CHAPTER 2
DEVONIAN AND EARLY CARBONIFEROUS LAURUSSIA

INTRODUCTION

With the late Caledonian suturing of Laurentia–Greenland and Fennosarmatia–Baltic along the Arctic–North Atlantic Caledonides, the Laurussian megacontinent was formed. This large cratonic block remained intact for some 350 Ma until its break-up along the Arctic–North Atlantic Caledonian megasuture during the earliest Eocene.

Following the latest Silurian–earliest Devonian consolidation of the Arctic–North Atlantic Caledonides, the Iapetus subduction system became inactive. This entailed a fundamental plate reorganization.

POST-CALEDONIAN PLATE REORGANIZATION

In the North Atlantic domain, the three-plate convergence system, which governed the evolution of the Caledonian fold belts of Western and Central Europe (Ziegler, 1982a; Soper and Hutton, 1984), gave way at the transition from the Silurian to the Devonian to a two-plate convergence system (Fig. 2). The Devonian and Early Carboniferous evolution of the southern margin of Laurussia was governed by the continued northward subduction of the Proto-Tethys plate at an arc-trench system extending from the Appalachian domain along the southern margin of the Ligerian–Moldanubian Cordillera (Baker and Gayer, 1985) into the area of the Caucasus. For the latter, Adamia et al. (1981) visualize a more southerly located intraoceanic Pontid–Transcaucasus arc that was separated by the Greater Caucasus interarc basins from the Greater Caucasus arc-trench system marking the southern margin of the Sarmatian platform (Plates 2–4).

Devonian subduction of the Proto-Tethys plate was accompanied by the continued northward rafting of the Gondwana-derived Traveler and Avalon–Meguma microcontinents (O'Brian et al., 1983; Williams and Hatcher, 1983; Baker and Gayer, 1985; Keppie et al., 1985) and of the Aquitaine-Cantabrian and the more hypothetical Intra-Alpine continental fragments (Perroud et al., 1984; Ziegler, 1986). These cratonic blocks collided during the Middle Devonian with the Proto-Tethys subduction complex, giving rise to the Acadian–Ligerian orogeny (Fig. 1). It is suspected that the convergence rates between the Proto-Tethys and the Laurasian plates varied though Silurian to Early Carboniferous times. This is evident by the Early Devonian development of an extensive back-arc rift system in Western and Central Europe that remained intermittently active until the Early Carboniferous onset of the Variscan orogeny (Ziegler, 1982a, 1984; Behr et al., 1984). Back-arc extension was, however, temporarily interrupted during the Middle Devonian Acadian–Ligerian orogeny and during the Bretonian orogeny at the transition from the Devonian to the Carboniferous (Plate 2).

During the Devonian and Early Carboniferous, the oceanic Ural plate continued to converge with Laurussia (Zonenshain et al., 1984). However, with the final locking of the Iapetus megasuture and the cessation of related subduction processes in the domain of the Arctic–North Atlantic Caledonides at the transition from the Silurian to the Devonian, the last step in the suture progradation from the Arctic–North Atlantic Caledonides to the Sakmarian arc-trench system, which had already begun during the Middle Silurian, was effected during the Early Devonian. Correspondingly, the Devonian and Carboniferous evolution of the eastern margin of Laurussia was governed by the Sakmarian (-Magnitogorsk) subduction system (Plates 2–4).

Continued convergence of the Siberian craton with Laurentia–Greenland culminated during the latest Devonian and earliest Carboniferous in the Ellesmerian orogeny, which resulted in the consolidation of the Innuitian fold belt (Trettin, 1973; Trettin and Balkwill, 1979; Fig. 1).

These plate movements were paralleled by a major sinistral translation between the Laurentia–Greenland and the Fennosarmatian subplates along a complex fracture system that transected the axis of the Arctic–North Atlantic Caledonides (see *Arctic–North Atlantic Megashear,* this chapter; Ziegler, 1984; Harland et al., 1984). The development of this megashear can be considered as evolving from the sinistral oblique collision of Laurentia–Greenland and Fennosarmatia during the Caledonian orogeny (Soper and Hutton, 1984). The Devonian and Early Carboniferous northward translation of Fennosarmatia relative to Laurentia–Greenland was presumably accompanied by the progressive closure of the hypothetical Lomonosov ocean and ultimately the collision of the Barents Shelf with the Siberian block and their suturing along the largely hypothetical Lomonosov fold belt (see *Innuitian Fold Belt*, this chapter).

The northern foreland of the Innuitian–Lomonosov fold belt is formed by the East Siberian craton, encompassing Chukotka and the New Siberian Islands. It is uncertain whether this block was separated during the Late Devonian–earliest Carboniferous by an oceanic domain from the West Siberian block or whether these were connected and formed together the larger Siberian Craton, as suggested in Fig. 7 (Scotese et al., 1985). If the latter applies, it may be assumed that during the consolidation of the Innuitian–Lomonosov fold belt a shear zone transected the Siberian Craton in the prolongation of the Sakmarian arc. Such a shear zone could have facilitated the Early Carboniferous separation of the West Siberian block from the East Siberain block and the subsequent convergence of the former with the eastern margin of Fennosarmatia (see *Innuitian Fold Belt*, this chapter, and Ziegler et al., 1979).

The Devonian and Early Carboniferous evolution of Laurussia is summarized by Plates 2–4. On these maps, the palinspastic reconstruction of areas located to the south of the Variscan deformation front is incomplete. Major tectonic units and facies belts shown for areas that later became overprinted by the Variscan and Alpine orogenies are therefore distorted to varying degrees. Moreover, the outline of continental fragments (allochthonous terranes) that converged during the Devonian with the Proto-Tethys subduction zone is only shown schematically on these maps.

