rock samples collected near measured gravity stations. These closely spaced data insure that the gravity profile shown for the cross section is of very high quality. An intensive collaborative revision of gravity data offshore was performed by colleagues from the Bedford Research Institute of the Geological Survey of Canada, the USGS at Woods Hole, Mass., and the Woods Hole Oceanographic Institution (D.R. Hutchinson, unpub. data). Marine gravity line spacing was approximately 14 km in the Gulf of Maine and approximately 20 km for Georges Bank. The offshore data were initially gridded at 7.5 km.

Aeromagnetic data for individual surveys were projected and gridded at a 0.8-km interval. If necessary, the data were continued to a surface 1000 ft above the terrain, and regional trends were removed. The grids for individual surveys were smoothly merged into a master grid, which was machine contoured at 100-nanotesla intervals. Estimated error is ± 25 nanoteslas. A similar compilation of magnetic data has been prepared for the Gulf of Maine and adjacent land areas by Macnab and others (1990).

The seismic refraction data on land were obtained in 1984 by the USGS and the Geological Survey of Canada using arrays of 200 2-Hz seismometers at approximately 800-m spacing having an estimated maximum survey location error of 25 m. Shotpoints were spaced 20 to 45 km apart using 900 to 1800 kg of chemical explosive shot mostly in drilled 90-m holes (Murphy and Luetgert, 1986, 1987). Shallow (<10 km) seismic wave velocities were computed from refraction first-arrival times and amplitudes by using two-dimensional raytracing techniques and synthetic seismograms (Luetgert, 1988a, b, c; Klemperer and Luetgert, 1987). Wide-angle reflection data from mid- to deep-crustal reflectors present in the records were interpreted from normal move-out corrected data (Klemperer and Luetgert, 1987; Luetgert and others, 1987). Detailed explanations of data processing and interpretations of results for the across-strike lines are given in Spencer and others (1989) and Luetgert and others (1987). Similar details and results for the along-strike lines are given by Klemperer and Luetgert (1987), Luetgert and Mann (1990), and Hennet and others (1991). Observed arrival times fit computed arrival times to within 0.1 s, so that the uncertainty on any velocity determination is approximately ±0.1 km/s. The depth uncertainty of features in the upper crust based upon agreement within 0.1 s between predicted and observed traveltimes is probably approximately ±0.8 km, increasing to ±2 km at Moho depths. Seismic anisotropy is shown by the differences in velocities at the same site, commonly with slower across-strike and higher along-strike velocities. At some sites, however, the differences are small and could have arisen from different approaches or methods used for data processing. At other sites, the differences exceed the uncertainties of the two compared velocities. It might seem that the site near kilometer 325 on the cross section where the two longest refraction profiles cross would be the best point for comparison. In fact, that point was a common shotpoint for two different seismometer deployments to the northwest and southeast, respectively, so there are not good data to define the across-strike velocity at this point. Seismic anisotropy having faster velocities parallel to regional strike also has been observed by Kafka and Reiter (1987) for the group velocity dispersion of short-period Rayleigh waves generated by the same shotpoints. Kafka and Ebel (1988) also obtained similar results using short-period surface-waves and body-waves generated by the same explosions.

Seismic reflection data were gathered by using vibrators on land and air guns at sea. Recording was for 15 s for onshore lines 1, 2, 3A, 3B, and 4 and offshore lines 1 and 1A; fold was normally 133 onshore and 30 offshore. Onshore line MERQ was recorded to 4 s for the northwesternmost 58 km and to 6 s for the southeastern 107 km. Marine line 8 (24 fold) was recorded to 6 s, and line 19 (36 fold) was recorded to 12 s. The variety of problems solved during processing of seismic reflection data are described by Hutchinson and others (1988), Spencer and others (1989), and Hutchinson and Lee (1989). Velocities for stacking seismic reflection data were obtained by trial velocity stacks at numerous intervals along the profiles and from the seismic refraction studies.

A plot of the compressive wave velocity versus depth for all of the velocity profiles derived from the refraction experiments in Maine by Luetgert (*in* Unger and others, 1987) showed that the crustal velocities for the region could be reasonably modeled by a linearly increasing function. Digitized line drawings of the principal reflectors on all profiles except line 19 were migrated using a raytracing migration program (Unger, 1988a) using the function $V = V_o + kz$ (where V_o is the initial surface velocity, taken as 5.98 km s^{-1}; k is the velocity gradient, taken as 0.027 s^{-1}; and z is the depth in km). Line 19 at the southeastern end of the transect, where thick Mesozoic and Tertiary strata overlie older rocks, could not be migrated using a single linear velocity function, so the digitized reflection section was divided into three sections. Within each section a different linear velocity model, which fit the observed velocity data, was used for migrating the reflectors. The Moho reflectors were migrated with the function $V = 5.75$ km $s^{-1} + 0.001\ s^{-1}z$, and the reflectors within 0 to 5 s two-way traveltime out to the edge of the continental shelf were migrated with the function $V = 2.8$ km $s^{-1}z + 0.3\ s^{-1}z$. The reflectors outboard of the edge of the continental shelf also required use of the function $V = 1.0$ km $s^{-1} + 0.5\ s^{-1}z$ in order to fit the observed velocity profiles and to account for the increased water depth.

MAJOR TECTONOSTRATIGRAPHIC TERRANES CROSSED BY THE TRANSECT

This transect from craton to Atlantic oceanic crust crosses the six major tectonostratigraphic terranes of this part of the Appalachian orogen. Figure 1 displays and labels these major terranes and thirteen of their principal components. Different shades of colors delineate belts on this figure and show the composite nature of the terranes. This means that rocks that have significantly different tectonic settings of origin or early history can be identified within such a terrane. A careful effort has been made to show the bounding faults between the major tectonostratigraphic terranes on the transect maps and sections. Red ticks along both long edges of all the transect maps mark these faults, and the same red ticks are used along the tops of the sections where the bounding faults reach the surface. A tectonostratigraphic evolution diagram gives our interpretation of the time-space history of the different terranes. A description of each of the major tectonostratigraphic terranes from craton to ocean is given below.

Laurentian Craton

The basement of the Laurentian craton within the transect area was formed by the late Middle Proterozoic Grenville continent-continent collision. Crystalline rocks exposed at the surface include high-grade gneisses, migmatites, and deformed intrusive rocks having a wide range of compositions and ages (Moore and others, 1986; Rankin and others, 1989). The craton was extended in Late Proterozoic time. Late Proterozoic rifting formed graben that were then filled with Late Proterozoic sediments and bimodal volcanic rocks (Rankin and others, 1989). Following deposition of this sequence, platformal and slope-rise sediments were transgressively deposited upon the craton from Early Cambrian to Early Ordovician time. Taken together the tectonic setting and geologic maps and cross sections for this transect show the craton with unconformably overlying platformal rocks of Cambrian and Ordovician age. Compression toward the craton during the Middle and Late Ordovician Taconian orogeny formed an imbricate stack of thrust sheets that carried the platformal sediments and successively outboard strata onto the craton along southeast-dipping faults such as the Aston and Logan's Line. The interpretation of St.-Julien and Hubert (1975), St.-Julien and others (1983), and Ando and others (1983) is that the décollement rises into younger strata toward the craton. These authors interpret the thrust faults to have been propagated toward the craton although some of the external nappes may have been emplaced by gravity sliding. Except near the southeastern end of the MERQ profile where new seismic data are more complete (Spencer and others, 1989), we have used these syntheses practically unchanged for the interpretation of the crust at depths less than 10 km shown on the geologic cross section. Crustal processes involved in the Taconian orogeny are well summarized in Stanley and Ratcliffe (1985); their ideas and earlier proposals are discussed in Drake and others (1989, p. 151–155).

