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FOLD-AND-THRUST BELTS IN SEDIMENTARY ROCKS.

PART 1: TYPICAL EXAMPLES

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ABSTRACT. The Appalachian Valley and Ridge province is a well known and early deciphered example of a very widespread type of deformational belt, in which a thick wedge-shaped (miogeosynclinal) prism of sedimentary rocks, deposited on the margin of a stable craton, has been thrown into long parallel folds generally cut by long parallel thrust faults of large displacement. Both folds and thrust faults verge fairly consistently toward the craton, the vergence resulting from the wedge shape of the prism. Such belts occur along the outer margins of most orogenic belts or mountain chains. Other classical examples are the Subalpine chains of France and the Helvetic zone of Switzerland, along with the nearby Jura, and the Canadian Rockies, to which should be added the west flank of the Ural and the western foothills of Taiwan. The belt in Taiwan is being formed now, diachronously from north to south.

Characteristically, the tectonics of such belts is thin-skinned; that is, the folds and thrust faults end downward in a main level of décollement and do not involve the cratonal basement on which the sedimentary prism was deposited or the lowest layers of the prism itself (the tegument) up to the main décollement level. Commonly there are other décollement levels higher in the prism, and their extent and distribution, controlled in turn by facies changes, synsedimentary faults, and so forth, have determined much of the structural pattern and style of the individual belts. Temperature gradients within the prism may also have brought about sharp changes in style of deformation, by activating or deactivating different deformation mechanisms in the décollement zones at certain temperature thresholds.

Under some of these belts (Canadian Rockies, western Taiwan), the basement with its tegument presents for the most part a gentle even slope down from the cratonal side toward the inner margin of the belt, a slope doubtless inherited from the floor of the original wedge-shaped prism but increased during deformation. Under other belts, the slope is broken by down-to-basin, commonly synsedimentary faults or by more complicated fault patterns (as in the Jura). In a few places, especially near promontories of the craton (recesses of the fold belt), slices of basement have been caught up in the thrusting. But in general basement shortening to match the obvious, large short-

ening in the sedimentary cover must have taken place in more interior parts of the orogenic belt.

The thrust faults in fold-and-thrust belts are often thought to form as the folds become tighter and more asymmetrical, but according to the hypothesis of John L. Rich (1934) the folds form as the thrust faults move and ramp from one décollement level to another. His hypothesis explains many features of the belts, notably the prevalence of folds of parallel but *not* concentric geometry, with flat-panel limbs and sharply curved or even angular hinges (several per fold), and it has predictive value. Doubtless both situations can arise in different belts or in different parts of the same belt.

A TYPE EXAMPLE—THE VALLEY AND RIDGE PROVINCE OF THE CENTRAL AND SOUTHERN APPALACHIANS

The Valley and Ridge province of the central and southern Appalachians (fig. 1) extends for about 1500 km northeast-southwest along the southeastern margin of the North American craton, from southeastern New York State to central Alabama. Its general structure has been well known and indeed accepted as the classical example of a fold-and-thrust belt ever since the superb description of the portion from Pennsylvania through Virginia by the brothers H. D. and W. B. Rogers in 1842 (Rogers and Rogers, 1843). Furthermore, they provided the first clear proof that the strong folding so common in mountain ranges results not from the individual uplift of each fold but by deformation within a single stress field involving the whole folded belt (they were less clear about the associated thrust faults). There have been a number of (mostly partial) descriptions since, notably the classical physiographic description by W. M. Davis (1889), who made the Appalachian landscape as well known as the Appalachian structure that underlies it. I too have described it, perhaps rather too exhaustively (Rodgers, 1970, chap. 3, p. 31–65); there have been advances in our knowledge since then, but the basics have not changed a great deal. Nevertheless, for this article I would like to recall the main facts and inferences.

The stratigraphy of the province is reasonably well known. It is underlain by Precambrian basement, consolidated in the Grenville cycle of deformation (1050 ± 100 Ma) or even earlier. A quartz-sandstone unit transgresses northwestward, cratonward, at the base of the Paleozoic sequence, but as it is not involved in the deformation it is not well known within the borders of the province; it is Lower Cambrian to the southeast (where in large areas uppermost Precambrian, post-basement, mainly clastic strata underlie it) but as young as Upper Cambrian on the craton northwest of the orogenic belt. There follows a sequence of shallow-water carbonate 1 to $2\frac{1}{2}$ km thick (made of limestone and dolostone in varying proportions) and ranging up to the Lower Ordovician. Coupled regressions and transgressions introduced sand and mud into this sequence at various levels, the most notable being across the Lower-Middle Cambrian boundary. Anhydrite beds are known here and there near that boundary, and salt may also have been present; the

A nearly but not quite universal disconformity (locally with low-angle discordance) separates "Lower" from "Middle" Ordovician strata.¹ Thereafter the stratigraphy becomes much more complex and facies changes abound. Perhaps the dominant feature is the appearance of several clastic wedges or deltaic complexes with eastern to southern, outboard, sources: commonly they begin with black shale, followed by turbidites, shallow-marine clastics, and continental redbeds. If the sea returned, it generally deposited clean quartz sandstone that transgressed eastward or southeastward, followed by shallow-marine clastics, and, unless another clastic wedge appeared fairly soon, by shallow-water limestone. In general shallow-water limestone continued to form farther west on the craton, beyond the reach of the clastic wedges, until the middle of the Carboniferous.

Despite the irregular distribution and overlap of the various clastic wedges, the total effect of all this sedimentation was to produce a triangular prism or wedge extending all along the Appalachian trend, its thin edge (perhaps 2 km thick) merging northwestward into the undeformed strata on the craton, its thick side (10 km or more, especially if the underlying uppermost Precambrian strata be included) now hard to decipher because involved in the overturned and mostly strongly thrust northwest limb of the Blue Ridge anticlinorium southeast of

¹ This statement is cast in terms of the traditional North American subdivisions of the Ordovician system into three series (Twenhofel, Chairman, 1954)—Lower, Middle, and Upper—according to which the Lower Ordovician equalled the Canadian (now Ibexian) series, roughly Tremadoc-Arenig, and the Middle Ordovician began with the Chazyan stage, roughly Llanvirn with or without Llandeilo. The disconformity mentioned also separates the two series over most of the central North American platform. Working in more continuous sections off the platform, Cooper (1956, p. 7–8) introduced a sub-Chazyan but still Middle Ordovician Whiterock stage, but later work showed that its upper part includes the Chazyan stage, and it has now been reclassified as the Whiterockian series (Ross and others, 1982, p. 10–12), upper Arenig to basal Caradoc. But in some areas strata belonging to the lower Whiterockian (and probably as young as lower Llanvirn) lie *below* the disconformity. Furthermore, it is quite possible that the disconformity (where present) is diachronous, becoming younger onto the platform, as suggested by Jacobi (1981) using plate-tectonic arguments.

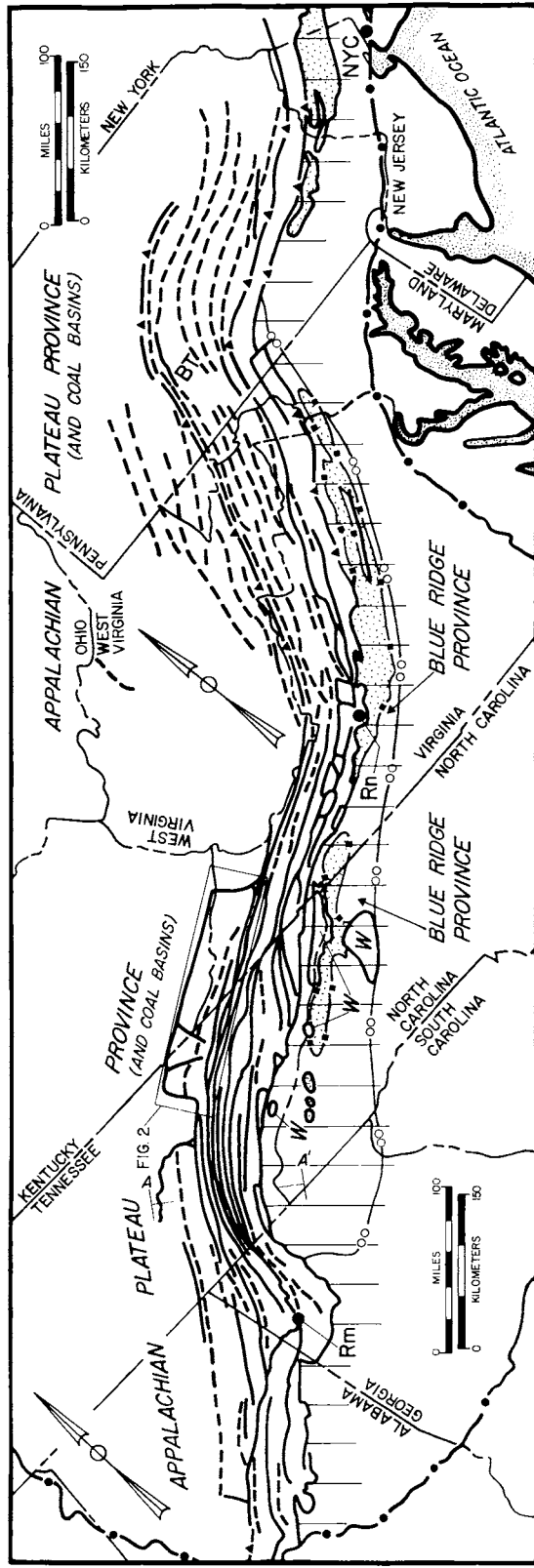


Fig. 1(A) Schematic tectonic map of central and southern Appalachians.

Heavy full line: major fault (selective; most of those shown are thrust faults dipping southeast)

Heavy dashed line: major anticline or anticlinorium in Valley and Ridge and Appalachian Plateau provinces

Heavy line with triangles: structural front—continuous belt of vertical or overturned dips, only locally broken by thrust faults (triangles point to downthrown side)

Thin vertical lines: provinces southeast of Valley and Ridge province—Blue Ridge province and northwest part of Piedmont province

Areas with short-dash pattern: main outcrop areas of Precambrian basement in Blue Ridge province

Line with squares: margin of Precambrian basement where not a major fault

Line with double circles: southeast boundary of Blue Ridge province

Heavy line with filled circles: overlap of Cretaceous and Cenozoic strata of Atlantic and Gulf Coastal Plains

Thin-line rectangle: outline of figure 2

Line with two dashes: state boundary

Letters: BT—Broad Top coal field; NYC—New York City; Rm—Rome, Georgia; Rn—Roanoke, Virginia; W—tectonic window (in Blue Ridge province)

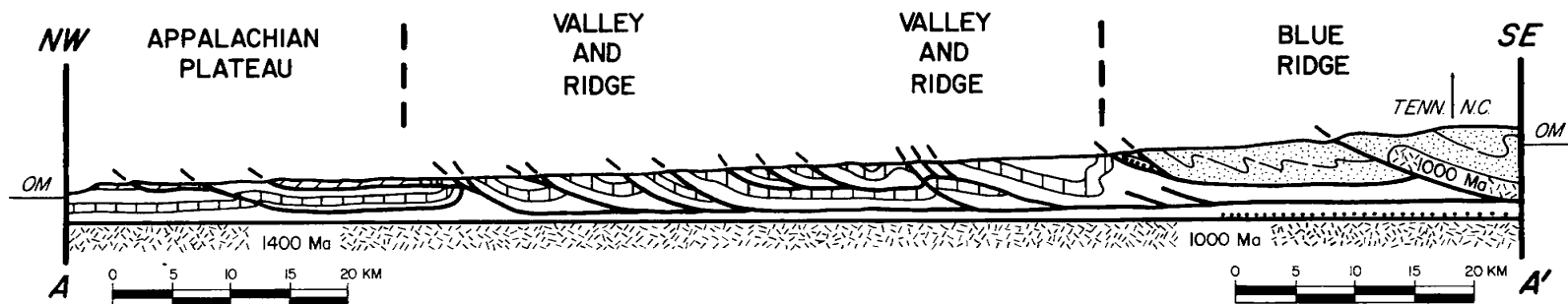


Fig. 1(B) Cross section A-A' (location shown on map); simplified from Roeder, Gilbert, and Witherspoon (1978), structure section 9.