VARISCAN GEOSYNCLINAL SYSTEM

Following the late Caledonian diastrophism, back-arc extension affected the Central Armorican–Saxothuringian–Barrandian successor basins as well as the Mid-European Caledonides. In the area of the latter, the Rhenohercynian Basin (Cornwall–Rhenish–East Sudetic Basin) began to subside during the Early Devonian under a tensional regime (Fig. 3; Ziegler, 1982a). Together, these basins formed important parts of the geosynclinal system out of which the Variscan fold belt of Europe developed during the Late Carboniferous. Their sedimentary record is summarized in Plates 23 and 24. The Rhenohercynian and the Central Armorican–Saxothuringian basins were separated in the west by the Normannian–Mid-German trend of highs but were connected in the east via the East Sudetic Basin. Marine transgressions entered the Rhenohercynian Basin from the east as well as from the west during the Gedinnian, with deeper water conditions being established in its axial parts during the Siegenian and Emsian. A major, in part reef-fringed, carbonate platform occupied the northern margin of the Rhenohercynian Basin during the Middle and Late Devonian and during the Early Carboniferous (Czerminski and Pajchlowa, 1974; Gardiner and MacCarthy, 1981; Ziegler, 1982a; Engel et al., 1983; Langenstrassen, 1983). Clastic influx from the Mid-German High abated during the Siegenian but accelerated again during the Late Devonian and Early Carboniferous (Weber, 1984; Franke and Engel, 1986; Plates 2–4).

In the Central Armorican Basin, Early Devonian shallow marine carbonates and clastics accumulated in depositional continuity with Silurian marine strata (Plate 25). A major influx of deltaic and continental clastics from the Ligerian Cordillera occurred during the Middle Devonian and again during the Early Carboniferous. The latter was preceded by a late Famennian pulse of deeper water flysch deposition. These pulses of clastic influx reflect the orogenic reactivation of the Ligerian Cordillera during the Acadian-Ligerian and the Bretonian diastrophic cycles (Fig. 1; Lardeux et al., 1977; Guillocheau and Rolet, 1982; Rolet, 1983).

In the Saxothuringian Basin, in which marine sedimentation was continuous across the Silurian–Devonian boundary, Devonian and Early Carboniferous series are represented by deeper marine shales and clastics (Plate 25). Shallow marine carbonates developed on local intra-basinal highs during the Middle and Late Devonian and the Early Carboniferous. In the Barrandian Basin, Early and Middle Devonian carbonate deposition was interrupted during the Givetian by the influx of flysch derived from the Moldanubian Cordillera. This is taken as evidence that the Moldanubian Cordillera was also affected by the Acadian–Ligerian diastrophism. The Bretonian orogenic pulse is reflected in the Saxothuringen Basin and in Moravia by the onset of the Early Carboniferous Culm flysch deposition and the reactivation of a south-plunging A-subduction zone (Ziegler, 1982a; Weber, 1984).

The Devonian and Early Carboniferous evolution of the Central Armorican–Saxothuringian and Rhenohercynian back-arc basins was accompanied by a repeated rift-induced, intracontinental alkaline mafic-felsic bimodal volcanism (Floyd, 1982; Ziegler, 1982a, 1984, 1986; Wedepohl et al., 1983; Behr et al., 1984). Facies patterns in these basins were, at least in part, controlled by rift tectonics and by the development of rift-induced volcanic edifices. In the Rhenohercynian Basin, back-arc extension probably progressed during the Early Devonian in some areas to crustal separation and the opening of oceanic basins, as for instance in the Cornwall subbasin (Badham, 1982; Isaac and Barnes, 1985; Rolet et al., 1986) and in the southern parts of the Rhenish subbasin (Engel et al., 1983).

In the Central Armorican and Saxothuringen basins, back-arc extension and volcanism were temporarily interrupted during the Mid-Devonian Acadian–Ligerian and the latest Devonian–earliest Carboniferous Bretonian orogenic pulses. The Early Devonian to earliest Carboniferous evolution of the Rhenohercynian Basin, on the other hand, was governed by persistent back-arc extension as indicated by an almost continuous bimodal alkaline volcanism and a generally low level of clastic influx from the Normannian–Mid-German High (Plate 24).

The eastward continuation of the Variscan geosynclinal system into the Dobrugea–Black Sea area and the Caucasus, schematically shown on Plates 2–4, is essentially based on the paleogeographic maps of Vinogradov (1969). The southeastward prolongation of the Moravian Basin, along the southwestern margin of the East Silesian Massif, is conjectural.

The Donets Graben, which came into evidence during the Middle and Late Devonian, may be considered as forming part of the Devono-Carboniferous back-arc rift system. Its subsidence was accompanied by the updoming of the Ukrainian and Voronesh highs (Vinogradov, 1969).

The Early Devonian development of an extensive back-arc rift system in Western and Central Europe may be interpreted as resulting from a post-Caledonian decrease in the convergence rate between the oceanic Proto-Tethys and the continental Laurussian plates. According to the subduction model of Uyeda (1982), such a decrease of the convergence rate between two colliding plates is associated with a steepening of the Benioff zone and a partial decoupling of the subducting and the overriding plate. With this, compressive stresses exerted on the back-arc areas decrease to the degree that back-arc convection systems can assert themselves and thus can give rise to back-arc extension. This mechanism is apparently reversible if convergence rates accelerate again (Fig. 4).

This model suggests that the Devonian to Early Carboniferous evolution of the Variscan geosynclinal system was governed by the interplay of back-arc extension and back-arc compression, whereby periods of back-arc extension were characterized by an alkaline-bimodal intraplate volcanism and periods of back-arc compression by the shedding of synorogenic clastics from orogenically reactivated highs into the adjacent basins.