Grenvillian basement continues at depth far to the southeast of its surface outcrop. Seismic reflection profiles (Spencer and others, 1989) show a strong reflector at the top of Grenvillian rocks for a distance of about 175 km southeast of autochthonous outcrops. This reflector is interpreted to be mylonite along a décollement possibly initiated along a Late Proterozoic rift fault that was reactivated in the Taconian orogeny and again in the Acadian orogeny. Southwest of the transect area, Trzcienski and Marchildon (1989) describe Grenvillian xenoliths from dikes that straddle the Guadeloupe fault, more than 100 km southeast of exposed Grenvillian rocks. Our interpretive cross section indicates that, at a structurally equivalent site to that for Trzcienski and Marchildon's samples in this transect, Grenvillian rocks are involved in imbricate crystalline thrusts; autochthonous Grenvillian crust occurs at a depth of about 15 km. The three-dimensional shape of the strong reflector at the top of Grenvillian crust beneath the Chain Lakes block has been determined from seismic reflection data (Unger, 1988b); it dips southeasterly at approximately 13°, and it can be traced to depths of approximately 25 km southeast of the Chain Lakes block. Beyond this point, the identity of the lower crust as Grenvillian becomes more speculative and its southeastern limit is conjectural due to deep crustal modification during and after the Paleozoic orogenies. This topic is discussed in the section on the Central Maine Composite Terrane.

Brompton-Cameron (Boundary Mountains) Composite Terrane

The Brompton-Cameron (Boundary Mountains) composite tectonostratigraphic terrane (Zen, 1989; Boone and Boudette, 1989) occurs southeast of the Brompton part of the Baie Verte-Brompton line of Williams and St.-Julien (1982). These authors interpreted the Brompton line as the suture between the Laurentian craton and accreted terranes unrelated to Laurentia. Included in this composite terrane are rocks that were deposited in oceanic and island-arc tectonic settings and a highly metamorphosed diamictite or sedimentary mélange of granodioritic or feldspathic graywacke bulk composition. The latter, known as the Chain Lakes massif or block, was formerly thought to be about 1500 Ma (Naylor and others, 1973), significantly older than Grenvillian crust. However, on the basis of U-Pb dating of single zircons from Chain Lakes rocks, G.R. Dunning and P.A. Cousineau (1990 and unpub. data), Olszewski and others (1990), and Cousineau (1991) have demonstrated that the Chain Lakes rocks are actually younger than Grenvillian rocks and are possibly as young as Early Cambrian. The Chain Lakes block may have originated as a continental margin deposit or as a sedimentary mélange from a continental terrane. The continental terrane may have been the Laurentian craton to the northwest (G.R. Dunning and P.A. Cousineau, unpub. data), a rifted part of the Laurentian craton to the southeast, or some other Proterozoic craton. We speculate that the rocks of the Chain Lakes block were deposited upon Grenvillian crust (see the interpretive cross section). The increase in the seismic refraction velocity of about 0.25 km/s at a depth of about 4 km may be caused by the change from Chain Lakes rocks to Grenvillian rocks. However, as there is no reflector on the seismic reflection profile at that depth, the increase in seismic velocity must occur over a broad depth interval.

The Brompton-Cameron composite terrane is interpreted to be entirely allochthonous upon the Laurentian craton. Neither of the faults that bound this terrane is a strong seismic reflector. The Brompton line that forms the northwestern terrane boundary was mapped as a near-vertical thrust fault zone (St.-Julien and Slivitsky, 1985; Moench and St.-Julien, 1989, fig. 1, p. 21). The fault zone is not a seismic reflector near the surface (St.-Julien and others, 1983) but appears at depth to be a weak reflector with southeast dip. The Squirtgun fault, a northwest-directed Acadian thrust fault, forms the boundary between rocks of the Brompton-Cameron and Central Maine terranes. This fault is not imaged near the surface on seismic reflection line 2 (Spencer and others, 1989) but may be one of the many southeast-dipping reflectors imaged at depth on seismic reflection profile 3A.

Relations among constituent terranes of the Brompton-Cameron composite terrane are complex, and their history is only partly deciphered. The Quebec ophiolite and related Cambrian intrusive rocks are overlain by mélanges and olistostromes that are as much as 8 km thick (Hébert and Laurent, 1989). This tract of rocks formed in the oceanic tectonic setting of the Iapetus Ocean, which was of unknown width, although commonly inferred to have been many hundreds of kilometers wide. Geochemical and isotopic evidence (Laurent and Hébert, 1989; Harnois and others, 1990; Church, 1987; Crocket and Oshin, 1987; Shaw and Wasserburg, 1984) suggests that the igneous rocks associated with the Quebec ophiolite were formed by subduction-related magmatism in a fore-arc basin. This would require the opening and closing of an oceanic tract at least hundreds of kilometers wide. However apparent time constraints of formation in the Cambrian and closure in the Middle Ordovician makes this scenerio problematic. Possibly significant strike-slip motion along some faults occurred to bring oceanic crust closer to Laurentia. A small tract of early Paleozoic calc-alkaline island-arc volcanic rocks has been thrust northwest an unknown distance to its present position over the ophiolite-mélange terrane. The Chain Lakes continental terrane with its cover sequence could not have been far outboard from the Quebec ophiolite and the early Paleozoic island-arc terrane because kilometer-size blocks of distinctive Chain Lakes rocks are found within the Quebec olistostromes, notably in the small triangular region on the northeastern side of the transect (Cousineau, 1991). The constituent Brompton-Cameron (Boundary Mountains) terranes were joined together and transported northwesterly by the late Early and Middle Ordovician Taconian orogeny (St.-Julien and Hubert, 1975; Spencer and others, 1989). Their space-time relations are shown on the tectonostratigraphic evolution diagram. The Hurricane Mountain belt of ophiolites and olistostromes and other constituent terranes beneath the Late Ordovician and younger sedimentary rocks of the Central Maine composite terrane may have been joined to the Brompton-Cameron terrane prior to the accretion of that composite terrane to Laurentia during the Taconian orogeny (Boone and Boudette, 1989).

Terranes within the Brompton-Cameron composite terrane deformed during the Taconian orogeny are unconformably overlain by a sequence of black pelite and basalt of probable Silurian age, taken here to have been deposited in an incipient rift, the Frontenac rift. Conformably overlying rocks (Marvinney, 1989) of the Connecticut Valley trough, a successor basin, include Early Devonian rocks (Hueber and others, 1990) in part coeval with rocks of another successor basin to the southeast in the Central Maine synclinorium. These rocks are thin above the Chain Lakes block, but thicken markedly southeastward across a tectonic hinge zone (Boone and others, 1970; Hatch and others, 1983; Moench and Pankiwskyj, 1988). The tectonic hinge may have resulted from the buoyant rise of the overthickened crust to the northwest of the Central Maine synclinorium after the Taconian orogeny (Stewart, 1989). Younger rocks overlying the Chain Lakes block include sandstones, some with volcanic clasts, and thick rhyolites and garnet rhyolites of late Early Devonian (Siegenian) age (Rankin, 1968, 1980). Such garnet rhyolites can originate only by fractional melting of felsic rocks at lower-crustal or upper-mantle pressures, probably at depths of more than 50 km. This implies that a very thick crust was formed during Taconian and early Acadian compression and by crustal loading of thrust sheets. On the tectonic setting map, these distinctive rhyolites are shown to result from continent-continent collision. The onset of the Acadian orogeny terminated sedimentation. The crust was significantly foreshortened by tight folding and extensive thrust faulting. Associated metamorphic grades increase to the southwest to facies as high as medium amphibolite. Numerous calc-alkaline plutons were emplaced immediately after the major deformational phase of the