Diagonal lines—Upper Carboniferous (Pennsylvanian)

Blank—Lower Carboniferous (Mississippian) to Middle Ordovician

Vertical lines—Lower Ordovician and Upper Cambrian (most competent unit in section)

Blank—Middle and Lower Cambrian (includes main décollement level), heavy dots indicate basal Cambrian clastic group

Dots—Upper Precambrian sedimentary rocks (metamorphic near southeast end of section)

Dashes—Precambrian ("Grenville") basement

the Valley and Ridge province. This thick wedge of sediments was first understood by James Hall in the 1850's (Hall, 1883); he also understood that it was genetically related to the folding and the existence of the mountains. Later Dana (1873, p. 430) gave it the name "geosynclinal" (nowadays it would be called miogeosynclinal or miogeoclinal).

The main deformation acted on this strongly asymmetrical geosynclinal prism, affecting all the strata above the décollement zone about equally; hence within the province, at least in the central Appalachians, it must be in the main very late Carboniferous or younger (but in the southern Appalachians the youngest strata still preserved are only middle Late Carboniferous—"Allegheny" in the eastern United States time scale). The deformation was certainly over well before the deposition of Upper Triassic strata in the Newark and related half-graben of eastern North America; it is generally assigned a Permian age. It used to be called the Appalachian orogeny or Revolution, but, as it is now known to be only the last of a series of orogenies that affected different parts of the Appalachians (but the others are mostly reflected in the Valley and Ridge only by the clastic wedges), we now generally call it by Woodward's name Alleghany orogeny (Woodward, 1957, p. 1440; 1958—note the *a* in the third syllable, in contrast to the *e* in the "stage" name mentioned above). It is becoming increasingly clear, however, that it too was polyphase (Geiser and Engelder, 1983), and more internally it may have begun during the late Early Carboniferous.

The folds and thrust faults of the Valley and Ridge province are remarkable for their length and parallelism. Their trends outline two major, smoothly curved salients, each about 550 km long (the Pennsylvania and Tennessee salients, corresponding to the central and southern Appalachians—see fig. 1), marked off by three more angular recesses (the New York, Roanoke, and Rome recesses), in which folds from the adjoining salients give the impression of crossing and interfering with each other. Southwest of the southwestern (Rome) recess, the trends are less curved, but they extend about 250 km toward an even more angular recess hidden under the overlapping Cretaceous sediments of the Coastal Plain. Some individual anticlines can be followed for a few hundred kilometers and at least three thrust faults for 600 km, though both folds and faults branch; some groups of en echelon folds and related thrust faults also form long conspicuous alignments. In certain parts of the belt the axial plunges of the anticlines and synclines are also remarkably parallel; in others, as among the en echelon fold groups, they are less so, but the overall regularity is striking. The vergence of both folds and thrust faults is very generally toward the craton—northwestward—with less than 5 percent of exceptions among the major structures, more among the minor. The belt is about 50 km wide in and near the recesses but widens to 75 km in the Tennessee salient and to 100 km in the Pennsylvania salient.

Thrust faults are much more important in the surface geology of the southern Appalachians than in that of the central, where large folds

dominate. The reason for this difference has been debated: perhaps it depends on differences in the stratigraphic column (the thickest clastic wedges are in the central salient), perhaps on differences in the total shortening (greater in the south). Subsurface evidence shows however that the difference is much less fundamental than we used to think: thrust faults are also important in the central Appalachians, but many of them are blind, never reaching the surface. A long argument as to whether or not décollement is important (thin-skinned versus thick-skinned tectonics) has been pretty well settled: it is important and indeed ubiquitous.

Perhaps the most significant development in Valley-and-Ridge geology in the last two decades concerns the shapes of the folds, especially those in the northwest part of the province (the Valley and Ridge par excellence), where body deformation is minimal and is confined to the less competent shale sequences. The folds are certainly parallel (thicknesses are reasonably constant, and there is little internal strain, at least in the competent strata), but they are not concentric, as we used to think, for they are not simply arcs of circles with centers near the underlying zone of décollement but are made mainly of relatively plane panels (Nickelsen, 1963, p. 19–20; Faill, 1969), though in the more competent layers the panels are joined by much shorter curved sections (more than one per anticline or syncline).

It is sometimes stated that the degree of deformation increases steadily from northwest to southeast across the Valley and Ridge province, from the gentle folds in the Appalachian Plateau province to the intense deformation, commonly associated with slaty cleavage and incipient metamorphism, at the margin of the Blue Ridge province. But in fact the deformation increases by “quantum jumps,” as is particularly clear in the classical section across central Pennsylvania. Adjacent to the relatively low folds of the Plateau on the northwest (wave length 10–20 km, amplitudes up to 800 m, dips rarely over 10°) is the highest (7 km) and steepest of the folds in the northwest part of the province, the Nittany arch, whose northwest flank is a wide belt of vertical or overturned strata (overturned locally to dip 30° southeast!)—a structural front—with associated complications (windows through an otherwise blind thrust fault—Moebs and Hoy, 1959). Southeast of the Nittany arch, the major folds are somewhat smaller—wave lengths 3 to 10 km, amplitudes up to 5 km—though still much larger than any on the Plateau.

Farther southeast, well beyond the mid-line of the province, is another structural front or wide belt of vertical to overturned strata, beyond which (in the physiographic “Great Valley”) the style of folding changes from the large “classical” folds to much tighter and more closely spaced folds, similar rather than parallel (even in the more competent layers) and associated with obvious body deformation—oölite deformation, cleavage, incipient metamorphism. The southeast margin of the province is a third such structural front, in which the

tegument—the sedimentary layers beneath the décollement—and the basement, both unaffected beneath the Valley and Ridge province itself, are brought up on the vertical to overturned northwest flank of the great South Mountain or Blue Ridge anticlinorium. Each of the structural fronts is parallel to the general trend of the folds in the Valley and Ridge province, but none is strictly continuous along strike from one end of the province to the other, for they show en echelon offsets or changes of character: thus southwestward the belts of vertical strata tend to be replaced, at the present land surface, by thrust faults of especially large displacement. Hence different cross sections show different features, but none shows a simple, gradual, southeastward increase in deformation.

On the Appalachian Plateau, a typical foreland basin lying on the cratonal margin immediately northwest of the Valley and Ridge province, the folds are low in central Pennsylvania, but southwestward certain of them become a good deal higher and begin to resemble the large folds in the Valley and Ridge province, even to being cut by northwest-verging thrust faults, though commonly they are separated from the edge of that province by flat-bottomed synclinal basins. (One such basin, the Broad Top coal basin, is preserved well within the Valley and Ridge province in southern Pennsylvania.) A remarkable example is the Pine Mountain fold and thrust fault (fig. 2), with a combined structural relief of about a kilometer, which extend together from a point on the Virginia-Kentucky border (Skaggs Gap) west-southwestward for 200 km across southeastern Kentucky into Tennessee, parallel to the trend of the nearest folds in the Valley and Ridge belt but separated from them by a flat-bottomed coal basin, the Middlesboro basin, from 13 to 40 km wide. The Pine Mountain thrust fault is cut off sharply at each end, beyond which there are only flat coal strata, but the Pine Mountain fold, much lower, extends a few kilometers east of the eastern end of the fault before dying out. At the southwest end in Tennessee, fault and fold turn abruptly through nearly a right angle and cut south-southeastward 15 km straight across the flat coal measures to the edge of the Valley and Ridge province, which they offset another 15 km and beyond which they turn back more gradually through south to southwest and become normal Valley-and-Ridge structures, extending on across the state. Where the fold and fault (Jacksboro fault) cross the strike, they clearly record a combination of thrust and left-handed strike-slip movement, the latter about 15 km.

At the northeast end of the Pine Mountain structure, the flat-lying strata of the Middlesboro basin seemed at first to be continuous around that end into the surrounding strata on the main Plateau, but Wentworth (1921) recognized that a line of badly disrupted rocks, trending southeast to the edge of the Valley and Ridge province, is a zone of right-handed strike-slip faulting (Russell Fork fault) with about 3 km of displacement. Not much later, windows were discovered (Butts, 1927) in the first (flat-topped) anticline (Powell Valley anticline) southeast of the

southwest part of the Pine Mountain fault and fold: their presence shows conclusively that the thrust fault, though it dips about 30° southeast where it crops out along the Pine Mountain fold, lies flat within a Devonian black shale unit (Chattanooga shale) under the entire Middleboro basin and only starts to dip down again under the anticline. It was here that Rich (1934) first described and explained ramp-and-flat thrusting, showing that under these circumstances the folding results from the thrusting, not the other way round—that is, with the thrusts developing secondarily on the oversteepened or overturned limbs of growing folds as had seemed intuitively obvious before.

A very similar northwest-verging fold, the Sequatchie anticline in southern Tennessee and northern Alabama, also cut by a thrust fault (though the fault is not continuous along the entire fold), is over 300 km long though the displacement is less; again it is separated from the Valley and Ridge province by a belt of flat-lying coal strata 12 to 25 km wide. The strike-slip fault system at the north end is clear, but that at the southwest end is not. Still a third fold and fault pair on the Plateau in Alabama, the Murphrees Valley anticline, is, though shorter, like the others in everything except its reversed vergence: indeed it was for this fault and fold that Smith (1893) coined the term “underthrust.”

Rich’s ramp-and-flat “model” also provides the most satisfactory explanation for the structure of the Valley and Ridge province, at least for its northwestern part as far southeast as the structural front beyond the mid-line of the province, as suggested by Rich himself followed by several others. It also explains very well the flat-panel shape of the folds (Suppe, 1983). Starting from this model, more and more sophisticated balanced cross sections have been prepared, at least for parts of the belt, notably those of Roeder, Gilbert, and Witherspoon (1978) and of Woodward (1985; see also Boyer and Elliott, 1982, fig. 29—their section concerns the adjacent Blue Ridge more than the Valley and Ridge but is an excellent illustration of the method and its development for duplex structure). All these reconstructions demand that the pile of strata above the main basal décollement be greatly shortened across the trend of the belt, whereas the strata below (the tegument) and the basement are not shortened here at all but must be shortened somewhere else, farther southeast (unless the shortening of the cover be ascribed to gravity sliding from a zone of tectonic denudation, for which no evidence exists). As the strata cratonward of the Valley and Ridge belt (or of the outlying folds on the Plateau) have clearly not moved relative to their basement, the relative horizontal displacement of the strata within the belt must increase in the opposite direction, and its maximum at the opposite margin must be much larger (10 times?) than the average thickness of the strata (5 km or so) being displaced. The requirement that a pile of strata be pushed laterally much farther than its own thickness has always been the most difficult point in the décollement hypothesis to accept, seeming at first to be counter-intuitive (this difficulty has led many to prefer hypotheses that explain the shortening not by compression but by simple gravity).

The strong (but not invariant) asymmetry in vergence of the folds and thrust faults has commonly been interpreted as showing that “the push came from the southeast,”

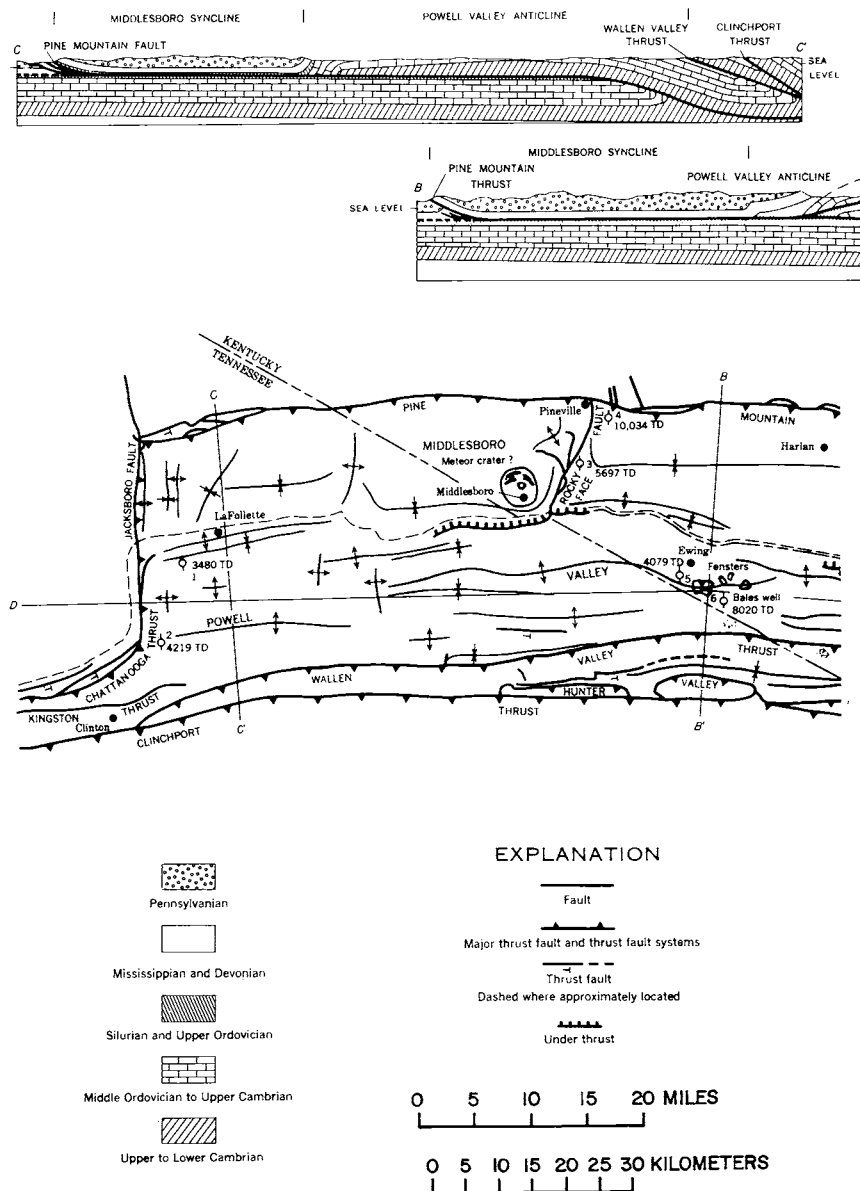


Fig. 2.