The general evolution of the Variscan geosynclinal system shows that it began to subside under a back-arc extensional setting immediately after the late Caledonian orogenic pulse. During the Middle Devonian Acadian–Ligerian orogeny, the southern parts of the Central Armorican, the Saxothuringian, and the Barrandian basins became deformed by back-arc compression while the Rhenohercynian Basin remained in a tensional setting. Back-arc extension resumed in the Central Armorican and Saxothuringian basins during the Late Devonian but was again interrupted during the Bretonian orogeny. At this time, compressive stresses apparently affected also the Mid-German High from which clastics (culm-flysch) were shed into the southern parts of the Rhenohercynian Basin (Weber, 1984). During the Tournaisian and early Visean, back-arc extension again controlled the evolution of the Central Armorican–Saxothuringian and of the Rhenohercynian basins as indicated by renewed rift-induced volcanic activity (Plates 24, 25).

In the following chapters, the geodynamics of the Acadian–Ligerian and Bretonian orogenic cycles are further discussed.

ACADIAN–LIGERIAN AND BRETONIAN OROGENIES

The Middle Devonian Acadian–Ligerian orogeny and the Bretonian orogeny straddling the Devonian–Carboniferous boundary resulted from the progressive northward subduction of the oceanic Proto-Tethys plate and the ensuing collision and accretion of Gondwana-derived continental fragments (allochthonous terranes) to the southern margin of Laurussia. In view of their association with the Proto-Tethys subduction zone,

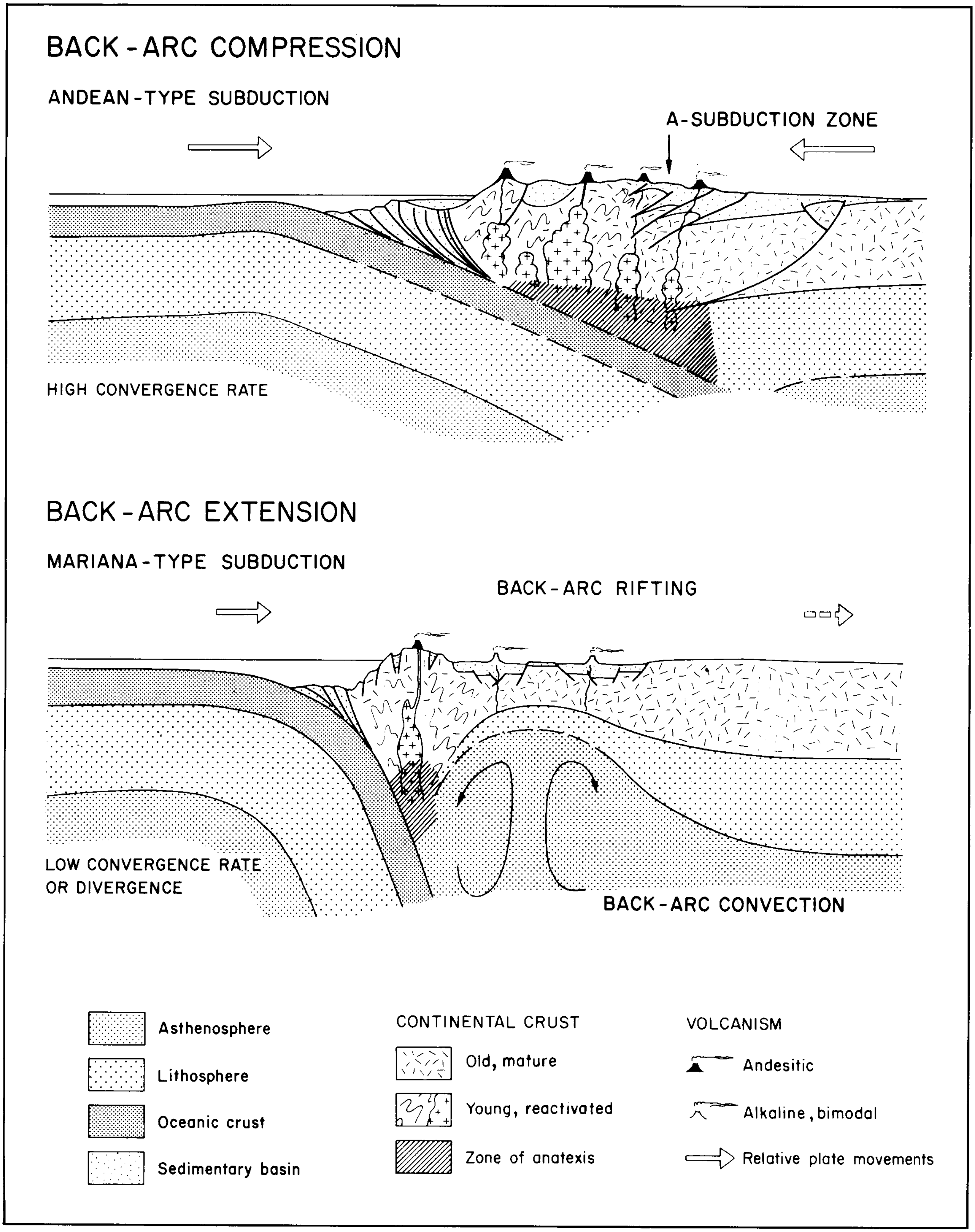

Figure 4—Subduction models. Modified after Bally and Snelson (1980) and Uyeda (1982). (Not to scale.)

both the Acadian–Ligerian and the Bretonian fold belts can be regarded as "Pacific-type" or "Accretion-type" orogens, as defined by Uyeda (1982). These intermittent orogenic cycles were possibly associated with temporary increases of the convergence rate between the Proto-Tethys and the Laurussian plates and partly correspond to collisional events during which Gondwana-derived continental fragments became accreted to the southern margin of Laurussia. Paleomagnetic data suggest that during the Devonian and the Early Carboniferous the Proto-Tethys narrowed progressively, culminating in the Visean collision of Gondwana and Laurussia and the onset of the "Himalaya-type" Variscan orogeny (Jones et al., 1979; Perroud and Bonhommet, 1981; van der Voo, 1983; Perroud et al., 1984; Ziegler, 1984, 1986).