Acadian orogeny. The depth of pluton emplacement also increases to the southwest; southwest of the transect in Maine, both high-grade Carboniferous metamorphism and granitic plutonism resulted from crustal melting because of overthickening during the Acadian orogeny (DeYoreo and others, 1989). These authors conclude that regional uplift of overthickened crust across the transect in Maine was approximately 6 km per 100 km along regional strike to the southwest. Subsequently, steep normal faults, such as those along the northwestern side of the Chain Lakes block with displacements as much as 5 km, formed in late Paleozoic and early Mesozoic times. Finally, a composite pluton of gabbro and granite was emplaced late in the Mesozoic (Eby, 1984).

Central Maine Composite Terrane

In the Central Maine composite terrane rocks older than Late Ordovician that were deposited in different tectonic settings are covered by the thick (up to ~13 km) sequence of conglomerates, sandstones, and calcareous siltstones of the Central Maine synclinorium, a post-Taconian successor basin. The volume of the rocks of the Central Maine synclinorium is enormous and possibly once amounted to as much as 2200 km^3 per km of strike length along the synclinorium (Stewart, 1989). Most of these rocks were deposited in Silurian and Early Devonian time and were severely deformed shortly thereafter during the Acadian orogeny. Thrust faults bound both margins of the Central Maine composite terrane. The suture line between the Brompton-Cameron (Boundary Mountains) composite terrane and the Central Maine composite terrane is a northwest-directed Penobscottian or Taconian thrust fault, the Squirtgun fault, that forms the northwestern boundary of the Hurricane Mountain belt. This fault was reactivated in the Acadian orogeny. The Hackmatack Pond thrust fault and related faults mark the southeastern boundary of the Central Maine composite terrane. Clearly the Central Maine composite terrane had been assembled by the end of the Acadian orogeny. Unfortunately, most of the relations between the component terranes and their basements beneath the blanketing sediments of the Central Maine synclinorium are poorly known or conjectural despite the availability of modern high-quality seismic reflection and refraction data. Some of the relations shown on the tectonostratigraphic diagram were projected along strike into the transect.

The Hurricane Mountain ophiolite and olistostromes belt (Boone and Boudette, 1989; Boone and others, 1989) is as much as 4 km thick and differs from the Quebec ophiolites and olistostromes in the rock types of the ophiolite sequence and in the clast assemblages of the olistostromes. The Hurricane Mountain ophiolite may be slightly younger in age, possibly Middle Cambrian. In addition, the geochemistry of the Hurricane Mountain ophiolite differs from that of the Quebec ophiolite (Coish and Rogers, 1987). The lower part is chemically similar to island arc rocks and was interpreted to have formed near a subduction zone. The upper part is chemically similar to midocean ridge rocks and may have formed in a proximal marginal basin. The ophiolite and olistostromes are overlain by a volcaniclastic flysch carapace, older than Middle Ordovician, that may be several kilometers thick.

Several different suites of Cambrian or Ordovician rocks underlie the Central Maine composite terrane. Southeast of the Hurricane Mountain belt, volcanogenic rocks of the Bronson Hill anticlinorium crop out along the southwestern edge of the transect (see fig. 1). These rocks are interpreted to be a Middle and Late Ordovician volcanic arc (Tucker and Robinson, 1990; Leo, 1985, 1991) and thus represent a tectonic setting different from that of the Hurricane Mountain belt. Compressional orogenic movements brought these two volcanogenic assemblages into contact, but the timing and nature of this event is controversial (Boone and Boudette, 1989; Tucker and Robinson, 1990). Thick sections of Cambrian and Ordovician sedimentary rocks are exposed within the Central Maine composite terrane northeast of the transect (Osberg and others, 1985) and are shown only on the tectonostratigraphic evolution diagram. Some of these rocks contain distinctive fauna of non-North American type (Neuman and others, 1989) as well as evidence of a pre-Taconian deformation (Neuman and Max, 1989). Boone and Boudette (1989) interpret these rocks as a separate terrane that was joined to terranes to the northwest during the Penobscottian orogeny. Lower Paleozoic rocks of the Nashoba-Casco-Miramichi composite terrane are known from magnetic and seismic reflection data to underlie the Central Maine composite terrane for at least 25 km northwest of the Hackmatack Pond fault.

The nature of the Precambrian basement beneath the Central Maine composite terrane is poorly known, although it is sialic and appears to be broadly homogeneous on the basis of its seismic refraction velocity and its internal seismic reflectors. Where basement is exposed some distance from this transect in the core of one of the domes on the Bronson Hill anticlinorium (the Pelham dome in northern Massachusetts), high-grade granulites, not unlike Grenvillian gneisses, are found, but have a U-Pb zircon age of 613±3 Ma (Tucker and Robinson, 1990). The Grenvillian crust beneath the Brompton-Cameron composite terrane is known to have been extended prior to rupturing to form the Iapetus Ocean and a passive margin at about 650 to 570 Ma (Rankin and others, 1989). As illustrated on our interpretative cross section for the present Atlantic passive margin, such extension drastically thinned the continental crust prior to rupture. This crust also was heated during the Late Proterozoic extension and may have reached granulite facies at midcrustal depths, which would have reset most isotopic systems. Hence, the core of the Pelham dome may be Grenvillian granulite that was reset during Iapetan crustal rupture. This interpretation also is supported by lead isotopic data from zircons in younger granitic plutons in New Hampshire (Harrison and others, 1987) and by characteristic lead isotopic compositions of feldspars from plutons that formed by partial melting of deep crustal rocks in the northern Appalachians (Ayuso and Bevier, 1991). The isotopic composition of lead in feldspars changes gradually across the Central Maine composite terrane but has a more uniform radiogenic composition near the coastline in the Atlantica[1] composite terrane (Ayuso and Bevier, 1991). One interpretation of this change is that a crustal boundary or suture exists beneath the Central Maine composite terrane. However, the gradual nature of the change does not permit the determination of a precise location for the suture or whether one type of crust might overlie another of different lead isotopic composition. It has been suggested previously that a third type of basement (the "Central" continental basement of Osberg, 1978, or "X" of Zen, 1983) lies beneath the Central Maine composite terrane. We therefore show this other basement by a different color on our interpretive cross section. However, as Nicholas Rast (oral commun., 1989) has remarked, a third type of basement between very long segments of the two known basements presents a serious topological problem in the

[1]Atlantica has been proposed by Zen and others (1986; see also Zen, 1989, 1991) as a more inclusive name for terranes commonly called Avalonian. This topic is discussed in greater detail below in connection with the description of numerous terranes within the Atlantica composite terrane.

context of the genesis of the entire Appalachian-Caledonide orogen. Given the available geologic, isotopic, and seismic evidence, and our new information on the nature and post-Paleozoic age of the reformation of the lower crust, we are skeptical that a third type of basement actually exists. We prefer to interpret most of the Late Proterozoic ages of basement rocks in regions that experienced large amounts of extension in the Late Proterozoic and extensive deformation in the Paleozoic to be reset Grenvillian ages. We reserve Atlantican basement to be those rocks upon which sediments were deposited at high paleolatitudes that contain a distinctive Late Proterozoic (Ediacaran) to Early Devonian fauna. As we think that the lower crust gained its principal features such as thickness and internal structures during late Paleozoic and early Mesozoic extensional processes, we have shown it separately, for reasons given in the section on the Evolution of the Present Crustal Framework.