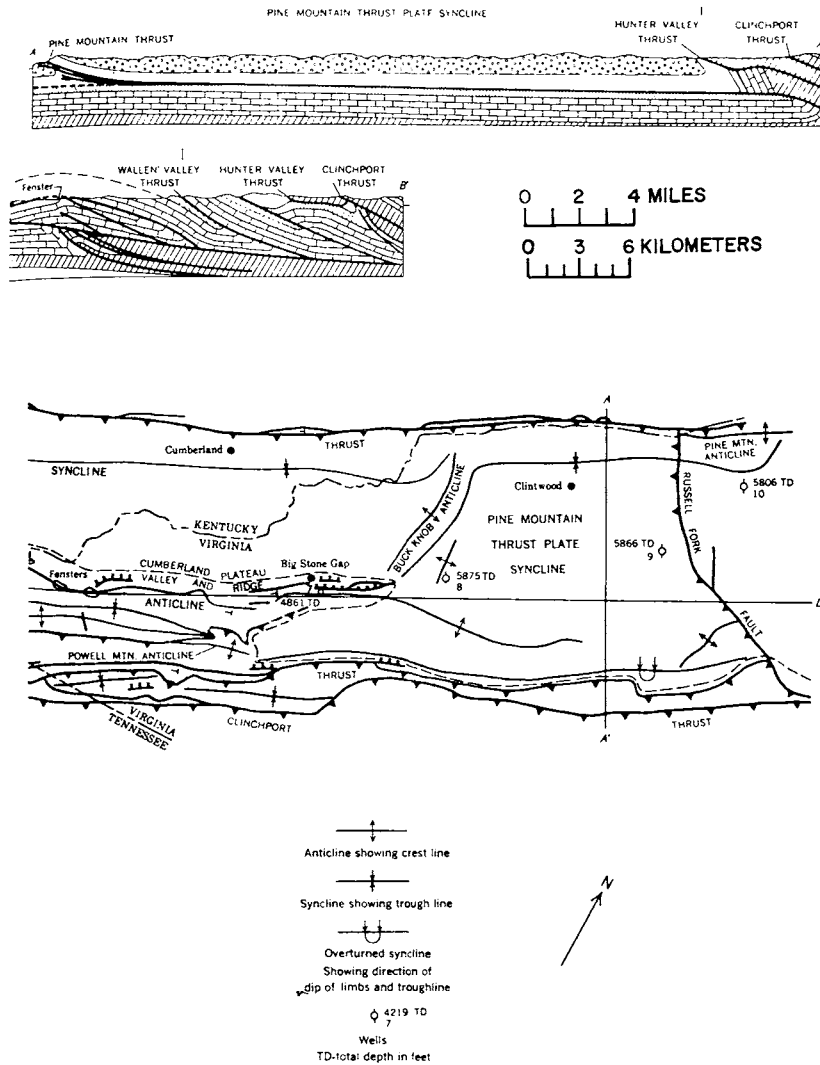


Fig. 2. Structure map of the Pine Mountain thrust block (from Harris, 1970, fig. 1, which gives original sources of data). Dashed line is boundary between Valley and Ridge province (southeast) and Appalachian Plateau province (northwest). Figure reproduced by permission of the copyright holders: Wiley-Interscience, New York.

which would also seem to explain the increase in displacement and deformation in that direction. But Newton's third law, or simply an easy thought experiment, disproves this idea; if a pressure box were mounted on wheels, it would make no difference whatever whether the piston were pushed to move into the stationary box or the box were pushed to move against the stationary piston (assuming that accelerations were small enough that inertial forces could be neglected). The asymmetry of the folds and thrust faults of the Valley and Ridge province does not depend on the "direction of push" nor on the "absolute motion," but on the asymmetry of the original boundary conditions; that is, of the geosynclinal wedge or prism.

The "quantum jumps" in the degree of deformation across the structural fronts within and bounding the province presumably record mechanical thresholds between different types of deformational behavior; the apparent relation of at least the Blue Ridge front to the onset of metamorphism in the Paleozoic strata suggests that the thresholds are thermally controlled, so that the fronts are in effect isotherms. The relatively mild deformation in the Appalachian Plateau province is clearly related to décollement levels higher than the main level in the Valley and Ridge itself; perhaps movement on the main level, probably a major evaporitic zone, was activated thermally so that beyond some isotherm it could no longer function and shortening could be accommodated only on higher zones (some are certainly shale layers, but one, near the top of the Silurian, is a fairly thick zone of halite). The front near the mid-line of the province separates parallel folds of large wave-length in the competent layers of massive carbonate and sandstone (the incompetent layers between show folds of all scales and much wedging) from tight similar folds accompanied by body deformation in the same competent rocks; the change may have become possible because the temperature was higher. Beyond that front the Rich model may not be applicable. Finally, the structural front that separates the Valley and Ridge province from the Blue Ridge anticlinorium records the isotherm where the basement ceased to be so rigid as to resist deformation but instead was softened enough to take part in it, so that the whole style of deformation changed again. But the structure of the Blue Ridge and comparable anticlinoria or "external massifs" is the subject of another article (Rodgers, in preparation).

EXTERNAL FOLD-AND-THRUST BELT OF THE WESTERN ALPS: THE SUBALPINE CHAINS OF FRANCE AND THE HELVETIC ZONE OF SWITZERLAND

A belt of relatively large-scale Cenozoic folding and thrust faulting a few tens of kilometers wide swings right around the great arc of the Western Alps from the Mediterranean near Nice to the Rhein River in eastern Switzerland (fig. 3, shaded belt). From Nice to Genève, this belt forms the Subalpine chains of France; its folds (wave-length up to 5 km) verge with fair but not complete consistency away from the core of the Alps in all cross sections. In the southern Subalpine chains thrust faults are as important as folds, but in the northern chains folds predominate. As the belt passes into Switzerland, one enters the classic ground of the

Helvetic nappes², also outward-vergent, and thrust faults are again important. Individual structures here are somewhat oblique to the belt as a whole, so that the northern Subalpine folds of France are not in line with the Helvetic nappes of Switzerland. Moreover, tectonic units from deeper in the Alpine chain have been superposed on the fold belt here, forming the Préalpine and other klippen (including the type klippen in central Switzerland), but I do not discuss those higher units in this article.

This belt in France and Switzerland is described briefly in articles introductory to the relevant guidebooks of the 26th International Geological Congress (Debelmas and Kerckhove, 1980, esp. p. 26–32; Trümpy, 1980, esp. p. 52–61; these introductions were reprinted in the collection: *Geology of the European Countries*, 1980, Dunod, Paris). Unfortunately, these articles carry no references.

The cover sequence involved in these folds and thrusts ranges from Triassic to mid-Cenozoic in age and from 2 to 5 km in thickness. Its stratigraphy varies considerably from area to area: moreover stratigraphic and structural trends are not parallel to each other but distinctly oblique. Normal faults formed graben or half-graben occasionally during Triassic time and especially in several episodes in the Early and Middle Jurassic, following which oceanic troughs formed in more internal parts of the Alpine region, and the belt here discussed became a trailing margin (Günzler-Seiffert, 1941; Lemoine and Trümpy, 1987). Hence Triassic and Jurassic strata range from thick to thin or even absent, and the structural style varies considerably along the belt.

The Triassic generally begins with sandstone, which remains structurally with the basement as a tegument; above are weak continen-

² I have tried to use the word *nappe* (German: *Decke*) in its original meaning in tectonics for a fairly large body of rock that has been transported laterally over another body of rock for a distance of at least 5 km (this limit was proposed by Cornelius, 1940, p. 274). Nappes can be of two kinds, as suggested by Termier (1906); those of the first kind are recumbent folds, those of the second are what North American geologists would call large thrust (or “overthrust”) sheets (the words *nappe* and *Decke* mean cover or blanket, not too far from sheet or, for that matter, bedding). The word “over,” as used in the first sentence in this footnote, should imply only the *relative* motion of the two bodies of rock, but much of the older literature was so written as to imply *absolute* lateral motion of the upper body over the lower.

The word Helvetic seems to have entered the Alpine geologic literature as the name of a stratigraphic facies whose Cretaceous and lower Tertiary strata differ considerably from those in other Alpine areas, notably the Préalpes. A few years later, the strata of that facies were shown to form a well defined group of nappes, likewise distinct from those of the Préalpes (Lugeon, 1901, p. 812, 817—“nappes à faciès Helvétique”), and soon the term was transferred to the structural zone formed by those nappes. More recently, the lowest of these nappes, Morcles and Doldenhorn, have commonly been excluded from the Helvetic group (by Trümpy, 1963, p. 422–423, and Lemoine and Trümpy, 1987, fig. 2, for example, but not by Ramsay, 1981, or Burkhard, 1988) and classed as parautochthonous, on the grounds that they root considerably more externally than the higher nappes—in the frontal parts of the Aar and Mont Blanc massifs instead of behind them—and that they come from a different sedimentary trough, or arm of a trough, although the stratigraphic sequences are comparably “Helvetic.” Personally I prefer to follow Lugeon and retain the Morcles and Doldenhorn folds in the Helvetic zone.