Acadian Orogeny

During the Middle Devonian Acadian orogeny, the composite Traveler–Avalon–Meguma terrane collided with the Appalachian subduction system and became accreted to the southern margin of Laurentia (Poole, 1977; Schenk, 1978, 1981; Bradley, 1983; Baker and Gayer, 1985; Keppie, 1985). These terranes apparently became rifted off the northern margin of Gondwana sometime during the Ordovician–Early Silurian. Their Late Ordovician Gondwana affinity is emphasized by the occurrence of glacio-marine deposits in Nova Scotia (Schenk and Lane, 1981).

The megasuture between Laurentia and the Traveler–Avalon–Meguma terrane is characterized by intense deformations, metamorphism, and an extensive Middle to Late Devonian plutonism (Williams and Hatcher, 1983; Chorlton and Dallmeyer, 1986). Consolidation of the Acadian fold belt was accompanied by regional sinistral and local dextral shear movements (Keppie et al., 1985). Because much of the Avalon–Meguma terrane is now buried beneath Mesozoic and Tertiary sediments on the shelves of Nova Scotia and Newfoundland, it is difficult to assess to what extent this cratonic block was deformed during the Acadian orogeny. In Nova Scotia, where the Meguma terrane is exposed, it is characterized by open folds and numerous granitoid intrusions that range in age between 360 and 370 Ma (Reynolds et al., 1981; Keppie, 1982, 1985). These intrusions may be interpreted as being related to the Middle to Late Devonian development of a new B-subduction system along the eastern margin of the Avalon–Meguma block facing the Proto-Tethys. The second, Late Acadian orogenic phase straddles the Devonian-Carboniferous boundary (Keppie, 1985) and is coeval with the Bretonian orogeny of Western and Central Europe. In southwest Newfoundland, time-equivalent metamorphism and granitic intrusions are thought to be related to Bretonian wrench faulting (Chorlton and Dallmeyer, 1986).

In the Canadian Maritime provinces, the accumulation of neo-autochthonous red beds commenced during the late Givetian in the Fundy synclinorium (Magdalen Basin) and possibly also in the Sidney and the St. Anthony basins (Fig. 5). The evolution of these basins was accompanied by repeated wrench deformations and a generally bimodal intraplate volcanism that can be related to the Middle Devonian to earliest Carboniferous Arctic–North Atlantic translation (see Chapter 2). During the Early Carboniferous, marine incursions originating from the Rhenohercynian Basin reached the St. Anthony, Sidney, and Fundy basins where they gave rise to the accumulation of evaporites and carbonates (Plate 22; Howie and Barrs, 1975a, 1975b; Jansa et al., 1978; Bradley, 1982; Barr et al., 1985; Fyffe and Barr, 1986).

Ligerian Orogeny

The Ligerian orogeny of Western and Central Europe, which is roughly time equivalent with the Acadian orogeny, was associated with the collision of the Gondwana-derived Aquitaine–Cantabrian and the more hypothetical Intra-Alpine terranes with the Proto-Tethys arc-trench system paralleling the southern margin of the Ligerian–Moldanubian Cordillera. At the same time, the Aquitaine–Cantabrian microcraton collided presumably with the eastern margin of the Avalon–Meguma terrane to which it became sutured along the Central Iberian fold belts (Ziegler, 1986).

It is not certain whether contemporaneous deformations occurred along the southeastern margin of Fennosarmatia. For the Greater Caucasus area, Adamia et al. (1981) indicate the Devonian development of a tensional intra-arc basin in which flysch series accumulated.

The development of the Central Iberian fold belt is related to the closure of a relatively narrow Siluro-Ordovician oceanic basin, the opening of which was preceded by Late Cambrian to Ordovician rifting (Iglesias et al., 1983; Ribeiro et al., 1983) and the intrusion of Ordovician anorogenic granitoids (Priem and den Tex, 1984; Casquero et al., 1985; Lancelot et al., 1985). The occurrence of Ashgillian glacial deposits (Arbey and Tamain, 1971; Carls, 1975; Robardet, 1981) indicates that the Aquitaine-Cantabrian microcontinent became separated from northern Africa presumably during the Early Silurian and began to converge with the South Portuguese Craton with which it ultimately collided during the Middle Devonian. Consolidation of the resulting Central Iberian fold belt was associated with the obduction of a major, east-verging ophiolite nappe in Galicia (Martinez-Garcia, 1972, 1980; Iglesias et al., 1983). In the Ossa Morena zone of Central Iberia, Givetian emplacement of northeast-vergent nappes was followed by sinistral tangential deformations (Chacón et al., 1983). This suggests an oblique collision between the Aquitaine–Cantabrian microcontinent and the South Portuguese Craton; the latter may well form part of the larger continental Avalon–Meguma terrane (Plate 2).

The Middle Devonian consolidation of the Central Iberian range was followed by back-arc extension in the area of southwest Iberia as evident by a Late Devonian–Early Carboniferous bimodal volcanism in the Pyrite Belt and the development of the South Portuguese–Ossa Morena Basin in which Late Devonian and Early Carboniferous shallow marine clastics and flysch series accumulated (Oliveira, 1982; Julivert et al., 1983). This could be taken as an indication that a new subduction complex had developed during the Middle to Late Devonian to the south of Iberia.

In the South Armorican and the Central Massif of France, the Ligerian orogeny gave rise to high-pressure metamorphism, important plutonic activity that persisted into the Late Devonian and the emplacement of major south-vergent basement involving nappes (Bernard-Griffiths et al., 1977, 1985; Autran and Cogné, 1980; Autran and Dercourt, 1980; Matte, 1983, 1986). These important deformations are compatible with the concept that the Ligerian orogeny corresponds to a major continent–continent collisional event (Brun and Burg, 1982; Burg et al., 1987).