Several large normal faults break the rocks of the Central Maine synclinorium and its basement to midcrustal depths based on seismic reflection profiles and shown on the interpretive cross section from kilometer 335 to kilometer 375 and from kilometer 290 to kilometer 310. These faults are post-metamorphic and likely are Mesozoic in age. A buried mafic pluton in the upper crust near Waterville, Maine, was inferred from gravity data by Potts and Doll (1987) and is shown as Mesozoic in age on the geologic cross section, although it could be of late Paleozoic age. The geologic map of Maine (Osberg and others, 1985) shows other such plutons in southwestern Maine. McHone (1978, 1984) reports numerous Mesozoic dikes oriented normal to the principal northwest-southeast direction of extension in the transect region, although these dikes are too small to be shown on the geologic map.

Nashoba-Casco-Miramichi Composite Terrane

The Nashoba-Casco-Miramichi composite terrane (Zen and others, 1986) is the complex, narrow, fault-bounded Orrington-Liberty belt of lower Paleozoic metavolcanic and metasedimentary rocks that was deposited upon a sialic non-Grenvillian basement (Ayuso and Bevier, 1991). Basement is neither exposed in the transect nor for hundreds of kilometers along strike to the northeast but may be like the Late Proterozoic metasedimentary and metavolcanic rocks in eastern Massachusetts and Connecticut (Zen, 1989). In the transect the Orrington-Liberty belt contains Ordovician metasedimentary and felsic and mafic metavolcanic rocks that have been metamorphosed to kyanite grade (Hussey, 1988) and Silurian anatectic granites and migmatites that were modified by middle- to high-grade amphibolite Acadian metamorphism and granitic plutonism. The seismic reflection profile (kilometer 350 to kilometer 435) shows many strong reflectors in the crust that dip northwestward beneath the Central Maine composite terrane. These are interpreted to be stacked, southeast-directed thrust sheets of a Silurian thrust belt whose character was not obliterated by Acadian or subsequent events. In the northern Appalachians, evidence for a strong Silurian tectonic belt along the southeastern side of the orogen along strike is extensive (Hepburn and others, 1987; Van der Pluijm and Van Staal, 1988; Stewart, 1989; Dunning and others, 1990; Bevier and Whalen, 1990; Ayuso and Bevier, 1991).

The near-vertical Norumbega fault zone within the Nashoba-Casco-Miramichi terrane experienced 10 to 20 km of right lateral strike-slip movement (Wones, 1978). It extends across the transect and for hundreds of kilometers both southwest and northeast. It has had a long history of movement for not only are conglomerates of Mississippian(?) age found in the fault zone but it cuts Carboniferous plutons. Both dip-slip and strike-slip movement along this fault continued into the Mesozoic (West and others, 1988). Distinctive lithotectonic sequences of rift volcanic and successor basin rocks are juxtaposed in the Norumbega fault zone. Metamorphic assemblages in pre-Silurian rocks northwest of the fault zone experienced as much as 4 kb higher pressures than rocks to the southeast. There is, however, no seismic reflection or refraction evidence in the map area that the Norumbega fault cuts entirely through the present crust; instead it appears to be cut off by a Mesozoic listric fault at a depth of about 20 km (see the interpretive cross section). Doll and others (1989) describe seismic reflection profiles that cross part of the Norumbega fault zone about 150 km northeast of the transect profile. There the Norumbega fault zone cuts through the crust and appears to offset the Moho by approximately 1.5 km. The complex history of motions along the Norumbega fault zone remains to be deciphered, but it is not the fundamental suture with the Atlantican continental block that many workers have taken it to be.

The Hackmatack Pond fault forms the northwestern boundary of the Nashoba-Casco-Miramichi composite terrane. A strong seismic reflector that originates at the mapped fault trace near kilometer 370 dips northwest at about 25° to depths of about 10 km, although it is offset at kilometer 365 at a depth of about 3 km by a normal fault with about 2 km of south-side-down movement. Metasedimentary rocks of the Central Maine synclinorium lie above the reflector, and rocks of the Falmouth-Brunswick sequence of the Cushing Formation (Hussey, 1988) occur below it. The Falmouth-Brunswick rocks are magnetic sulfidic gneisses, granofels, and amphibolites, metamorphosed equivalents of volcanic rocks, possibly of Early Ordovician age (Olszewski and Gaudette, 1988). These rocks contain kyanite and staurolite and have been metamorphosed to significantly higher pressures and temperatures than rocks of the Central Maine synclinorium. Newberg (1985) and Hussey (1988) interpret the Hackmatack Pond fault as a southeast-directed thrust fault. Osberg (1988) and Osberg and others (1989) considered it to be an unconformity; but in section *G–G′* in Hatcher and others (1990), Osberg shows it as a folded thrust fault. Our data support its interpretation as a southeast-directed thrust fault. The reflection from the fault trace at depth crosses the seismic reflection profile as a nearly linear feature and is subparallel to several strong deeper reflectors shown on the seismic reflection profile between kilometer 350 and kilometer 387 that also are interpreted to be thrust faults. One of the deeper reflectors reaches the surface along the Liberty-Orrington fault at kilometer 387. The magnetic anomaly map and especially bandpass-filtered intermediate wavelength magnetic data (J.D. Phillips, unpub. data) show that the fault surface at the top of the magnetic metavolcanic Ordovician rocks has broad undulations but dips northwest overall like the strong seismic reflector.

The southeastern border of the Nashoba-Casco-Miramichi composite terrane is marked by northwest-directed thrust faults, the principal of which is the Sennebec Pond fault (St. George fault of Berry and Osberg, 1989). Thrusting along this fault brought a suite of early Paleozoic slope-rise conglomerates, impure carbonate rocks, sulfidic pelitic rocks, and minor basalt onto the earlier formed buttress of the Orrington-Liberty belt. Seismic reflection profile 4 has a moderately strong reflector that projects to the trace of the Sennebec Pond fault. The fault dips to the southeast at about 20° and is only imaged to approximately 6 km where the reflection profile ends. A similar reflector imaged at 8 to 10 km depth on seismic reflection profile 8 beneath western Penobscot Bay may be the base of an obducted sequence of Atlantican rocks and, if so, indicates that the Turtle Head fault is not the fundamental suture between different continental blocks postulated by Stewart and Lux (1988).