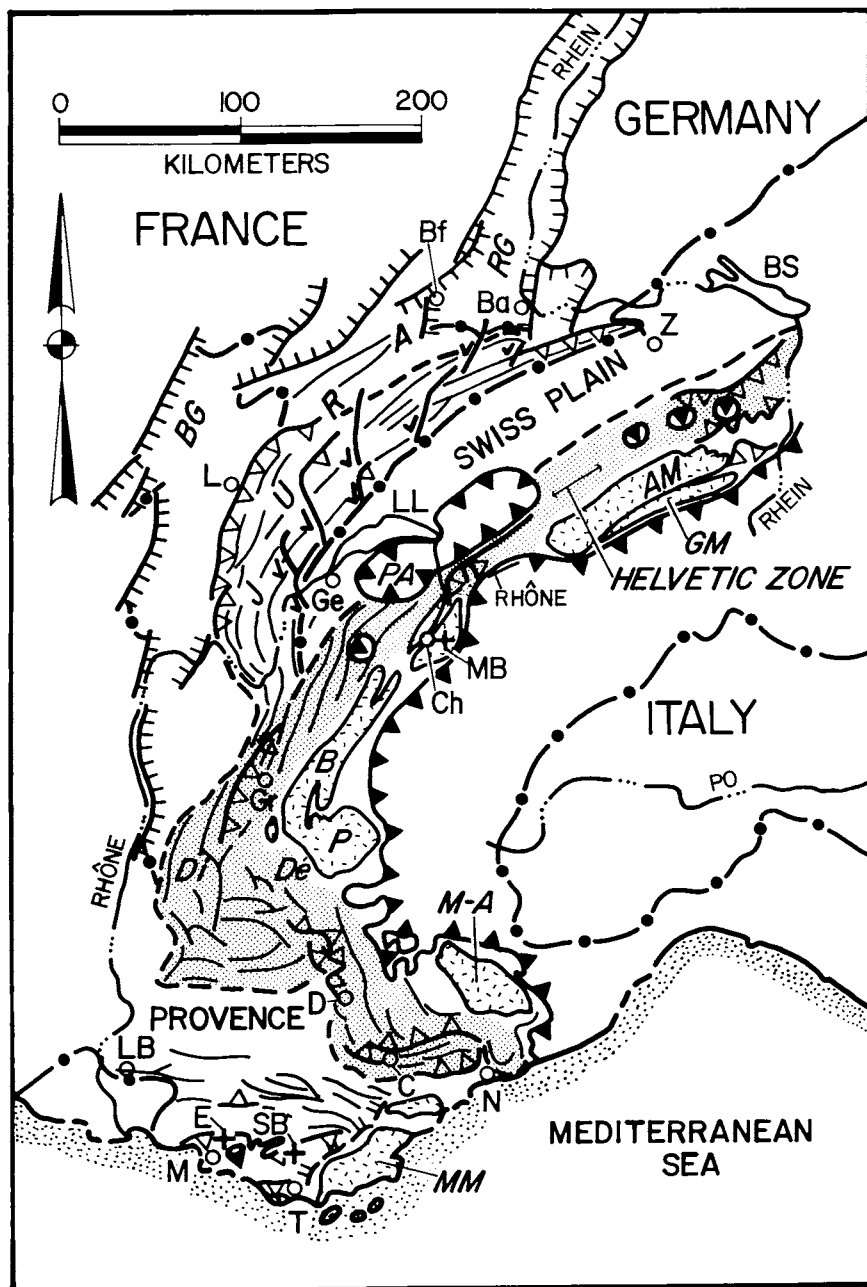


Figure 3

EXPLANATION

	Overlap of larger areas of Cenozoic strata
	Normal fault, ticks on downthrown side
	Front of strong folding in Alps and Jura
	Anticlinal axis
	Thrust fault (<i>very selective</i> ; for Helvetic zone shown only at each end); triangles on upthrown side
	Strike-slip fault (with associated thrust fault or anticline)
	Frontal thrust fault of internal, Pennine, zone, and boundaries of klippen
	Basal decollement in eastern Provence
	Subalpine-Helvetic zone of Western Alps (Subalpine chains in France, Helvetic zone in Switzerland)
	Variscan "basement" in external massifs of Alps and in Massif des Maures (not shown outside area of Tertiary deformation)
	Mountain
	River
	Locality

Fig. 3. Schematic tectonic map of the outer zones of the western Alps and of related ranges, from the Mediterranean Sea to the Rhein River.

Letters: AM—Aar massif; B—Belledonne massif; Ba—Basel; Bf—Belfort; BG—Bresse graben; BS—Bodensee (Lake of Constance); C—Castellane; Ch—Chamonix; D—Digne; Dé—Dévoluy; Di—Diois; E—L'Etoile; Ge—Genève; GM—Gotthard massif; Gr—Grenoble; L—Lons-le-Saunier; LB—Les Baux; LL—Lac Leman (Lake of Geneva); M—Marseille; M-A—Mercantour-Argentera massif; MB—Mont Blanc; MM—Massif des Maures; N—Nice; P—Pelvoux; PA—Préalpes; RG—Rhein graben; SB—La Sainte-Baume; T—Toulon; Z—Zürich.

tal red pelitic rocks, including evaporites (both anhydrite-gypsum and halite) that in certain regions occur at two levels separated by thin but more competent marine limestone or dolostone. Where the Jurassic strata are thick, they are chiefly deep-water calcareous mudstone or shale (marl, marne, or Mergel in the European terminology), with some thin limestone units; where they are thin, they are more varied and include conglomerate and breccia derived from the adjacent horsts. By late Jurassic time, however, conditions were more uniform, and a sheet of limestone spread over the whole region, shallow-water on the periphery but pelagic to deep-water over most of the belt. The pelagic facies is assigned to a special "stage", the Tithonian, which is only roughly equivalent to the standard Portlandian stage.

Lower Cretaceous sediments are more differentiated again. Deep-water deposits, mainly marl, continued in the Vocontian trough, which cut diagonally across what are now the south-central Subalpine chains in France but connected to more internal troughs in Switzerland. Elsewhere, shallow-water shale and limestone predominated, culminating in a great sheet of massive limestone, partly a rudistid reef, a facies of Barremian-Aptian age to which again a "stage" name, Urgonian, was assigned. The presence of this sheet of limestone was one of the distinguishing characters of the "Helvetic facies." The two limestone sheets mentioned, Tithonian and Urgonian, are both resistant to erosion, and they dominate the landscape through much of the belt, except in the former Vocontian trough where only the Tithonian is present.

Upper Cretaceous strata are irregularly distributed and range from "pure" limestone to shaly and sandy limestone; some of the region was emergent, and east-west folds formed in part of the Vocontian trough. Where lower Tertiary strata are present, they are disconformable on the Cretaceous (the Upper Cretaceous may be partly or entirely missing) and form a "triad" composed of a thin nummulitic limestone layer below, a zone of "marl", and a body of flysch or flyschoid sandstone above. This triad is another mark of the Helvetic facies. The whole cover sequence may be characterized as "miogeosynclinal", but its thickness is quite irregular in detail, and it is less obviously wedge-shaped than in the Appalachians.

The folds and thrust faults in the southern Subalpine chains form two smoothly curving salients or arcs, known as the arcs of Nice and Castellane, and a more complex area of polyphase deformation north of Digne. Triassic evaporites provide the décollement (tegument is locally involved near Digne), and there are higher décollement levels and zones of disharmonic folding in the Jurassic marls and even in the Lower Cretaceous. Deformed Pleistocene deposits and recent earthquakes suggest that the arc of Nice is still actively deforming.

West and north of the complex area north of Digne, Triassic strata are not exposed, perhaps because they are thin or absent (as they are along parts of the margins of the adjacent external massifs from the Pelvoux northward), and décollement was in the Jurassic, especially in

the lower Upper Jurassic (Oxfordian) black clays or marls. In the area of the Vocontian trough, where the stiff sheet of Urgonian limestone is absent, the folds have shorter wave-lengths and curved axes and show less continuity, inconsistent vergence, and polyphase interference patterns, notably in the Diois. In the nearby Dévoluy, an angular unconformity within the Upper Cretaceous is particularly clear; above it is a unit of massive (Senonian) limestone.

The Urgonian limestone sheet reappears just north of the Diois, and the northern part of the Subalpine chains forms a gentle salient or arc past Grenoble to southeast of Genève. A typical, classical cross section of this portion of the belt is provided by the cluse of the Isère River west of Grenoble (fig. 4, from Gignoux and Moret, 1952, fig. 48, p. 221). Here the lowest unit visible, except at the extreme eastern end, is the Upper Jurassic black marl, and the Tertiary is represented not by the "triad" but by Miocene Molasse (coarse poorly consolidated deltaic or shallow-marine clastics), which is pinched in tight synclines under thrust faults, especially at the west end of the section, and is thus clearly involved in the deformation. The folds and thrust faults verge west.

The folds in the easternmost part of the Isère section are not merely asymmetric or partly overturned like the others but recumbent; the axial planes have been tipped beyond the horizontal to dip west, so that some anticlinal noses are now synformal. To the north, the individual folds diagonal westward across the belt of folds (the westernmost fold or thrust slice here becomes separated from the rest and continues into the southern end of the Jura; see p. 340), and hence more and more of the belt comes to be dominated by recumbent folds, some of which stand on their noses or heads as it were (*plis plongeants*). Along the Arve cross section, followed by the much-travelled road from Genève to Chamonix and Mont Blanc, only the western folds (at Cluses and to the west) are still in "normal" position, the rest, magnificently displayed on the northeast wall of the valley, are recumbent.

Between the Arve and the Rhône (in Switzerland), the larger and higher recumbent folds become detached from those below and form the first and lowest of the nappes originally described by Lugeon (1901), the nappe de Morcles. Minor recumbent folds (digitations) on the lower, overturned limb of the Morcles nappe are beautifully displayed in the Dent de Morcles east of the Rhône and the Dent du Midi west of it.

East of the Rhône, a strong axial plunge to the east-northeast carries the Morcles nappe beneath two higher nappes (apparently of the second kind, with distinct thrust surfaces cutting out much or all of their overturned limbs), the nappe des Diablerets and the nappe du Wildhorn; the latter then becomes the Wildhorndecke as it crosses the linguistic frontier between French and (Swiss) German. The Diablerets nappe can be considered only an especially prominent lower digitation of the Wildhorn nappe. In central Switzerland, the exposed Helvetic zone is composed mainly of the Wildhorndecke. The lower nappes or Decken appear only to the southeast close to and involved with the northern

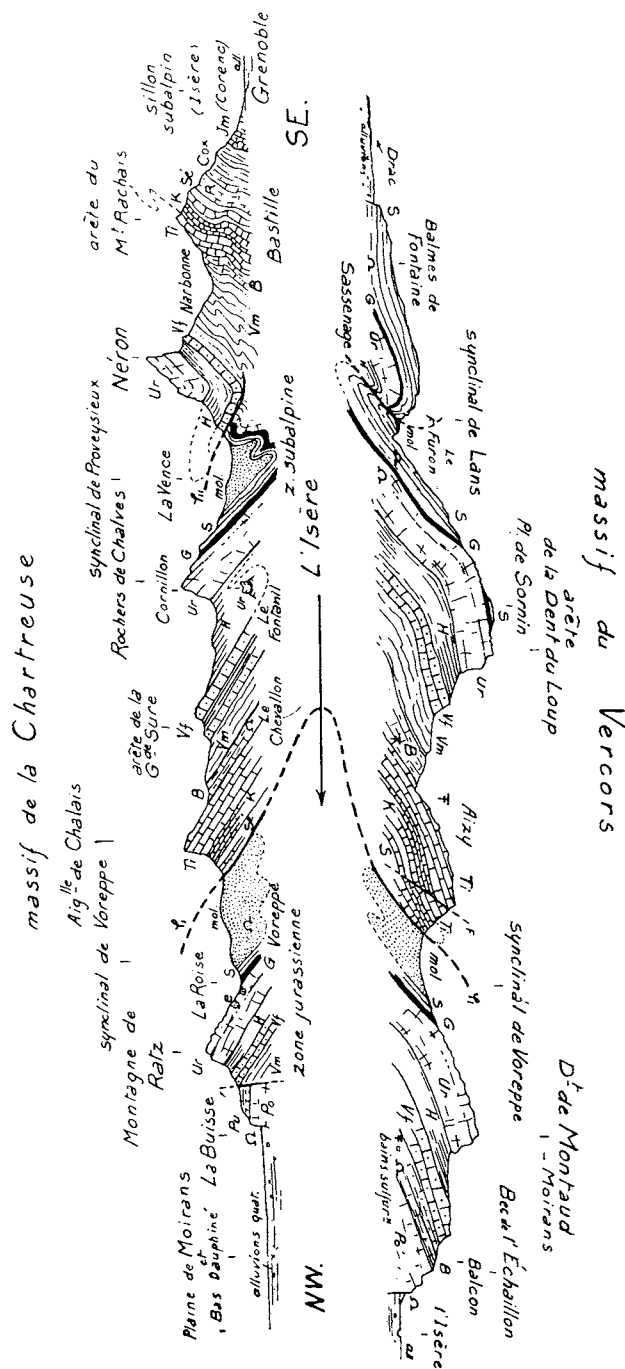


Fig. 4. Geologic structure of the cluse of the Isère, from Gignoux and Moret, 1952, fig. 48, p. 221. The upper section is on the north side of the cluse, the lower on the south side. Figure reproduced by kind permission of the copyright holders: Masson, Paris.

Letters, in stratigraphic order (top to bottom): mol—Miocene Molasse, S—Senonian, G—Albian, Ur—Urgonian, H—Hauterivian, Vf—Valanginian limestone, Vm—Valanginian marl, B—Berriasian, Pu—Purbeekian (with "bastard marble"), Po—Portlandian (coralline limestone), Ti—Tithonian, K—Kimmeridgian, Sé—Séquanian, Cox—Callovo-Oxfordian (marl), Jm—Middle Jurassic. Φ and F—thrust faults, Ω —quarry.

margin of the Aar massif (which here plays the same rôle as the Blue Ridge in the central and southern Appalachians), whereas the Wildhorndecke has its roots behind that massif. The Wildhorndecke itself becomes more and more complex eastward, each of its parts receiving separate names, and in eastern Switzerland it splits up into a pile of thrust sheets or Decken, which in general consist of only one part of the stratigraphic column (here including thick Permian conglomerate beneath the Triassic, which is thin and not always evaporitic in this region). These sheets can be traced around the east end of the Aar massif to roots behind it.