The above suggests that by the end of the Acadian–Ligerian orogeny the Aquitaine–Cantabrian terrane was rimmed on three sides by coherent fold belts. Its stratigraphic record shows, however, that in itself this stable cratonic block was little deformed and that its central parts continued to be occupied by carbonate and mixed carbonate-clastic platforms, which were open to the Proto-Tethys ocean to the southeast (Julivert et al., 1983; Kulmann et al., 1982; Plate 25). The deeper water troughs flanking this platform and the Central Iberian and Ligerian ranges were, however, largely destroyed during the Late Carboniferous Variscan orogeny.

Although there is stratigraphic evidence that the Moldanubian Cordillera was also affected by the Ligerian diastrophism, there is only a limited radiometric record of Middle Devonian granitoid plutonism (Bernard and Klominsky, 1975; see Plate 25, Barrandium). It is possible that during this orogenic phase the Barrandian Block became sutured to the Saxothuringian Block (W. Franke, personal communication, 1987).

Areas affected by the Acadian–Ligerian diastrophism possibly extend from the Massif Central, the West Alpine External Massifs, the Vosges, and the Bohemian Massif into the area that is now occupied by the autochthonous basement of the Alps and the basement of the Penninic and Lower and Middle Austro-Alpine nappes (Autran and Cogné, 1980; Lameyre and Autran, 1980; Kornprobst, 1980). The latter contain Siluro-Ordovician and Early and Middle Devonian sedimentary sequences and have yielded limited evidence for a Middle to Late Devonian plutonism (Schönlaub, 1979). Radiometric age determinations indicate, moreover, that the Variscan basement complex of the Central Alps contains Cadomian and older continental crustal elements that were apparently affected by a Caledonian distensional phase (Gebauer and Grïienfelder, 1982). This is compatible with the hypothesis that these areas consist of Gondwana-derived continental fragments and thus represent a truly allochthonous terrane, referred to here as the Intra-Alpine terrane, which was probably accreted to the southern margin of Fennosarmatia during the Ligerian diastrophism. More radiometric datings are, however, required to confirm this

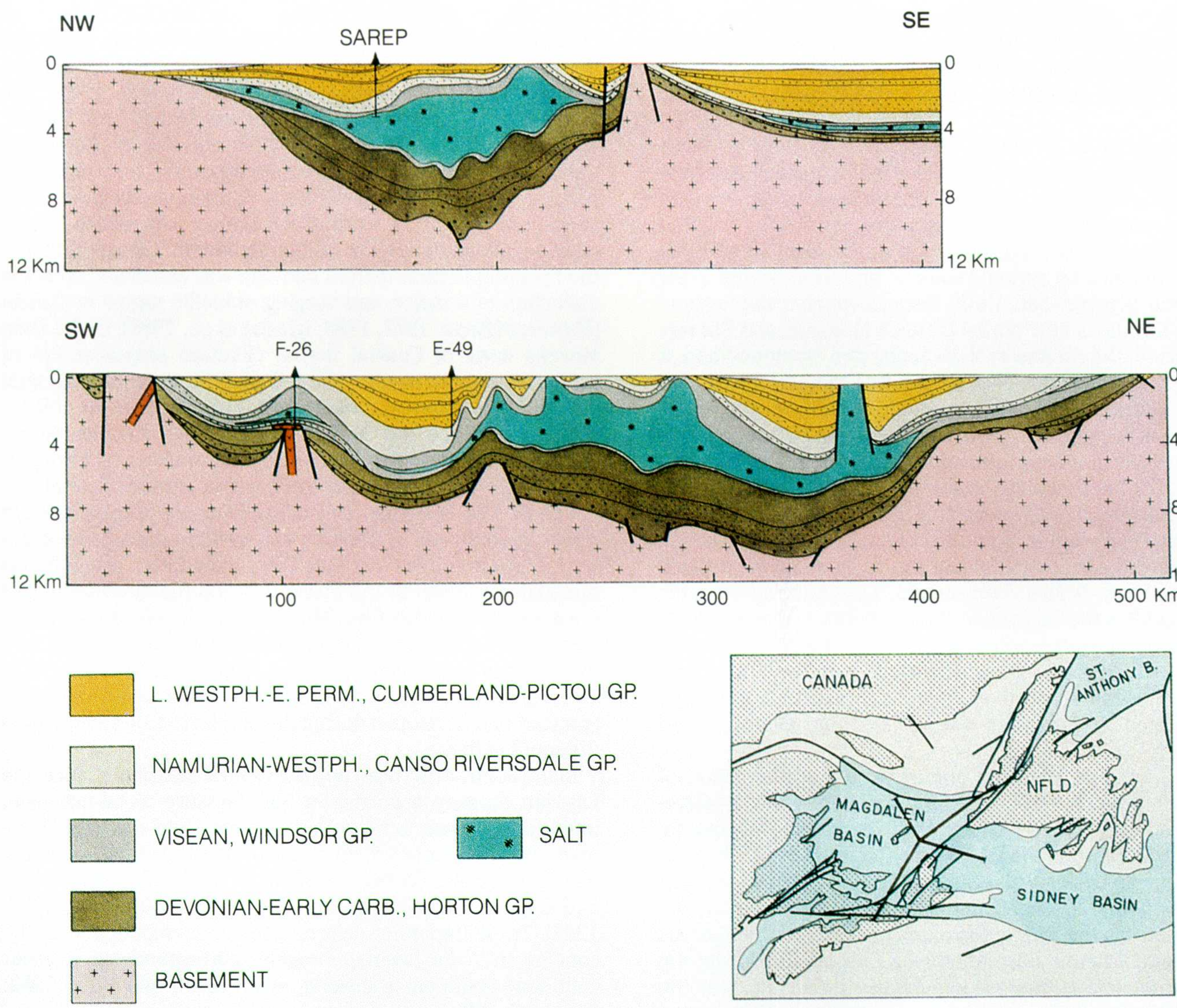

Figure 5—Regional structural cross sections through Magdalen Basin, Gulf of St. Lawrence. After Howie and Barss (1975b).

concept (Plate 2).