Atlantica Composite Terrane

The Atlantica composite terrane southeast of the Sennebec Pond fault consists of several component terranes. Each terrane contains Cambrian and Ordovician strata deposited in different tectonic settings. Although each terrane has a different geologic history, all are non-Laurentian. This assemblage of terranes shares major paleogeographic and faunal similarities that differ substantially from those of North American terranes and areas outboard from these terranes during the Late Proterozoic and early Paleozoic. Zen and others' (1986) concept of Atlantica derives from the early recognition by Walcott (1889, 1891) that Cambrian fauna from eastern New England and the Maritime Provinces of Canada constituted a distinctive Atlantic Province quite different from faunas of typical North American aspect. As summarized by Neuman and others (1989), faunal distinctions have been found to occur from at least the Late Proterozoic (Ediacaran) through, locally, the earliest Devonian (Gedinnian). The composite terrane here called Atlantica has been called Avalonia or Avalon by others (for example, Nance, 1986; Rast and Skehan, 1983) and has been interpreted to be composite as faunal and stratigraphic evidence from the typical Avalonian platformal rocks argues for their separation into two, or more, terranes (Secor and others, 1983). In the region called Atlantica on this transect, several other terranes share paleogeographic and faunal similarities with parts of "Avalonia" but are from different tectonic settings and have different geologic histories, thus demonstrating the composite nature of Atlantica. Pratt and Waldron (1991) described an Acado-Baltic (our Atlantican) faunal assemblage in sandy turbidite of the Goldenville Formation of the cover sequence of Meguma terrane. Therefore the Meguma terrane could properly be another component of the Atlantica composite terrane as used by Zen (1989). However we distinguished the Meguma terrane on this transect prior to the publication of Pratt and Waldron's paper, and because its basement, although poorly known, may be different.

Onshore in this transect, the Atlantica composite terrane southeast of the Sennebec Pond thrust fault contains the St. Croix and the Ellsworth-Mascarene terranes. There is little specific evidence to identify the offshore terranes, but samples collected by the submersible *Alvin* are undeformed alkalic plutonic rocks typical of the Boston-Avalon zone of Hepburn and others (1987), which are equivalent to most of the rocks of the Esmond-Dedham terrane of Gromet (1989). It is not known if the high-grade basement gneisses typical of Gromet's Hope Valley terrane are present, but the highly magnetic rocks within and just southeast of the Gulf of Maine fault zone (Hutchinson and others, 1989) may be gneisses like those of the Hope Valley terrane.

The St. Croix terrane (Ludman, 1987) contains rocks of Cambrian through Late Ordovician age. Conglomerates, quartzite, impure carbonate rocks, and abundant black pelitic rocks and minor bimodal volcanic rocks suggest a continental slope-rise sequence that passes upward into a marginal basin formed upon rifted continental crust (Fyffe and others, 1988). We speculate that the continental basement involved was part of Atlantica and that the slope-rise sediments were deposited on its western side. However, no fossils having diagnostic Atlantica affinities have been found in these rocks. The St. Croix terrane had been deformed, metamorphosed, and accreted to the Ellsworth-Mascarene terrane by Middle Silurian time (West and others, 1992), and Late Silurian conglomerate was deposited across the boundary faults near the Maine-New Brunswick boundary (Gates, 1989).

The accretion of the already-joined St. Croix and Ellsworth-Mascarene terranes to the Nashoba-Casco-Miramichi composite terrane occurred early in the Acadian orogeny. Accretion may have resulted from a continent-continent collision that caused large thrust faults such as the Sennebec Pond fault as well as faults that combined strike-slip and thrust movements such as the Turtle Head fault zone. The Turtle Head fault zone separates the St. Croix terrane from the Islesboro block, which is taken to be a fragment of the Ellsworth-Mascarene terrane. The Islesboro block has a Late Proterozoic basement and a thin platformal cover sequence of conglomerate, quartzite, and impure carbonate rocks and slate with some volcanogenic sedimentary rocks interpreted to be distal to typical Ellsworth-Mascarene rocks (Stewart and Lux, 1988; Stewart, 1974). Southeast of the Turtle Head fault zone, blocks of St. Croix terrane rocks are juxtaposed by strike-slip faulting with blocks of the volcaniclastic lithologies of the Ellsworth-Mascarene terrane. Rocks in the St. Croix terrane northwest of the Turtle Head fault zone reached upper amphibolite facies of metamorphism prior to completion of right-lateral strike-slip and thrusting movement along the fault zone. These high-grade metamorphic rocks of the St. Croix terrane yield Silurian cooling ages and have not been subjected to any significant Acadian metamorphism (D.P. West, Jr., written commun., 1992). Lower Paleozoic rocks southeast of the Turtle Head fault zone have not exceeded greenschist facies of metamorphism (Guidotti, 1989). Plutons intruded across the Sennebec Pond and Turtle Head faults establish that this part of Atlantica was joined to North America by Middle Devonian time.

The Ellsworth-Mascarene terrane in the transect has a continental Proterozoic basement that is intruded by Late Proterozoic granitic pegmatite (Stewart and Lux, 1988). These rocks are unconformably overlain by three volcanic-rich sequences. The oldest sequence is unfossiliferous and consists predominantly of a bimodal continental rift suite of metabasalt and metarhyolite containing interbedded metasiltstone, quartzite, and marble. On the basis of field mapping and seismic reflection data, this sequence is at least 7 km thick. The rocks are Cambrian or Ordovician in age and were recrystallized to greenschist, probably in the Early Ordovician. A younger volcanic sequence[2] is relatively unmetamorphosed. This sequence is mostly a bimodal basalt-rhyolite suite that contains interbedded shallow marine volcaniclastic sedimentary rocks. It is at least 2 km thick in the transect area. The youngest volcanic rocks unconformably overlie the older volcanics and contain metamorphosed clasts of the older volcanic rocks and a characteristic Atlantica (Baltic) fauna of Middle Silurian to Lower Devonian age. Numerous Silurian, Devonian, and Mississippian plutons have intruded all three volcanic sequences.

Seismic reflection and refraction data indicate that the St. Croix and Ellsworth-Mascarene terranes have been thrust northwestward to their present positions. Seismic reflection profiles 4 and 8 show reflectors that dip to the southeast and can be interpreted to mark the bases of these terranes. Seismic refraction line 8, as interpreted by Luetgert and Mann (1990), suggests that the lower Paleozoic rocks of the Ellsworth-Mascarene terrane rest on basement at a depth of 7 to 10 km, and that at about 20 km, velocities increase to those characteristic of the lower crust. However, Luetgert and

[2]Shown as Silurian and Devonian age on the transect geologic map, a zircon sample has been dated at 490±4 Ma (Early Ordovician) by McLeod and others (1991).

Mann found no unambiguous refraction evidence for a midcrustal detachment fault as is shown in the interpretive cross section on the basis of reflection profiles 4 and 8. Seismic reflection profile 3B across the strike of the Ellsworth-Mascarene terrane (Stewart and others, 1985) images numerous subhorizontal reflectors throughout the upper crust at depths equivalent to the 7- to 10-km-thick Paleozoic cover sequences known from regional mapping and Luetgert and Mann's refraction studies. One of the deepest of these reflectors is moderately strong to weak and north of the end of seismic reflection profile 3B projects toward the trace of the Sennebec Pond thrust fault. It could be interpreted to be the detachment fault along which these rocks were transported northwestward. There are few reflectors in the interval below the reflector interpreted to be the Sennebec Pond thrust fault and depths of about 20 km. Below 20 km numerous strong reflectors occur throughout the lower crust to the Moho.