It is here in the canton of Glarus, north and northeast of the east end of the Aar massif, that the idea of large-scale thrusting was first suggested by Escher von der Linth (1841), just before the publication of the Rogers brothers on the Appalachians, and that Marcel Bertrand (1884) showed (without visiting the area!) that the famous "double fold of Glarus," supposed to be two recumbent folds that had moved 10 or 15 km toward each other, was in fact a single great folded nappe (of the second kind) with not less than 45 km of displacement relatively toward the north-northwest. The story of the development of the nappe concept in the Alps is beautifully told by E. B. Bailey in his book *Tectonic Essays mainly Alpine* (Bailey, 1935).

There is no doubt that the Subalpine-Helvetic zone from Nice to the Rhein is governed by décollement, by thin-skinned tectonics. The main décollement level is in the Triassic evaporites or, as in the Isère section, in the Jurassic marls, but in a few places slices of Lower Triassic or Permian tegument or of Variscan basement are brought up, showing that locally the basal thrust clipped off pieces of the rock mass beneath. The most extraordinary feature of the belt taken as a whole, and indeed of the whole West Alpine chain, is the great curve it makes; in eastern Switzerland the folds and thrust sheets verge north-northwest, in the Isère cross section they verge west, and in the arc of Nice they verge south. Thus the geometric constraints on any kinematical explanation are extreme.

The Rich ramp-and-flat model of faulting and folding, mentioned above (p. 329), appears to be generally applicable to the simpler parts of the Subalpine chains, but cleavage, body deformation, and ductile strain are present in the less competent rocks in those chains and predominate in the Helvetic belt. Perhaps the two regions can be compared to the two parts of the Valley-and-Ridge province in central Pennsylvania (p. 327–328). Ramsay, Casey, and Kligfield (1983) have shown how the presence of ductile shear zones modifies ramp-and-flat geometry to produce the complex folds in the Helvetic nappes.

Stratigraphic analysis of the thrust sheets of eastern Switzerland (for example, de Sitter, 1939, fig. 12) has shown that even some of the details of the distribution of thrust faults can be accounted for by the lateral extent of weak zones that serve as décollement levels. In particular, Günzler-Seiffert (1952) and Trümpy (1969, Tafel II) have shown that abrupt thickness changes in Mesozoic strata across synsedimentary, commonly listric, normal faults limiting half-graben (displacement is mostly down to the south-southeast,

basinward) have guided the later thrusting; in many cases the older faults have moved again, cancelling or reversing their original displacement. Similar analysis in other fold-and-thrust belts has proved fruitful, as in the Canadian Rockies and the Appalachians (explanation of the limits of the Pine Mountain thrust sheet in terms of the thickness and siltstone content of the Chattanooga shale, the décollement level).

THE JURA

The folded belt of the Jura stands somewhat apart from the Alps, although a part of the Alpine mountain system (see fig. 3). Its main body extends for nearly 300 km in a broad arc through northwest Switzerland and east-central France, from a narrow and abrupt east end 14 km north-northwest of Zürich, where the trend is south of west, to a rather sharp inflection in a wider part of the belt on the order of 50 km southwest of Genève, where the trend turns from west-of-south to southeast. Through this distance, the Jura is separated from the Subalpine-Helvetic zone by the Swiss Plain, a belt up to 50 km wide underlain by mostly flat-lying Oligocene to Miocene Molasse, an alternation of shallow-marine and fresh-water sandstone and conglomerate clearly derived from the rising Alps (although most of the fragments are from higher nappes than the Helvetic, which formed rather late in the Alpine history). Thus the Jura lies on the wrong, far side of a débris-filled foreland basin.

Beyond the inflection, the Jura rapidly narrows and approaches the French Subalpine chains, cutting across the southwestern termination of the Swiss Plain in which, south and southwest of Genève, several anticlines appear, rather of Jura than of Subalpine type. The Jura folds merge southward with the front of the Subalpine chains and turn back to follow their south-southwest trend. The thin strip of Miocene Molasse near the west end of the Isère section (fig. 4) is the last surface remnant of the Molasse of the Swiss Plain, from whose main body it extends southward for about 80 km as a narrow band never more than a kilometer wide and mostly less.

The southeastern anticline or group of anticlines in each cross section of the Jura is the highest, both topographically ("la haute chaîne") and structurally. At its highest point, Crêt de la Neige west of Genève, it reaches 1718 m; northeastward into Switzerland it decreases in height, only gradually at first, then more rapidly.

The stratigraphic sequence in the Jura differs from that in the Helvetic zone in containing more shallow-water limestone and less marl in the Jurassic and in being thinner (1–2 km; thicker to the southwest), not so much geosynclinal as platformal or cratonic (like the section under the Appalachian Plateau and farther west as compared to that in the Valley and Ridge province). The evaporitic Triassic is universal; to the west the main décollement level is in the Upper Triassic (Keuper), but to the east it steps down below the Middle Triassic carbonate (Muschelkalk). Upper Cretaceous is present only locally. The Tertiary, disconformable or even locally unconformable, is especially thin and in many areas discontinuous, in strong contrast to the thick Molasse (2–5 km thick—a wedge thickening toward the Alps) under the Swiss Plain

and also to the thick Tertiary fillings (more than 3 km) of the Rhein and Bresse (or Saône) graben adjacent on the other side. Clearly the folded strata now visible at the present surface in the Jura were never buried more than a kilometer or two.

Much of the Jura is characterized by its generally parallel anticlines, spaced $1\frac{1}{2}$ to 10 km apart and up to 40 km long; of the synclines between, some are also long and narrow, but others are rhomb-shaped or rectangular and have flat floors. Thrust faults accompany many of the anticlines, especially toward the two ends. In contrast to the Appalachians and the Subalpine-Helvetic zone, the asymmetry of the folds is strikingly inconsistent, although the thrust faults, at least the larger ones, tend to verge northwest. Many of the anticlines are box folds, or double box folds, or box folds with a "dimple" on top, and they commonly contain thrust faults in their cores; even more obviously than in the Appalachians, they are made of plane panels connected by tight arcs or separated by the thrust faults. Also prominent in the Jura are transverse strike-slip faults, or zones of strike-slip faults, which curiously show a more consistent (though by no means perfect) asymmetry than the folds; the majority, and the longest ones (cutting across most of the fold belt), are left-handed and strike to the right of the perpendicular to the fold trends. The folds show that the pile of strata has been shortened northwest-southeast, the strike-slip faults that it has been extended northeast-southwest. The interplay of the strike-slip faults with the folds (and thrust faults) is somewhat like that in the Pine Mountain structure in the Appalachians (p. 328) but much more complicated (Laubscher, 1965). The dimensions of the Jura are about 50 percent greater than those of the Pine Mountain structure.

The characteristic Jura folding dominates the whole southwestern third of the chain, which reaches a width of 60 km northwest of Genève, and the whole northeastern third, which narrows from 30 km southwest of Basel to 0 km north of Zürich, but only the southeastern part of the middle third, a band about 30 km wide next to the Swiss Plain; these regions are called the folded Jura. The northwestern part of the middle third, called the plateau Jura, is dominated instead by flat-bottomed basins or subhorizontal slabs, topographically plateaux, mostly rather rhombic, which are separated by narrow anastomosing and broken folds or zones of tight folds, most with complicated fault patterns. Moreover, not all these folds are anticlinal; in some of them strata are preserved at levels *below* their positions in the adjacent slabs. Again diagonal transverse faults (and fold trends) are common, contributing to the rhombic pattern, and those in the left-handed strike-slip position greatly predominate.

A third Jura, the tabular Jura, lies north of the east part of the folded Jura and extends northeast into Germany—for example the Schwäbischer Jura—but it lies entirely outside the Alpine orogenic belt, mostly lacks folds and thrust faults, and does not concern us here.

It was in the Jura that August Buxtorf (1907a, b) first demonstrated the tectonics of *décollement* (German: *Abscherung*). By care-

fully scaled cross sections of the box folds, using excellent subsurface data from tunnels as well as surface mapping, he showed that the folds must end downward at the level of the Triassic evaporites and that the basement (and tegument) below cannot be directly involved. Debate then centered on whether the Jura was shortened by sliding laterally over entirely unshortened basement all the way from the Alps (the Swiss Plain also sliding laterally but with almost no deformation), or whether shortening deformation of some other kind in the basement under the Jura provoked the thin-skinned Jura folding. Aubert (1949) argued that thrust faults in the basement beneath shortened it as much as the cover, and Wegmann (1961, 1963) and Pavoni (1961) that the basement was simultaneously shortened northwest-southeast and extended northeast-southwest by strike-slip movement along a network of diagonal (“conjugate”) strike-slip faults. Indeed, it had long been known that faults belonging to the Rhein graben system and to the system connecting the Rhein to the Bresse graben can be followed into the Jura, where some of them project into the strike-slip zones that cross the folded Jura and others into the zones of tight folds in the plateau Jura.

But drilling showed (Michel and others, 1953; Lefavrais and others, 1957) that, at the very west edge of the Jura near Lons-le-Saunier, the Mesozoic cover has slid at least 7 km westward over the already dead eastern border fault of the Bresse graben and over Middle Miocene strata that overlap that fault. Adding to this the shortening recorded by the several Jura folds farther east and by the large Risoux thrust fault, the total shortening of the cover in this cross section is not less than 20 km, probably as much as 30 km, considerably more than could be produced by blind thrust faults or strike-slip faults within the basement. Moreover the timing is wrong; as shown by the drilling near Lons-le-Saunier and by mapping around the south end of the Rhein graben, the main movements on the graben faults preceded by a considerable period the décollement folding of the Jura. Where those faults enter the Jura, especially along what are now the transverse strike-slip zones, their movement at the time the graben were forming was mainly normal dip-slip, as shown by their effect on the present distribution of Cretaceous strata and on the character of the overlying Tertiary (Aubert, 1975, especially fig. 23). Because of this earlier faulting, therefore, the décollement did not take place over a plane surface, as generally shown in textbook drawings of such structures, but over a surface already cut up into steps and risers by faults with various strikes (see fig. 1 in Laubscher, 1961, p. 228). The strong strike-slip faulting and even more the curious down-faulted strips in the plateau Jura find a natural explanation if the Mesozoic cover in slipping laterally found itself forced to accommodate to sizeable breaks in the underlying surface. Thus faulting in the basement did not *cause* the Jura folding but strongly influenced it, posthumously as it were, and determined some of its special characteristics (Laubscher, 1965, 1987).

If the basement beneath the Jura shows much less shortening than the cover, then the cover shortening must reflect basement shortening

somewhere else, and the Alps are the only possible place. But if so, not only the Jura but the entire Swiss Plain must have slid laterally in relation to its basement, and this idea seemed at first to raise insurmountable mechanical difficulties, for the Molasse under the Swiss Plain is affected by very few folds, and those mostly gentle except very close to the Jura and in the highly deformed "Subalpine Molasse," a narrow zone along the front of the Alps directly under the frontal thrust faults of the Helvetic or higher nappes. Goguel (1943, chap. 23) showed however that the energy that would have been necessary to fold the thick sandstone-conglomerate sequence of the Molasse is of at least the same order of magnitude as that necessary to make it slip, undeformed, laterally northwest *and* to crumple the Jura, and Laubscher (1961, p. 240–251) adduced fluid-pressure arguments to the same effect; the mechanical difficulties therefore fall away.

Although not perfectly dated, the Jura folding is very late, near the end of Miocene time, and indeed much of the present topography in the folded Jura reflects the original anticlines. On the other hand, the main faulting in the Rhein and Bresse graben apparently ended during the early Miocene (Laubscher, 1987), though some parts of the Rhein graben farther north are still active.

THE CANADIAN ROCKY MOUNTAINS (AND THEIR EXTENSION INTO THE UNITED STATES)

The fold-and-thrust belt of which the Canadian Rockies are a part extends for more than 2000 km along the west margin of the North American craton in Canada and (beyond complicated recesses) continues on for 1500 km in each direction, westward into the Brooks Range of northern Alaska and southward into the Western Rockies, *et cetera*, of the lower United States. Of the three major arcuate salients in Canada, separated by more angular recesses near the Liard and Peace Rivers, I shall discuss only the south half of the southernmost, the 450 km from the Athabasca River near Jasper, southwest of Edmonton, south to the recess 50 km north of the international border (fig. 5); this part is by far the best known because great interest in petroleum exploration has led to detailed mapping and stratigraphic study, much drilling, and thousands of kilometers of seismic reflection profiles. Good summaries abound; let me cite Price and Mountjoy (1970), Bally, Gordy, and Stewart (1966), and the early but remarkably prescient analysis by Link (1949). Because the belt here is so well understood in three dimensions, it has served as the basis for significant theoretical analyses of the kinematics and mechanics of fold-and-thrust belts generally; I need only cite the (somewhat divergent) analyses of Elliott (1976) and Chapple (1978). More than 100 km wide, this part of the belt straddles the frontier between the provinces of Alberta and British Columbia and forms the continental divide between drainage via the Saskatchewan and Athabasca Rivers toward the Arctic Ocean and that via the Columbia River toward the Pacific Ocean.