It is uncertain whether the Aquitaine–Cantabrian and the Intra-Alpine terranes formed part of a single microcraton or whether they represented separate blocks prior to their accretion to the southern margin of Fennosarmatia.

During the Acadian–Ligerian diastrophism, compressive stresses temporarily overcame back-arc extension in the Central Armorican and the Saxothuringian basins. This was apparently accompanied by the first phases of closure of the southwest Cornwall back-arc oceanic basin (Rolet et al., 1986; Holder and Leveridge, 1986). On the other hand, the Middle Devonian evolution of the Rhenohercynian Basin continued to be governed by crustal extension (Ziegler, 1982a). This suggests that during the Acadian–Ligerian orogeny both the subducting Proto-Tethys and the overriding Fennosarmatian plate were partly coupled at the B-subduction zone that paralleled the southern margin of the Ligerian–Moldanubian Cordillera.

Furthermore, it is likely that during the Acadian–Ligerian diastrophism a new B-subduction zone developed along the southern margin of the Intra-Alpine Block. It is, however, uncertain whether this subduction zone linked up to the west with the post-Acadian Proto-Tethys subduction zone along the southeastern margin of the Avalon-Meguma-South Portuguese Terrane (see the beginning of this section, *Acadian–Ligerian and Bretonian orogenies*).

Following the Acadian–Ligerian diastrophism, back-arc extension resumed to control the subsidence of the Central Armorican and the Saxothuringian basins as evident by the resumption of alkaline-bimodal volcanic activity (Sider and Ohnenstetter, 1986). Furthermore, there is evidence that southern Iberia was also affected by back-arc extension causing the subsidence of the South Portuguese Basin and inducing an alkaline-bimodal volcanism (Oliveira, 1982). Whether back-arc extension also affected the Intra-Alpine domain has yet to be resolved.

The resumption of regional back-arc extension during the Late Devonian probably reflects a renewed decrease in the convergence rate (or even a gentle divergence) between the Proto-Tethys plate and the Fennosarmatian subplate, a commensurate steepening of the Proto-Tethys B-subduction zone, and at least a partial decoupling of the subducting and overriding plates. This interpretation is in keeping with the postulated Middle Devonian to Early Carboniferous sinistral translation of Fennosarmatia and Laurentia–Greenland (see *Arctic–North Atlantic Megashear*, this chapter), during which Fennosarmatia may actually have temporarily receded from the Proto-Tethys subduction zone. There is, however, still considerable uncertainty about the width of Proto-Tethys during the Late Devonian, mainly due to insufficient paleomagnetic data from Gondwana in general and specifically from Africa (Kent et al., 1984).

Bretonian Orogeny

The Bretonian orogenic pulse, straddling the Devonian–Carboniferous boundary, can be related to a second phase of increased convergence between the Fennosarmatia and the Proto-Tethys plates. Renewed shallowing of the Proto-Tethys B-subduction zone and partial coupling of the subducting and the overriding plates resulted, once more, in the exertion of compressive stresses on the back-arc areas. Again, mainly the Ligerian–Moldanubian Cordillera and the southern parts of the Central Armorican (Rolet, 1983; Rolet et al., 1986) and Saxothuringian basins (Behr et al., 1982) were affected by compressive stresses, while back-arc extension, giving rise to a commensurate volcanism, persisted in the Rhenohercynian Basin (Plates 3, 24, 25). The Famennian increase in clastic influx into this basin from southern sources indicates, however, that the Mid-German High also became reactivated during the Bretonian orogeny (Engel and Franke, 1983; Behr et al. 1984; Weber, 1984; Franke and Engel, 1986). Furthermore, there is evidence for continued closing of the south Cornwall back-arc oceanic basin (Holder and Leveridge, 1986).

In Iberia, evidence for a Bretonian orogenic phase is largely restricted to Galicia (Martinez-Garcia, 1972; Julivert, 1979, 1983). Contemporaneous granitoids are, however, lacking (Priem and den Tex, 1984). In the South Portuguese–Ossa Morena Basin, back-arc extension apparently persisted across the Devonian–Carboniferous boundary (Oliveira, 1982). In northern Algeria there is, however, evidence for Bretonian orogenic deformations (Bouillin and Perret, 1982), and it is therefore speculated that this area may have formed part of the arc system along which oceanic crust of the Proto-Tethys continued to be subducted.

Limited radiometric age determinations indicate that the Intra-Alpine domain was also affected by the Bretonian orogeny (Schönlaub, 1979; Trümpy, 1980; Oberli et al., 1981; Thélin and Ayrton, 1983). This, however, requires further clarification.

The stratigraphic record of the Gondwana-derived Austro-Alpine Block (Upper Austro-Alpine nappes) lacks evidence of a compressive Bretonian deformation phase. The same applies for the South Alpine–Carnic–Dinarid domain, which is also characterized by a nearly complete Gondwana-type Paleozoic sedimentary sequence.* Furthermore, a Devonian separation between the Austro-Alpine and the Carnic–Dinarid blocks is indicated by faunal differences (Schönlaub, 1979; Vai, 1975, 1980; Vai and Cocozza, 1986).

This suggests that the Austro-Alpine block constitutes a separate terrane. Whether the Carnic–Dinarid block was still attached to the northern margin of Gondwana during the Devonian and Early Carboniferous is uncertain. Its tensional evolution, which persisted into the Westphalian, is thought to be related to large-scale dextral strike-slip movements preceding and accompanying the Carboniferous convergence of Gondwana and Laurasia (Vai, 1975, 1980; Vai and Spalletta, 1982; Vai and Cocozza, 1986). Paleomagnetic data may shed further light on the Paleozoic evolution of the Austro-Alpine and the Carnic–Dinarid blocks.