A major fault zone mapped in southern New Brunswick and just off the southeastern Maine coast is projected to cross the transect along a band of strong magnetic highs (Hutchinson and others, 1989). Strong seismic reflectors dipping 25° to 30° south at mid- to deep-crustal levels reach the surface in the same region. They have been interpreted (Hutchinson and others, 1988) to be thrust faults of the Gulf of Maine fault zone, a probable Alleghanian feature by analogy to the geology of eastern Massachusetts and Connecticut (Gromet, 1989). Still another band of magnetic highs and the projection of another seismic reflection to the surface mark the trace of the Fundy fault. This fault had normal slip motion when it formed the large East Narragansett rift basin, and later during the Carboniferous, experienced Alleghanian thrust motion as well. This basin crosses the transect as a half graben. Normal slip along the Fundy fault also may have formed a Mesozoic rift basin to the northeast of the transect. The region between the Fundy fault and the northwestern side of the Gulf of Maine fault zone may well be made up of high-grade basement gneisses and metamorphosed cover rocks like those in the Hope Valley terrane of southeastern New England. Western and northwestern dips are common in the Hope Valley terrane and reflect Alleghanian compression as well as subsequent extension (Goldstein, 1989; Gromet, 1989; Getty and Gromet, 1991). The interpretive cross section between kilometer 475 and kilometer 525 shows imbrication of the northwest-dipping structures by younger northwestward-directed thrusts of the Esmond-Dedham terrane. In Massachusetts, geochronologic evidence indicates that the Esmond-Dedham terrane was joined to cratonal North America between about 370 Ma (Late Devonian) and 340 Ma (Mississippian) (Hepburn and others, 1987). In Massachusetts, fragments of the Ellsworth-Mascarene terrane (the Newbury Volcanic Complex) are caught in faults between the Nashoba-Casco-Miramichi composite terrane and the Esmond-Dedham terrane. The keel of the Mesozoic Middleton basin formed along another fault is exposed nearby. It is clear that faults in the Esmond-Dedham terrane have had a long history of movement as a result of several episodes of compression, transtension, and extension.

Weak, discontinuous arcuate subhorizontal reflections are imaged on the seismic reflection profile below the refraction Moho in the interval between kilometer 485 and kilometer 530. These resemble sub-Moho reflections observed near kilometer 100 to 150 on deep seismic reflection profiles 86–1 and 86–2 in the Gulf of St. Lawrence (Marillier and others, 1989). These reflections occur near the deepest parts of the Carboniferous Magdalen basin where Marillier and Verhoef (1989) proposed, from consideration of gravity and seismic refraction velocity data together with the depth to reflection Moho, that a thick layer of mafic rocks having a high density (3.05 g/cm^3) and a high velocity (7.35 km/s) underplated the lower crust during the Carboniferous extension that formed the basin. A swath of broad high magnetic and gravity anomalies extends southwest from the deep part of the Magdalen basin across southern New Brunswick and southeasternmost Maine and passes close to the coastline in the region of this transect, just northwest of the Gulf of Maine fault zone. Mississippian plutons intruded the rocks of this swath in southwestern New Brunswick, and the Mississippian and Pennsylvanian Fredericton basin lies just north of it. The extensional Carboniferous East Narragansett basin is underlain by the arcuate sub-Moho reflections. It is speculated on the interpretive cross section that the sub-Moho reflections result from Carboniferous mafic underplating. The Gulf of Maine fault zone, interpreted by Hutchinson and others (1988) to be the late Paleozoic compressional Alleghanian (Variscan) front, also is underlain by these sub-Moho reflectors. At the very least, the reflectors imply that the upper mantle is laterally heterogeneous locally.

The Esmond-Dedham terrane is mostly underlain by Late Proterozoic calc-alkaline volcanic rocks and coeval plutonic rocks that intruded mildly metamorphosed older alkaline volcanic rocks and quartz-rich, slope-rise sedimentary rocks deposited in rifts along the margin of an extending craton. Sedimentary cover rocks of Late Proterozoic age in rift basins consist of conglomerate, tillite, and basaltic to intermediate volcanic rocks. They are overlain by platformal Cambrian sedimentary rocks containing characteristic Atlantic (Baltic) fauna representative of cold high-latitude (south polar?) marine environments. All these rocks were intruded by Late Ordovician to Devonian peralkaline plutons that have minor associated volcanic rocks. One of these plutons, the Cashes Ledge, forms a large magnetic high on the southwestern side of the transect. All rocks of the Esmond-Dedham terrane are little deformed, and there was no regional metamorphism above the lower greenschist facies.

Meguma Terrane

The Cape Cod (Nauset) fault system is taken to be the boundary between the Esmond-Dedham block of the Atlantica composite terrane and the Meguma terrane, which underlies the bulk of Nova Scotia. Magnetic and gravity anomaly maps allow the identification of the Meguma terrane to at least as far west as 67° W. longitude (Pe-Piper and Loncarevic, 1989). But farther from land and especially where younger sediments are present, the location of the terrane boundary and even the presence of the Meguma terrane beneath the sea are not well established. Our choice of location is the northwestern border of a very extensive granitic batholith, the Franklin batholith, located near kilometer 610 to kilometer 640 by Hutchinson and others (1988, p. 178). The Franklin batholith is comparable to the South Mountain batholith that intrudes the Meguma terrane in Nova Scotia.

The Meguma terrane in Nova Scotia is made up of a very thick section (as much as 10 km) of turbiditic sandstone and shale of early Paleozoic age, unconformably overlain by shallow water marine sedimentary and bimodal volcanic rocks (Keppie, 1989). Large peraluminous granitic batholiths as young as Late Devonian intrude these rocks. Some of these plutons have been deformed by strike-slip faults associated with the accretion of the Meguma terrane to North America. Clasts from these plutons occur in conglomerates in deep Mississippian basins that overlap the suture. Within this transect no drill hole has penetrated the crystalline basement interpreted here as a southwestern extension of the

Meguma terrane, so its nature is virtually unknown. However, immediately to the west of the transect, at about 68°20′ W., 40°50′ N., drill hole G–1 bottomed in graphitic slate, schist, and phyllite dated at 540 to 450 Ma that may correspond to Meguma rocks (Klitgord and others, 1988; Schlee and Klitgord, 1988; Schenk, 1991).

South of approximately 42° N. latitude, this transect overlaps Continent-Ocean Transect E–1 by Thompson and others (in press). The interpretive cross sections on both transects in this area are based on seismic reflection profile 19. Thus, from kilometer 700 southeast to its end, our cross section coincides with their cross section. However, our transect stops north of theirs on what Klitgord and others (1988) call marginal oceanic crust. The region common to both areas is discussed by Klitgord and others (1988). The discussion of the Gulf of Maine embayment and Georges Bank basin is based on Klitgord and others (1988).

Mesozoic extension was so extensive that it led to continental breakup. Extreme extension is the dominant feature of the Meguma terrane within the transect, which is quite different from the type Meguma terrane in Nova Scotia. Numerous rift basins were formed during the Triassic and Early Jurassic. They are elongated northeast-southwest along the regional structural grain of the basement and were probably controlled by the reactivation of older faults. The rift basins contain as much as 5 km of terrestrial sediments, and some contain voluminous basaltic volcanic rocks as well as plutonic rocks locally. Some of the rift basins also contain evaporite beds near the top of the section. The rift basins in the northwestern part of Meguma on this transect have been deeply eroded so that only their lower parts remain. A tectonic hinge zone strikes northeast-southwest across the transect near kilometer 700. Southeast of the hinge zone, the rift basins are more completely preserved. The hinge zone developed shortly before crustal breakup produced oceanic crust (late Early Jurassic=175 Ma). The Georges Bank basin is a marginal basin that is a successor basin to the rift basins. It contains as much as 7.5 km of post-rift sediments in the transect region and becomes even deeper to the southwest. Northwest of the hinge zone, onlapping, platformal, Coastal Plain sediments are up to 2 km thick.