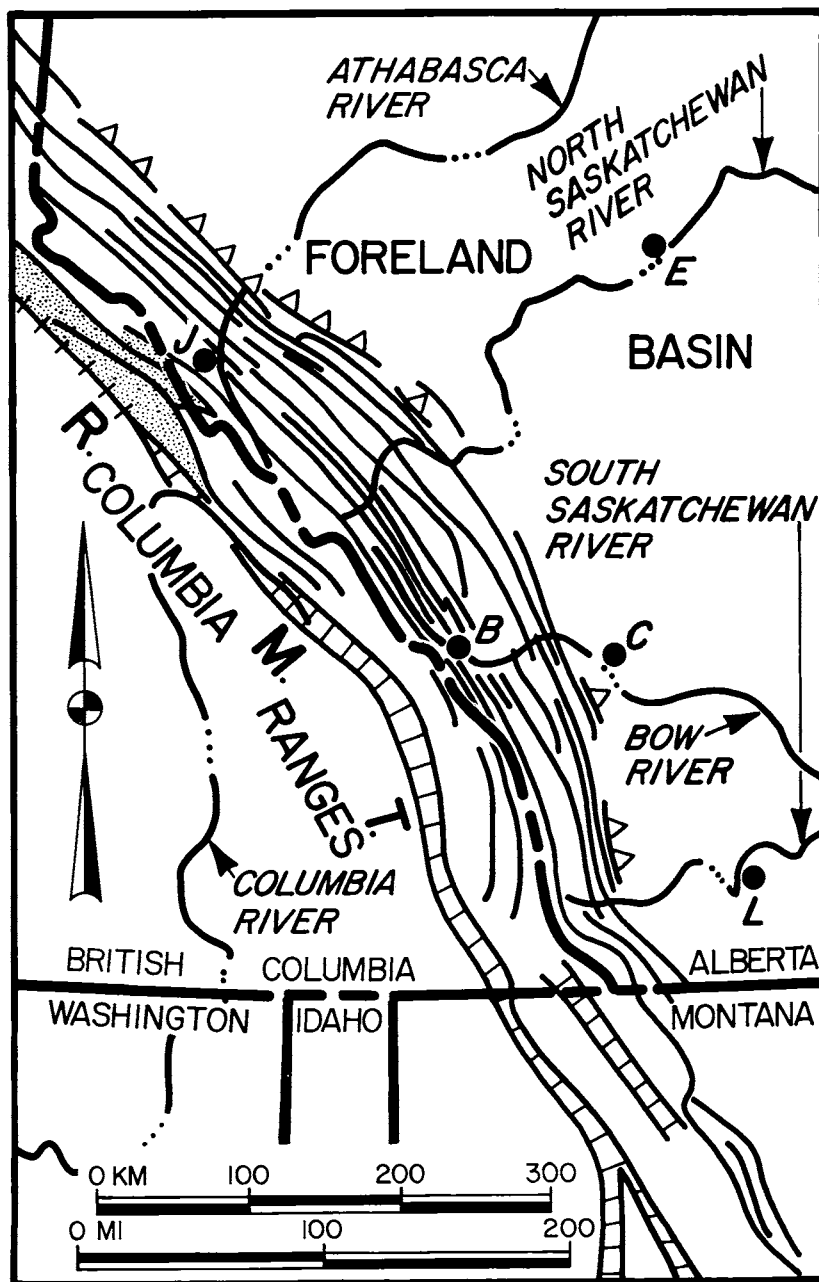


Fig. 5. Schematic tectonic map of the Canadian Rocky Mountains in Alberta and adjoining territory.

Unbroken line: Major thrust fault, mostly east-vergent (selective)

Line with open triangles: West-vergent thrust fault at front of folded belt—east margin of "triangle zone"

Stippled: Metamorphic rocks thrust into Western Ranges of Rocky Mountains

Double or single line with crossbars: Fault trough or fault-line valley of Rocky Mountain trench system

Heavy line with two dashes: International boundary

Heavy line with one dash: Province or state boundary

Line with three dots: River

Letters: B—Banff; C—Calgary; E—Edmonton; J—Jasper; L—Lethbridge;

R. M. T.—Rocky Mountain trench system

The stratigraphic sequence in the Canadian Rockies has many similarities with that in the Appalachian Valley and Ridge province but is considerably thicker and covers about half again as much time (even more if the underlying Middle Proterozoic Purcell strata be included). The basement belongs mainly to the Churchill province of the Canadian shield (last major metamorphism about 1700 Ma ago), and the overlying quartz sandstone, transgressing eastward onto the craton, is Cambrian under the eastern edge of the belt, Upper Proterozoic (Windermere) in the interior of the range. A great package of mainly massive, shallow-water carbonate extends from Middle Cambrian to Lower Carboniferous, underlain or interrupted by some shaly zones, the lower ones derived from the craton, those in the Devonian at least partly from uplifts deeper in the interior of the range. (A discordance like that within the Ordovician of the Appalachians (p. 323) appears to separate Devonian from older strata.) Thereafter clastic sediments became more abundant, and from Early Cretaceous to Paleocene time they flooded in from the west in the form of great delta complexes or clastic wedges.

Structurally, vergence is predominantly eastward toward the craton, and the belt can be divided into four longitudinal zones. In the Foothills to the east, the folds are fairly small, thrust faults are present but many are blind, and décollement occurs in the lower Mesozoic or upper Paleozoic strata, though upper Paleozoic rocks appear at the surface only exceptionally.³ In the Front Ranges (for example, the Ranges around Banff or east of Jasper), the full Paleozoic carbonate sequence is present (Mesozoic rocks are preserved here and there in synclines) and is generally broken into great thrust sheets: the thrust faults between, spaced 5 to 10 km apart, are long and roughly parallel, but no one extends the full length of the segment being discussed, let alone the whole salient. The main décollement levels here are in the shaly zones within the carbonate section, and especially at its base. In the Main Ranges along the continental divide, thick late Precambrian strata, more clastic than carbonate, join with the Lower and Middle Paleozoic carbonates; here large folds are prominent along with the thrust faults. In the Western Ranges, an abrupt facies change in the lower Paleozoic from massive shelf carbonate to argillaceous deeper water carbonate, along with increasing strain, cleavage, and incipient metamorphism, causes a change in style like that into the southeastern Valley and Ridge

³ In several sectors, at the very front of the Foothills belt, there is one west-vergent (east-dipping) thrust fault, and thus a narrow "triangle zone" between it and the first "ordinary" thrust fault is overridden from both sides but ordinarily is *not* depressed farther than the synclines between the thrust faults to the west or the undeformed strata to the east (Douglas, 1955, p. 53-56). In several places blind east-vergent faults have been encountered (in seismic profiles or drilling) beneath the triangle zone, so that the west-vergent fault is superficial. If movement on the blind faults had proceeded farther, the triangle zone would have been uplifted and quickly eroded away; thus as a rule such a zone can be preserved only at the very front of the deformed belt.

A similar structure (*Randunterschabung*) was reported by Renz (1937) and Habicht (1945, Tafel VI) at the edge of the "Subalpine Molasse" zone (p. 343) east of the Zürichsee.

province in Pennsylvania, and there are more exceptions to the dominant eastward vergence. Here also many faults originally thrust faults moved again as listric normal faults after the main deformation. Locally thrust sheets of frankly metamorphic rocks lap onto the Western Ranges, but mostly they lie farther west. There is however nothing like the Blue Ridge basement anticlinorium anywhere in the southern Canadian Cordillera; instead the Rockies are marked off from the more interior ranges (the Columbia Ranges) by the physiographic Rocky Mountain trench, which evidently is eroded along a series of continuous, or en echelon, strike (but only in part strike-slip) faults whose nature and continuity (and interpretation) vary greatly along the trend of the trench.

All the surface and subsurface data converge to prove the validity of the Rich model of ramp-and-flat thrusting in the Canadian Rockies, the flats being along various décollement levels and the ramps cutting from one to the next across the competent strata between. Beneath the whole edifice, the upper surface of the basement (with a bit of tegument) is virtually plane and descends gently from about 5 km below sealevel at the mountain front to about 12 km near the Rocky Mountain trench. Because the subsurface data are so unequivocal, the argument against décollement tectonics was rejected sooner here than in the Appalachians.

Although great thrust faults hundreds of kilometers long and with tens of kilometers of displacement are present in every cross section, nevertheless Price and Mountjoy (1970, p. 10) suggested that the overall deformation of the belt can be thought of as plastic rather than as brittle or rigid. The relative importance of gravity (spreading, *not* downhill sliding) and of true compressive shortening was vigorously debated (see among others the articles by Elliott and Chapple cited above, p. 343; see also below, p. 354–355).

For a long time, the evident involvement of Paleocene rocks in the folding and thrusting at the mountain front led to the view that all the deformation in the belt is Paleocene in age. It was only fairly recently (Price and Mountjoy again, 1970, p. 22–23), under the impact of the unequivocal evidence from the western Wyoming salient of the U. S. Western Rockies (see below, p. 347), that this view was overthrown and the progression (but not necessarily steady progression) of deformation eastward across the belt through almost the whole of Cretaceous time was accepted.

As mentioned above, the fold-and-thrust belt so well displayed in the Canadian Rockies continues southward nearly all the way across the United States (see Rodgers, 1987, fig. 1, p. 665), but it is much more broken up by Cenozoic normal faults. The only large unbroken segment south of the one in northern Montana (which strictly speaking is a continuation of the Canadian Rockies) is the western Wyoming salient (the “western overthrust belt” of the U. S. petroleum geologists). It extends barely 300 km from where it appears from under the late

Cenozoic Snake River volcanics in southeastern Idaho to where it is drowned in its own débris in the Bridger basin of southwesternmost Wyoming and adjacent Utah, just north of the transverse range of the Uinta Mountains in the Eastern Rockies. That part of the fold-and-thrust belt from the international border to northern Utah I call the Western Rockies (for definitions of Eastern and Western Rockies, and reasons for the terms, see Rodgers, 1987, p. 663–664). The rest does not have a general geographic name, but the whole, at least north to southwestern Montana, is now commonly called the Sevier orogenic belt, and the orogeny that formed it the Sevier orogeny (Armstrong, 1968), to distinguish it from the Laramide orogeny that formed the Eastern Rockies. Much of the belt, especially in Utah and Nevada but also well into Montana, is so badly cut up by Cenozoic Basin and Range normal faults that its essential continuity was long unrecognized. Moreover, it is delimited by complicated recesses in southwestern Montana and southeastern California, in both of which (as also locally in the smaller recess in Utah at the west end of the Uinta Mountains) basement rocks are clearly involved in the thrusting, which therefore loses its thin-skinned character.

It seems to be a general principle in fold-and-thrust belts that basement can be involved in the thrusting in recesses, especially the deeper ones, whereas in the salients completely thin-skinned décollement tectonics is the rule; as far as I know, the first to recognize this principle was W. A. Thomas (oral presentation, 1969; see Thomas, 1977, p. 1270). The reason is easy to grasp if we invert our viewpoint and replace the terms salient and recess (in the fold belt, as seen from the craton) by their equivalents embayment (or reentrant) and promontory (of the craton, as seen from the orogenic belt); as moreover the promontories generally bear a considerably thinner sedimentary cover than the embayments, clearly their basement would be more readily involved in shortening of the marginal part of the orogenic belt, even if it remained cold and rigid. To me, moreover, such basement involvement is easier to comprehend if the shortening results from true compression rather than from gravity spreading, but others may disagree.

The stratigraphy of the Sevier belt has some differences from that of the Canadian Rockies, but they are not important for our purposes; the structure is quite comparable, although the differentiation into longitudinal zones, where it can be made out, is different. I simply wish to recall that it was in the western Wyoming salient that the eastward progression of deformation throughout the Cretaceous was first unequivocally demonstrated (Armstrong and Oriel, 1965, building on earlier work by W. W. Rubey, 1955), though Longwell too (1950, p. 424) had seen and understood similar evidence in the southern Nevada part of the belt which, having been an artillery officer in the First World War, he described informally as a “rolling barrage.”