Following the Bretonian orogenic pulse, back-arc extensions resumed and dominated the Tournaisian and early Visean evolution of the Central Armorican, the Saxothuringian, and the Rhenohercynian basins and possibly also of the Intra-Alpine domain.

This may reflect a renewed decrease in the convergence rate of the Proto-Tethys–Gondwana plate and the Fennosarmatia subplate. On the other hand, paleomagnetic data suggest that by early Visean time the collision of Gondwana and Laurasia

*Gondwana-type Paleozoic sequences are characterized by Ordovician cold-water faunas and the lack of Caledonian compressional deformations. Sedimentation took place in a tensional setting that variably ranged from the Ordovician to the Silurian, Devonian, and Early Carboniferous. The widespread Ordovician magmatism (Caledonian thermal event) is of an essentially anorogenic nature and is probably related to regional tensional tectonics that affected the northern margin of Gondwana; these induced the separation of a number of microcontinents from the northern margin of the Afro-Arabian craton.

was imminent (Jones et al., 1979; Perroud and Bonhommet, 1981; van der Voo and Scotese, 1981; Kent, 1982; van der Voo, 1983). It can, however, not be excluded that already during the Bretonian orogeny cratonic promontories of Gondwana had already collided with the southern margin of Fennosarmatia (see Matte, 1986). This is possibly indicated by the latest Devonian-Early Carboniferous evolution of Northwest Africa where wrench faulting induced the Famennian-Dinantian subsidence of the Sidi Betache Basin in northwest Morocco (Piqué, 1981; Piqué and Kharbouche, 1983) and the concomitant differential subsidence of the Ougarta Trough that extends from the Anti-Atlas in a southeasterly direction into the Sahara Platform (Wendt, 1985; Fig. 8).

From the above it follows that the "Avalon plate," defined by Rast and Skehan (1983) as including, apart from the Avalon–Meguma Terrane, the London–Brabant Massif, the Central Armorican, and the Aquitaine–Cantabrian blocks as well, was not accreted to the southern margin of Laurussia as a single tectonic element but consisted of a mosaic of separate continental fragments that collided stepwise with the Proto-Tethys subduction zone(s). For instance, the London–Brabant, Armorican, and Saxothuringian microcratons were accreted to Laurussia during the Caledonian orogeny whereas the Avalon–Meguma, Aquitaine–Cantabrian, and possibly the Barrandian and Intra-Alpine microcratons were accreted during the Acadian–Ligerian orogeny.

The exact timing of docking and accretion of these different continental fragments to the southern margin of Laurussia is, however, difficult to determine, mainly due to still fragmentary information. Moreover, it should be kept in mind that the definition of the individual suspected allochthonous terranes contained in the European part of the Hercynian fold belt is severely hampered by their intense deformation during the Variscan and partly also the Alpine diastrophism, by limited exposure of the Variscan basement complex, and by incomplete data sets. It is likely that their number is actually greater and their configuration and lateral relationship is more complex than suggested by Plates 2–4. As results of ongoing research become available, the conceptual model developed above will have to be modified and in all likelihood will become considerably more complex.

ARCTIC-NORTH ATLANTIC MEGASHEAR

As discussed in Chapter 1 and illustrated in Plate 1, Fennoscandia and Laurentia–Greenland were dextrally offset during the latest Silurian by a minimum of 1500 km as compared to the Permo-Triassic Bullard Fit. The bulk of this offset was recovered during the Middle and Late Devonian by sinistral motions along a complex fault system, transecting the axis of the Arctic North Atlantic Caledonides. Relatively minor displacements may still have occurred during the Early Carboniferous (Plates 2–4).

This postulate is supported by modern paleomagnetic data (Irving and Strong, 1984) and is in keeping with the stratigraphic record of the North Atlantic borderlands and those of the Norwegian–Greenland Sea. These data refute the earlier held view (Kent and Opdyke 1978, 1979) that most of these movements had taken place during the Carboniferous (Kent and Opdyke, 1978; Keppie et al., 1985; Kent and Keppie, 1988). Further paleomagnetic support for the Devonian Arctic–North Atlantic translation comes from Scotland where Storedvedt (1987) recognizes a 400–600 km Middle to Late Devonian sinistral displacement along the Great Glen fault.

Definition of the fault systems pertaining to the Arctic–North Atlantic megashear is limited to onshore areas since thick Mesozoic and Cenozoic strata generally mask the Devonian and Carboniferous structural framework of the shelves flanking the North Atlantic and the Norwegian–Greenland Sea. Fault patterns shown on Plates 2–4 are therefore schematic and in offshore areas conceptual.

The stratigraphic record of Devonian and Carboniferous basins associated with the Arctic–North Atlantic megashear is summarized in Plate 22, for which an index is given in Fig. 9.

Canadian Maritime Provinces

In the Canadian Maritime Provinces, the present juxtaposition of the western and eastern parts of the Acadian fold belt was achieved during the late Visean to Namurian. The bulk of these sinistral displacements was taken up along the Lubec–Belle Isle–Cobequid–Cabot fault system transecting the Fundy Basin (Fundy Synclinorium or Magdalen Basin). The evolution of this complex basin was accompanied by wrench and pull-apart movements and repeated volcanic activity, spanning late Middle Devonian to early Namurian time (Fig. 3; Howie and Barss, 1975a, 1975b; Bradley, 1982; Fyffe and Barr, 1986). Data from Newfoundland indicate that intra-Carboniferous displacements were, however, below the paleomagnetic margin of error (Irving and Strong, 1984). Contemporaneous movements along the Minas (Chedabucto) geofracture, controlling the evolution of the Sidney Basin, are thought to have been dextral (Keppie, 1985). In northeastern Newfoundland, the Cabot fault splits up into the fault systems controlling the subsidence of the St. Anthony Basin (Cutt and Laving, 1977) and those associated with the Devono-Carboniferous basins of the northern British Isles. An important element of this fault system is the Great Glen Fault. In view of the paleomagnetic constraints on the movements along this fault (Storedvedt, 1987), it is likely that the bulk of the Devonian shear movements was taken up along a fault system located to the west of the Hebrides Isles.