The Jurassic and Early Cretaceous paleoshelf edge was a carbonate-bank complex that built upward and seaward as the crust cooled after continental rupture. The carbonate-bank complex separates shelf sediment deposits from slope-rise deposits seaward. Elongate salt diapirs or ridges intrude this sediment wedge just seaward of the paleoshelf edge. The continent-ocean boundary is marked by the East Coast Magnetic Anomaly, which is near the present continental shelf edge.

Seismic data suggest that the continental crust of Meguma was thinned to about half by extension in the Mesozoic prior to rupture. The propagation of seismic energy is impeded by the thick blanket of post-rupture sediments, and no deep crustal refraction studies have yet been made in this region. By analogy to seismic refraction studies along the same margin off the Southeastern United States it is probable that the deep crust was extensively intruded by mafic dikes and that it has higher average density (and seismic refraction velocity) than continental crust (Trehu and others, 1989; Austin and others, 1990). Underplating by upwelling mafic magma generated by pressure relief during rupture is quite probable but not demonstrated beneath Georges Bank. If underplated, the lower crust could be continuous from broken former continental to oceanic crust. We surmise that the strong magnetic anomaly about kilometer 830 represents the rise of Mesozoic mafic magma to form a magnetic pluton in the rifted crust. The interpretive cross section is based on control from nearby deep wells and the limited seismic reflection data.

Atlantic Oceanic Crust

The continent-ocean boundary, marked by the East Coast Magnetic Anomaly, is poorly defined by anomalies at the southeastern end of the transect. Its label on the magnetic map follows the trend shown on maps for larger regions (Magnetic Anomaly Map of North America Committee, 1987). The strong magnetic anomaly at the southwestern corner of the transect is associated with Bear Seamount, one of the extensive chain of New England seamounts of Cretaceous age.

Oceanic crust at the southern end of the transect contains seaward-dipping seismic reflectors (mafic volcanics?) in the basement rocks beneath the sedimentary cover. This type of oceanic crust, called marginal oceanic crust by Klitgord and others (1988), forms a swath about 25 km wide just seaward of the salt diapirs. Individual reflectors can be traced only part of the way across this swath. All the reflectors merge upward into a conformable unit that covers the entire swath. The reflectors disappear at depth, and the zone of reflectors has no distinct base. This type of oceanic crust lacks the hyperbolic echoes in seismic reflection profiles that are typical of oceanic crust generated by sea-floor spreading processes.

EVOLUTION OF THE PRESENT CRUSTAL FRAMEWORK

The cumulative effect of the various Paleozoic orogenies undoubtedly was the formation of thick (~60 km) continental crust as terrane after terrane was accreted. Structural features still preserved in the upper crust that attest to this thickening are (1) allochthonous sheets that can be imaged to depths of more than 20 km, (2) areas of younger metamorphic rocks that contain mineral assemblages indicative of pressures as high as 9 kb (~30 km) now exposed on the surface, and (3) extruded garnet rhyolite formed by fractional melting of felsic rocks deeper than 50 km. Our composite of seismic reflection profiles demonstrates that the upper crust still retains coherent regional structural dips; these may be opposed to each other in adjacent terranes as in the case of the Nashoba-Casco-Miramichi composite terrane and the Atlantica composite terrane between kilometer 325 and kilometer 550 on the interpretive cross section.

The crust beneath the craton today, however, lacks overthickened parts and is about 43 km thick over broad regions near the northern Appalachians. In this transect the composite of seismic reflection profiles shows a stepwise crustal thinning toward the ocean basin. Broad steps, 75 to hundreds of kilometers wide, and narrow risers, 2 to 4 km high and 10 to 20 km wide where the Moho shallows, are evident. Profiles elsewhere in the northern Appalachians (Marillier and others, 1989; Phinney and Roy-Chowdhury, 1989) also do not show overthickened crust, but show broad regions of crust having little variation in thickness and narrow regions where the Moho rises or falls from 2 km to as much as 6 km over a few tens of kilometers. But on these other profiles, there is no systematic rise seaward as shown by this transect. In the northern Appalachians, steps on the Moho are not simply related to any terrane boundary mapped at the surface. Since the late Paleozoic, deep crustal processes have obliterated evidence of earlier Paleozoic orogenies, and the crust has been restored to normal continental thickness overall but is thinned toward the ocean. Undoubtedly some of the continental crust has been removed by erosion. Estimates based on fission track dates

(Doherty and Lyons, 1980) suggest the erosion of about 6±1 km of rock since the Jurassic for western Maine. This must be a maximum. Obviously the amount of erosion was not uniform along the transect, as the Coastal Plain (shown by overprints on the geologic map) was the site of deposition for much of the time since continental breakup. Zen (1991) has described in detail the denudation in the New England Appalachians.

Tectonic restoration of normal crustal thickness had begun in middle Paleozoic times, as evidenced by the hinge line in the rocks of the Central Maine synclinorium, a Silurian and Early Devonian successor basin. Thermal recovery of thick crust led to partial melting and the diapiric rise of magma into and upon the cover sequences of deep basement, depleting the deep crust of quartz and alkali feldspar and thus imposing the compositional transition in the lower crust toward the higher density and seismic velocity observed now. The lower crust lost its stiffness as a result of added internal and external heat, some possibly resulting from mafic underplating. The upper crust remained stiff and brittle, although some heating may have occurred by the mechanical transport of heat by ascending magmatic diapirs and fluids. The softening of the lower crust together with global plate tectonic forces led to the beginning of crustal transtension in the Mississippian and the formation of deep half graben such as the East Narragansett basin near kilometer 530 on the interpretive cross section.

This transect demonstrates that the tectonic processes that disrupted Pangea, the supercontinent that was assembled near the end of the Paleozoic, imprinted their effects upon a swath of continental crust that is now about 650 km wide in the part still attached to North America. These effects increase in intensity from the craton toward the oceanic basin and are less apparent in the upper crust than in the lower crust. The effects in the upper crust are occult toward the craton and consist principally of normal faults and the intrusion of oriented dike swarms and small plutonic complexes. Only limited extension is indicated, possibly only a few percent. This appears to be the case from about kilometer 175 to kilometer 325 on the interpretive cross section. As the amount of extension in the upper crust increased, crustal blocks were rotated along favorably oriented older faults that were reactivated and probably become listric at midcrustal depths; rift basins were formed where the extension was localized in the upper crust. The upper crust may have been thinned by as much as 15 percent as is shown from about kilometer 325 to kilometer 625 on the interpretive cross section. More extensive rifting accompanied by basaltic volcanism occurred close to the locus of continental breakup. The upper crust may have been thinned as much as 40 percent when the breakup occurred. This thinning is shown from kilometer 625 to kilometer 875.

The effects of extension upon the lower crust and Moho were profound as is most evident in the seismic reflection and refraction data, as well as the broad southeasterly increase in the regional gravity field, most easily seen on the gravity profile. One of the most important consequences of extension was the demarcation of the lower crust itself. The top of the lower crust is defined by a gradual increase in the seismic compressional wave velocity from about 6.4 to about 6.8 km/s over a depth of as much as 4 km (Stewart and Luetgert, 1990). The top of the lower crust is not a reflector at near-vertical angles of incidence as can be seen by comparison of the seismic refraction and seismic reflection profiles. Seismic refraction data both parallel and normal to regional strike show that the top of the lower crust has little relief, rising smoothly from about 24 km depth under the Brompton-Cameron (Boundary Mountains) composite terrane to about 20 km beneath the northern part of the Atlantica composite terrane, a distance of over 300 km. The top of the lower crust is not offset by any terrane boundaries or faults, as seen on the interpretive cross section. Using seismic refraction data collected after this transect was completed, J.H. Luetgert (written commun., 1990) contoured the top of the lower crust for much of Maine, New Hampshire, and Vermont. Beneath this transect in Maine, Luetgert found that the top of the lower crust has a slight dip to the southwest, being as much as 2 km deeper near Farmington. Over a 600-km-wide region in northern New England, the relief on the top of the lower crust is only about 4 km and is not offset by any boundaries or faults.