THE URAL

The Ural Range (fig. 6) is strikingly like the Appalachians in geography, structure, and even age (events are a period or two later); it is a little lower (the highest points are 1894 and 2037 m respectively) and much farther north, for its southern end, the Mugodzhar, is at about the

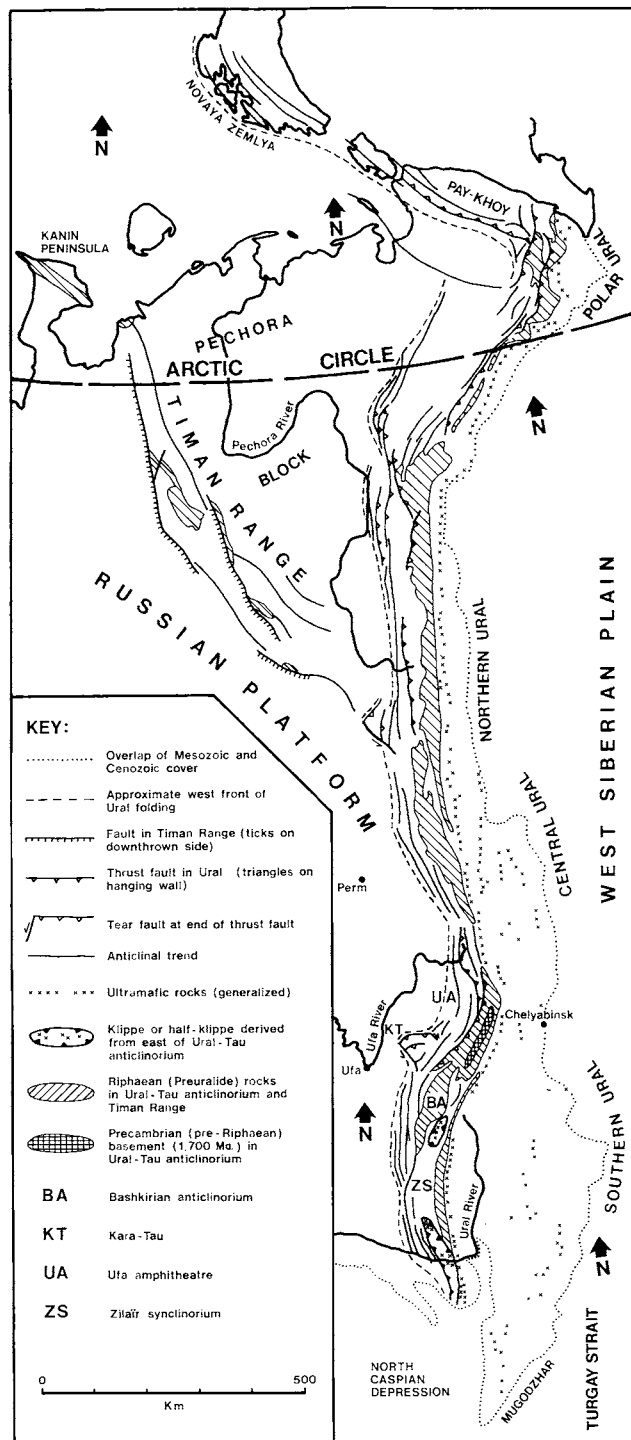


Fig. 6. Schematic tectonic map of Ural and Timan Ranges.

same latitude as the northern end of the Appalachians in Newfoundland. It extends roughly north-south for 2300 km (over 3000 km if the Pay-Khoy Range and the islands of Novaya Zemlya are included) along the eastern margin of the Russian or Baltic craton and displays a series of salients and recesses. Novaya Zemlya and the Pay-Khoy north of the Arctic Circle form one salient, though they are not generally considered part of the Ural proper; the chord of the salient is almost 1000 km, and its "height" 450 km. Its south end is a right-angle bend around the eastern corner of the Pechora block, a northeastern protuberance of the Russian craton half covered with Cenozoic deposits. The Polar, Northern, and Central Ural form a very large, perhaps compound salient, here called the Northern Ural salient, whose chord is 1400 km but whose "height" is only about 350 km. South of the Northern Ural salient is the double recess of the Ufa Amphitheatre (named for the headwaters of the Ufa River, not for the city, which lies to the southwest in front of the Southern Ural). The Southern Ural salient extends from there 550 km farther south to the point where the Ural River leaves the mountains to flow west and south to the Caspian Sea; the maximum "height" of this salient is about 200 km. Beyond the Ural River, the Mugodzhur may be the beginning of another salient, but within 350 km it plunges down beneath nearly flat-lying Cretaceous and Tertiary strata.

The width of the exposed Ural orogenic belt is on the order of 150 km in the Polar and Northern Ural but widens to a maximum of nearly 400 km in the Southern Ural. The west side fronts the Russian platform, where upper Paleozoic strata of the same age as those deformed in the Ural still lie nearly horizontal in a typical foreland basin. The east side of the Northern Ural salient disappears under the (mainly upper) Cenozoic strata of the West Siberian plain, that of the Southern Ural and the Mugodzhur under the lower Cenozoic (and locally Upper Cretaceous) strata in the Turgay "Strait" between the southwest corner of the plain and the Aral-Caspian depression south of the end of the mountains. The major watershed range, the Ural-Tau, is remarkably continuous and follows almost exactly a major anticlinorium of Blue-Ridge type between a "Valley-and-Ridge" province on the west slope and a "Piedmont" province on the east slope. Only the province of Valley-and-Ridge type is discussed here, and mostly that part in the Southern Ural, the only part I have visited and the subject of most of my reading. The modern literature on the Ural is mostly published in Russian in books that have had little circulation outside the Soviet Union (for example, Kamaletdinov, 1974).

The Precambrian of the Ural contains two elements. One is a gneissic-granitic basement, older than 1700 Ma, coeval and probably continuous with that under the nearer part of the Russian platform to the west; in the Ural, however, it is exposed for only about 140 km along the crest of the Ural-Tau anticlinorium opposite the recess of the Ufa Amphitheatre. The other consists of the rocks of a post-basement mountain range, the Pre-Ural (Russian: Doural or Douralidy), deformed

near the end of Precambrian or, at least in part, during Cambrian time. This older range seems to have a typical double structure: a miogeosynclinal, non-metamorphic fold belt on the west and a eugeosynclinal, metamorphic core zone on the east, the latter much less well known. The miogeosynclinal belt can be followed with a trend more or less coincident with that of the late Paleozoic Ural for nearly the full length of the Southern Ural salient, being exposed along the crest of the Ural-Tau and also forming the core of the Bashkirian anticlinorium (formerly called the Ufa horst), which rises within the "Valley-and-Ridge" province on the west side of the northern half of that salient. The belt probably continues northward for about 350 km along the Central Ural but then diverges to the northwest along the *southwest* side of the Pechora block (see fig. 6); it can be followed intermittently through the Timan Range (a chain of Late Paleozoic cratonal uplifts—Rodgers, 1987, p. 679—that follows the Pre-Uralide trend) onto the Kanin Peninsula, then even farther westward onto the Ribachiy (Fishermans) and Varanger Peninsulas along the Arctic coast west of the White Sea. The eugeosynclinal zone appears along the crest of the Ural-Tau anticlinorium through most of the Northern and Polar Ural, and it probably underlies most of the Pechora block.

The Pre-Uralian miogeosynclinal section is the Upper Proterozoic Rhiphaean "system" or super-group, consisting mainly of marine clastic and carbonate (stromatolitic) strata more than 10 km thick, with probably Vendian and perhaps Cambrian redbeds at the top. The Rhiphaean (named for the Ural, whose Latin name was Montes Riphæi) has its type region in the Bashkirian anticlinorium where the whole succession is well exposed, but it is also present, undeformed and thinning rapidly westward, above the basement in the subsurface of considerable parts of the eastern Russian platform.

The Paleozoic section on the western side of the Ural is again a typical miogeosynclinal wedge, reaching a maximum thickness of something like 8 km. It begins with Ordovician clastics; they are followed by a thick sequence of shallow-water carbonate, evidently the record of a great carbonate bank. In the eastern part of the belt, notably in and around the Zilaïr synclinorium (which lies east of the southern part of the Bashkirian anticlinorium and continues south to the Ural River), carbonate extends into the Middle Devonian and is overlain by a thick Middle and Upper Devonian flysch sequence, a clastic wedge that records a first major orogeny farther east in the Ural. (Like the Ordovician Taconic orogeny in the Appalachians, this orogeny appears to be the time of emplacement of all the main ultramafic bodies in the Ural, which form one of the world's longest and most nearly continuous ultramafic—"ophiolitic"—zones, running the entire length of the Ural along the east side of the Ural-Tau anticlinorium.) No younger strata than Upper Devonian are preserved in the Zilaïr synclinorium.

Farther west, however, the carbonate bank persisted well into the Carboniferous, when clastic wedges from the east began to encroach there as well, indicating renewed orogeny farther east, which probably then progressed westward. At the western edge of the fold belt, highest Permian and even Lower Triassic strata are deformed, though not

strongly. Thus the climactic orogeny probably lasted from Middle Carboniferous (Bashkirian or Namurian) through Permian time.

The fold pattern of the Southern Ural "Valley-and-Ridge" is much like that of the southern and central Appalachians, although the belt is somewhat narrower and is disturbed by the presence of the large Bashkirian anticlinorium, some of whose folds are Pre-Uralide though some are Uralide. Long continuous folds, some of them cut by thrust faults, verge west toward the craton. The Kara-Tau southwest of the Ufa Amphitheatre (Kara-Tau, which means Black Mountain in Turki languages, is a common name in the central part of the Soviet Union, and one must always specify which Kara-Tau is meant) consists of a group of folds and thrust slices bounded by strike-slip faults and projecting 40 km in front of the main fold belt into the otherwise flat-lying strata of the foreland basin at the margin of the craton. The pattern here is much like that of the Pine Mountain thrust sheet in the southern Appalachians, though the length along strike is only about 60 km. (There appears to be a similar structure in front of the southern part of the Northern Ural, just where the trend of the Pre-Ural diverges from that of the Ural, but it may belong to the Timan Range; I have no knowledge of that area.)

The Lower Paleozoic succession displayed in the main part of the fold belt is not the same as that exposed on the flanks of the Ural-Tau anticlinorium where, especially along the east flank, the lower Paleozoic strata are thinner and contain little carbonate but more argillite and chert, suggesting deeper water and a starved basin east of the carbonate bank up to Devonian time.

In 1932, Arkhangel'skiy (1932) summed up recently discovered evidence for thrust sheets or nappes of the second kind along the west side of the Ural, but later Russian geologists denied their existence and considered that each anticline and each thrust fault rooted separately in an uplift of the immediately underlying basement—typical thick-skinned tectonics. In the '60s, however, the Tatar geologist Kamaletdinov was able to direct an oil-drilling campaign to obtain critical subsurface information. One well, begun in Lower Devonian limestone in an anticline surrounded by Middle and Upper Devonian flysch (along the west side of the Zilaïr synclinorium south of the end of the Bashkirian anticlinorium), after penetrating Silurian limestone, passed abruptly, across a thrust fault, into Carboniferous *limestone*, then down through a complete carbonate section back into the Devonian (Kamaletdinov, 1965). Thus two very different facies of Upper Devonian strata are superposed, and the nearest surface thrust fault that could be the outcrop of the thrust fault in the well lies 20 km to the west, a minimum displacement for the fault. Other similar results have shown (Kamaletdinov, 1974) that the whole "Valley-and-Ridge" belt is thin-skinned; the Kara-Tau too has proved to be as much like Pine Mountain in the subsurface as on the surface.

The eastern part of the Ufa Amphitheatre (a topographic basin as well as a structural recess) is eroded into a thick, tightly folded Middle Carboniferous flysch (belonging to the Bashkirian stage, named here; it is the fully marine equivalent of the western European

Namurian stage). Here and there within the area of the flysch are small mountains, up to 5 km long, made of a heterogeneous assemblage of Ordovician, Silurian, and Devonian limestone and chert (but in general each mountain is made of only one rock type). These rocks were formerly interpreted as erosional islands buried by the onlapping flysch, but drilling directed by Kamaletdinov (1962) showed that in every case they rest *upon* the flysch. He interpreted them as erosional remnants of a single large thrust sheet, but they could also be interpreted as blocks that slid separately into the flysch basin (while deposition was going on) from the rising Ural-Tau anticlinorium, which half surrounds the Ufa Amphitheatre, again suggesting that orogeny had begun by Middle Carboniferous time.