The velocity change at the top of the lower crust is sufficient to be observed only within the part of the orogen that experienced late Paleozoic or Mesozoic extension, which is shown by oriented dike swarms, rift basins, or steps in the regional gravity gradient parallel to the regional structural grain. Because the seismic velocity in shallow Grenvillian crust is higher (~6.6 km/s), the top of the lower crust cannot be clearly identified therein (Hughes and Luetgert, 1991). The interpretation that the lower crust attained its present configuration during late Paleozoic and early Mesozoic extension is supported by several other lines of evidence.

Extension seems to have affected the properties of the Moho and the lower crust even where the effects in the upper crust may be occult. The Moho beneath the craton commonly is an indistinct gradational velocity change that does not yield a strong reflection at any angle of incidence or a sharp refraction. However, in regions where the orogen has been extended, a sharp crust-mantle boundary is reported (Taylor, 1989, fig. 5), which results in a stronger Moho signal. The transition can be seen from about kilometer 200 to kilometer 240 of the composite seismic reflection profile, and more distinctly in the more detailed figures in Spencer and others (1989). The Moho becomes stronger farther to the southeast and remains strong to near the end of the transect where it is lost beneath thick carbonate-bank sediments that inhibit transmission of reflected sound energy.

As the amount of extension increases, reflectors in the lower crust become more numerous and stronger, although commonly they are subhorizontal and discontinuous, and individual reflectors are only about 10 to 15 km long. This phenomenon has been observed widely, although seldom have corresponding refraction velocity profiles been available for the identification of the top of the lower crust, as is the case for this transect for over 300 km between kilometer ~200 and kilometer ~540. The numerous reflectors observed in the lower crust of this transect suggest that during deformation much of the lower crust yielded ductilely along subhorizontal anastomosing shear zones, causing the reflective velocity contrasts. Not all crustal blocks yielded in the same way. The bands of reflections in the Gulf of Maine fault zone and the Fundy fault pass continuously from the bottom of the upper crust through most of the lower crust until they merge into a layer of subhorizontal reflections, 3 or 4 km thick that marks the bottom of the lower crust and reflection Moho. The crust has the same thickness in this region as it does in blocks to the north and south, suggesting that ductile extension of the lower crust was concentrated at the bottom of the lower crust. It is possible that the dip of the bands of reflectors was reduced by southeastward plastic flow of the deep crust.

Two observations gained by comparing coincident seismic reflection and refraction data support the interpretation that most of the lower crust was homogenized by plastic flow and recrystallization during extension so that it lacks detectable through-going faults. The first is that the strong seismic reflector at the top of autochthonous Grenvillian crust beneath the Brompton-Cameron (Boundary Mountains) composite terrane dies out at the same depth as the top of the lower crust as determined by seismic refraction data. This would be expected if the boundary between the upper crust and the lower crust was a zone of rheological contrast that localized intense deformation and recrystallization, thus obliterating the strong fabric and compositional contrasts that cause the reflector. The disappearance of this reflector at the top of the lower crust makes interpretation of the southeastern extent of Grenvillian crust speculative; gravity and magnetic data are not definitive either. Geologic evidence from the northwestern end of the transect indicates that the Grenvillian crust was extended, so that it thinned and ruptured to form the Iapetan passive margin. The later thrust imbrication of thinned Grenvillian blocks, interpreted to underlie the Brompton-Cameron composite terrane, must represent a considerable shortening of the Iapetan passive margin. Geologic and isotopic evidence suggest that the basement that constitutes the upper crust of the Central Maine terrane may not be Grenvillian. However, the nature of the lower crust is not constrained. Therefore the interpretive cross section shows an arbitrary boundary for the southeastern limit of Grenvillian lower crust. Another possible interpretation is that the top of Grenvillian lower crust continues to deepen at about the dip of the décollement surface until it pinches out near kilometer 320 on the interpretive cross section. Such a tapered edge would be the result of extensive plastic flow, as the end of the ruptured Iapetan continental block probably would have been blunt-like crust at the present Atlantic continental edge.

The second line of evidence is that young normal faults that break the upper crust, such as the one in the southeastern part of the central Maine synclinorium (kilometer ~340), become listric and are lost at or near the top of the lower crust as indicated by the rapid increase of seismic refraction velocity at a depth of 20 km. A geographically coincident approximate 4-km rise of the Moho occurs below the footwall of this listric fault as indicated by well-constrained seismic reflection and refraction data. The intervening lower crust is strongly laminated with subhorizontal reflections. All of these observations support the interpretation of extensive plastic flow in the lower crust, in the direction of ultimate crustal rupture. Examples of the rise of Moho beneath the footwall of listric faults and the development of a well-laminated lower crust are numerous on seismic reflection profiles from the northern Appalachians (Phinney and Roy-Chowdhury, 1989; Marillier and others, 1989), although determinations of the refraction velocity changes are lacking. A conspicuous example occurs beneath the Franklin rift basin at about kilometer 675 on the interpretive cross section. Phinney and Roy-Chowdhury (1989, fig. 12d) give another good example from the eastern end of line USGS 36 on the Long Island platform beneath the early Mesozoic Nantucket rift basin. If the large normal displacements on mapped faults are considered (Lyons and others, 1991), COCORP line 6 in southwestern New Hampshire beneath the Bronson Hill anticlinorium and Central Maine synclinorium (Phinney and Roy-Chowdhury, 1989, fig. 16) can be interpreted to show the same phenomenon. Seismic refraction data from northern New Hampshire (Hughes and Luetgert, 1991) suggest that the top of the lower crust is at about 18 km depth. This corresponds closely to the depth on COCORP line 6 at which east-dipping reflectors, interpreted as décollements, die out, and where laminated subhorizontal reflectors appear in the lower crust in conjunction with a distinct Moho reflection.

The lower crust portion of the continental crust that has been extended more than 10 or 15 percent cannot be assigned with confidence to a terrane that lies above it. As reflectors that can be traced downward from structures at the surface are obliterated, identity is lost as one block of lower crust has much the same velocity structure and bulk composition as another. The problem becomes more difficult the greater the distance from the craton, as the lower crust is thinned significantly more than the upper crust (Keen and others, 1987). Hence we have shown lower crust in a different pattern from the upper crust outboard from the arbitrary boundary of Grenvillian lower crust. The isotopic evidence of Ayuso and Bevier (1991) best describes the crust as it was in the middle Paleozoic when it was being sampled by partial melting and the ascent of granitic magmas. These data are not directly applicable to the lower crust after it was reformed in the late Paleozoic and early Mesozoic. Suites of deep crustal xenoliths from Mesozoic dikes and plutons would provide more reliable samples of lower crustal compositional variation now. Deep seismic refraction velocities below 7.1 km/s show that the lower crust is of sialic composition, probably plagioclase-rich pyroxene and amphibole granulites. No tract of transitional or oceanic crust has been identified to underlie any terrane northwest of Meguma.

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