WESTERN TAIWAN

The island of Taiwan is not quite 400 km long and is 150 km wide at its widest point, yet the highest peak of its Central Mountain Range stands 3997 m above sealevel. The west slope of the range and its Western Foothills contain a fold-and-thrust belt (fig. 7) some 250 km long and a few tens of kilometers wide, expressed in strata ranging from Miocene to Recent and clearly still growing (as is the Central Mountain Range); anticlines are becoming topographic highs, and historic earthquakes have broken the ground surface. Older Cenozoic strata appear mostly as tegument beneath the belt or, in greenschist facies, in the Central Mountains. Moreover drill holes and some seismic profiles provide considerable subsurface data; in particular the drill holes make it possible to measure the present fluid pressure of the formation waters, which in the deeper parts of the fold belt is well above hydrostatic, reflecting disequilibrium maintained by active sedimentation and deformation in rocks of low permeability. Here therefore information is available on a fold-and-thrust belt during its formation.

Recent published and unpublished work by John Suppe (1987 and references therein), who worked in western Taiwan and had access to the subsurface data, shows that the belt has many resemblances to the décollement-dominated belts I have described above but also some differences over and above its unfinished state, such as the virtual absence of carbonates (except in the youngest strata and then mostly outside the folded belt). The underlying basement, continuous with the bedrock of the adjacent Chinese mainland, includes granitic and metamorphic rocks as young as Mesozoic; its upper surface is about 4 km deep at the west coast of the island (less under the Taiwan Strait to the west) but slopes about 6° eastward, reaching depths of 8 km or more near the east edge of the folded belt; it is nearly plane, except where broken by Oligo-Miocene down-to-basin faults. The sediments above form a great wedge thickening eastward; the wedge taper is increased by the deformation but would exist in any case.

Up to the Pliocene, these clastic sediments were all derived from the west, from what is now the Chinese mainland; only then, and at first only in northern Taiwan, did sediment begin to arrive from the rising mountain range to the east. Although the first sediments above the basement were deposited in shallow marine water (like the entire

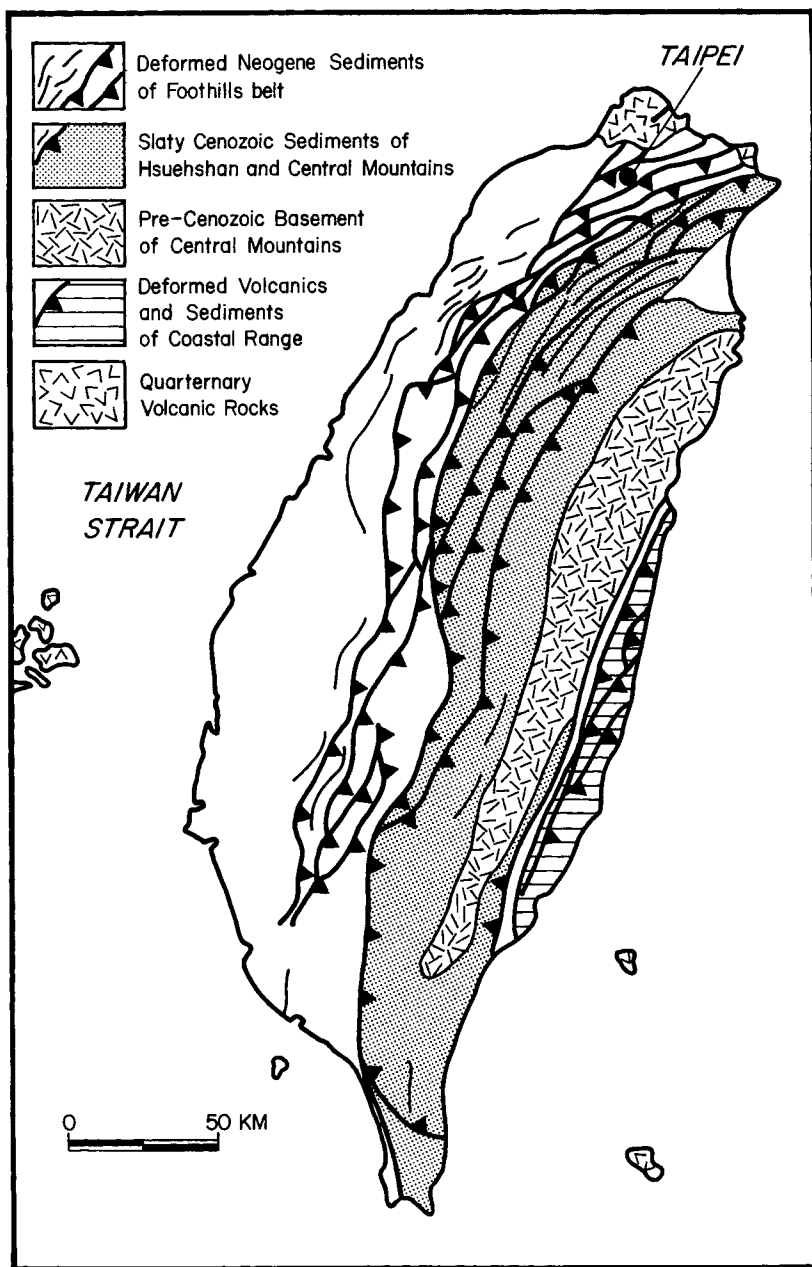


Fig. 7. Schematic tectonic map of Taiwan (after Suppe, 1987, fig. 4). Figure reproduced by permission of the copyright holder: Princeton University Press, Princeton, N.J.

sequence under the Taiwan Strait), the upper Miocene sediments in the eastern part of the belt were deposited in much deeper water, at depths up to those in the present Manila trench south of Taiwan (3–4 km). The youngest sediments in the fold belt are again shallow marine or indeed continental deposits derived from the Central Mountain Range. The time of maximum water depth is roughly that of the shift from west-derived to east-derived sediments in each cross section; moreover that time is clearly diachronous, ranging from early Pliocene (about 4 Ma ago) near the north end of the belt southwest of Taipei to the present at a point about 150 km south of the south tip of the island, near the north end of the deep part of the present Manila trench. Therefore, as Suppe has shown (1987, p. 280–281), successive east-west cross sections from that point northward can be considered as successive time-shots of the growth of the fold belt and of the island through the last 4 my. In and south of Taiwan, the Philippine Sea plate is impinging on the Asian plate, which includes the oceanic South China Sea, but at an angle to the continent-ocean boundary on the Asian plate, so that deformation of the sedimentary pile deposited along the continental margin on that plate, in the former northward extension of the Manila trench, began first in northern Taiwan, is now active over most of Taiwan and for a hundred kilometers to the south, and will in the future involve more and more of the trench.

Suppe (1983) has carried through consistently the Rich ramp-and-flat model for thrusting, with folds made of plane panels joined by angles or by relatively short arcs, and has shown that the model can be used to predict quite accurately the subsurface structure. Décollement levels are determined by the static properties of the sediments and by overpressuring in the formation waters. All in all, western Taiwan provides an unusual look at a fold-and-thrust belt being formed as we watch.

We can use this example to test mechanical hypotheses for the formation of such belts, in particular the rival hypotheses of shortening by real compression and of gravity spreading. The height of the Central Mountain Range certainly provides gravitational potential to deform the wedge of sediments between it and the Chinese basement, but the height of the range itself must be accounted for. It is not simply a static uplift, resulting from the uplift of an already existing, underlying basement: it is dynamic and diachronous, and the basement is being depressed beneath it. Hence it must be produced by lateral compression of the sedimentary cover as the Philippine Sea plate drives against the margin of the Asian plate (Dahlen, 1988); to use the simile carried out by Davis, Suppe, and Dahlen (1983), the Philippine Sea plate acts like a giant bulldozer pushing the relatively soft and (considered in bulk and to scale) plastic sediment ahead of it, so that both the steep slopes of the mountain range and the resulting gravitational potential are simply the result of the push of the bulldozer (of course only relative motion is implied; see above, p. 332 and footnote on p. 333). The force of gravity is always there, always working to bring down uplifted materials and

using their quasi-plastic behavior to convert its own downward pull into lateral shove ("gravity spreading"), but the height of the pile that permits gravity to work at all is itself the result of lateral compression, which is thus the primary force.

INTERIM SUMMARY

These six fold-and-thrust belts thus display many common characters and yet a considerable variety—in size and age, for instance. In general each formed from a wedge of miogeosynclinal sediments that thickens dramatically from the cratonal to the internal edge of the belt (the Jura is an exception, but perhaps it can be considered an over-sized Pine Mountain belt nearly detached from the related Alpine belt). The lower part of each wedge is mainly carbonate rock, except in western Taiwan; the upper part consists of deltaic complexes or clastic wedges (not in the Jura, and to only a minor extent in the Subalpine-Helvetic zone of the Alps), and the later parts of those complexes were contemporaneous with the onset of the folding and the thrusting.

The folding and thrusting were thin-skinned over a major décollement or décollements in the lower part of the sedimentary sequence, on evaporites or weak, probably over-pressured shale zones; basement is involved, if at all, only locally, mainly in recesses. Both folds and thrusts verge predominantly toward the adjacent craton (again the Jura is an exception, presumably because the sediments involved do not form a wedge). Fold trends in the salients are smoothly curved, filling embayments in the margins of the cratons; those in the recesses are more angular, appearing to intersect over promontories of the craton. The Rich ramp-and-flat model clearly applies to the outer parts of all these belts, but more internally the rocks may show finite strain and ductile behavior. The western Taiwan belt is forming today under compression caused by the convergence of well characterized lithospheric plates, and in my opinion all the others formed in the same way, at earlier times.

In the second part of this article, I will mention in less detail a number of other fold-and-thrust belts around the world, emphasizing those that differ in instructive ways from these typical examples.

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Probably I have spent more of my professional life in the Appalachian Valley and Ridge province than in any other region, and I have been helped to understand it by scores of other geologists; I cannot name them all, but I must mention two. Professor Charles M. Nevin started me off there at his summer field camp in central Pennsylvania; indeed, one of the mapping projects he set me included what later turned out to be the Birmingham window, a key to understanding the central Appalachians. Dr. Josiah Bridge steered me to East Tennessee and helped and inspired me in my work there. I was introduced to the Alps in 1959–1960, when Professor Paul Fallot accepted me in his laboratory at le Collège de France and arranged for leading geologists to guide me in the field. Among the many who showed me the Subalpine

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DEVONIAN PLANTS FROM SOUTHERN QUEBEC AND NORTHERN NEW HAMPSHIRE AND THE AGE OF THE CONNECTICUT VALLEY TROUGH

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ABSTRACT. The gray metasedimentary rocks of the Connecticut Valley trough extend from the east end of the Gaspé Peninsula southwest and south across Quebec and western New England. In northern New England, these rocks belong to the Waits River Formation and Standing Pond Volcanic Member and to the Gile Mountain Formation and its Meetinghouse Slate and Northfield Members (in ascending order). Ages based on reports of fossils in these rocks in New England have long been controversial and range from Middle Ordovician to Early Devonian. Many of the fossils have been proven to be inorganic, some misidentified, and others severely questioned.

Following a review of the earlier "fossil" finds, we herein present descriptions of Early Devonian fossil plants found in rocks of the upper part of the stratigraphic section composing the Connecticut Valley trough. We also present U-Pb data from analyses of zircons from a dike near Springfield, Vt., indicating a Silurian age for some rocks of the trough. Four groups of plants have been recovered from eleven sites: Nematophytales; Zosterophyllophytina; Lycophytina; and Trimerophytina. The best preserved and most numerous specimens occur in southern Quebec in very low grade metaturbiditic rocks of the Compton Formation (correlative with the Gile Mountain Formation). One site in northern New Hampshire contains fragmental *Prototaxites*. Farther south near Montpelier, Vt., and down section, poorly preserved echinoderm fragments and "pseudograptolites" that may be interpreted as plant fragments (*incertae sedis*) occur in the Waits River Formation. The presence of plant fossils in northern New Hampshire and southern Quebec supports a late Early Devonian (Emsian) age for the Gile Mountain Formation and its equivalent, the Compton Formation; the equivocal paleontological data in Vermont and the radiometric data indicate a Silurian age for the Waits River Formation. We conclude that the rocks of the Connecticut Valley trough accumulated continuously beginning as early as the Silurian and extending into the latest Early Devonian.

INTRODUCTION

The Connecticut Valley-Gaspé synclinorium (Cady, 1960), or Connecticut Valley trough (Hatch, 1988), is a distinctive belt of rocks that

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