The sedimentary record of the Fundy Basin is summarized in Plate 22. In the Post-Acadian basins of the Canadian Maritime Provinces, late Middle Devonian to Tournaisian continental clastics, attaining a thickness of up to 4000 m, are overlain by Visean evaporitic series containing minor carbonates and clastics (Howie and Barss, 1975a, 1975b; Boehner, 1983). Corresponding marine ingressions reached the Sidney and Fundy basins presumably from the West-Iberian segment of the Rhenohercynian Basin and the St. Anthony Basin from the grabens and troughs of southern Ireland (Plates 3, 4).

Northern British Isles

In the northern British Isles the late Ludlovian to early Emsian lower Old Red continental and lacustrine series accumulated in the largely tensional intramontane Orcadian Basin, the Midland Valley, and the Northumberland–Dublin Trough (Plates 22, 23; Figs. 3, 11). Their development was accompanied by an extensive postorogenic intrusive and extrusive igneous activity. Lower Old Red clastics reach a maximum thickness of 6600 m in the Midland Valley Graben (Leeder, 1976; House et al., 1977; Bluck, 1978, 1984).

During the Middle Devonian, sedimentation in the Northumberland–Dublin Trough and the Midland Valley Graben was interrupted by transpressional deformations giving rise to low relief folds in the Lower Old Red sandstone. In the Orcadian Basin, on the other hand, unconformity-bound Middle Devonian Old Red Series consist of some 5000 m of lacustrine shales and fluviatile sands. Their accumulation was accompanied by syndepositional deformations reflecting the interplay between transtensional and transpressional stresses that can be related to sinistral movements along the Great Glen Fault (House et al., 1977; Anderton et al., 1979; Watson, 1985).

The Middle Devonian Orcadian Basin probably extended across the northern North Sea to the Norwegian coast where great thicknesses of Old Red conglomerates and sand accumulated in the transtensional Hornelen, Solund and two other smaller basins (Nilsen, 1973; Steel, 1976; Steel and Gloppen, 1980; Ziegler, 1982a). The margins of these basins are partly marked by basement-involving overthrusts (Roberts, 1983). These Middle Devonian deformations, which coincide with the Acadian–Ligerian orogeny, can be related to the Arctic–North Atlantic translation. In onshore and offshore areas of northwest Scotland, the subsidence of half-graben-shaped Old Red Sandstone basins involved the tensional reactivation of Caledonian thrust faults. Some of these basins are associated with flat-lying mylonite zones and thus beg a comparison with the Basin and Range Province of the United States Cordillera. This illustrates that the Early to Middle Devonian evolution of the northern North Sea area reflects the interplay between transtensional and transpressional tectonics (Hossack, 1984; Beach, 1985; McClay et al., 1986; Enfield and Coward, 1987).

During the Late Devonian tectonic activity abated in the British Isles and sedimentation resumed in most of its Old Red Sandstone basins. As a result of the progressive overstepping of basin margins, which is partly related to eustatically rising sea levels (House, 1983), these basins became connected with each other and with the Rhenohercynian Basin via southern Ireland and the North Sea. Through these avenues, transgressions advanced from the latter during the early Tournasian. Dinantian and Namurian strata are represented by carbonates, mixed carbonate and clastics and deltaic-paralic series (Plates 3, 4; Ziegler, 1982a).

The Dinantian and early Late Carboniferous evolution of the sedimentary basins of the British Isles (Midland Valley, Northumberland, Solvay, Craven, and Dublin troughs) was governed by crustal extension as documented by the widespread occurrence of an alkaline, bimodal rift volcanism (Francis, 1978; Leeder, 1982; Ziegler, 1982a; Kirton, 1984). During the Early Carboniferous, these basins were connected to the southwest with the basins of the Canadian Maritime Provinces. Together they formed an important part of the Arctic–North Atlantic rift-wrench system (Ziegler, 1984, 1986).

Norwegian–Greenland Sea Area

Little is known about the Devonian evolution of the West Norwegian coastal and shelf areas. In the Trondheimsfjord, some 1300 m thick Early to Middle Devonian conglomeratic series are preserved in a narrow trough, the southeastern margin of which is overthrust by the Caledonian basement (Siedlicka and Siedlicki, 1972; Nilsen, 1973; Roberts, 1983; Norton, 1986). The age of deformation is unknown but probably Middle Devonian. In the adjacent offshore areas, reflection seismic data indicate the presence of thick pre-Permian sediments, contained in a half-graben, that may include Devonian and Early Carboniferous strata (Figs. 37, 38; Bukovics and Ziegler, 1985).

In Central East Greenland, some 7000 m thick Emsian to early Tournaisian Old Red series, consisting of conglomerates, sandstones, and lacustrine shales, accumulated in the Trail Ø-Hudson Land area (Plates 22; Fig. 9). The evolution of this basin was accompanied by repeated wrench deformations, referred to as the Hudson Land phases, volcanism, and granitic intrusions. The last phase of deformation, involving thrust faulting, is referred to as the Ymerland phase, which cannot be dated closer than intra-Dinantian (Haller, 1971; Friend et al., 1976). From the structural style of this basin it is evident that its development was governed by transtensional and transpressional deformations (Fig. 6). These can be related to the Arctic–North Atlantic megashear. Early Namurian to Early Permian continental clastics, however, were deposited under a tensional setting that persisted through Mesozoic time (Vischer, 1943; Haller, 1971; Pedersen, 1976). The Permo-Carboniferous East Greenland rift system can be followed over a distance of some 1400 km from Scoresby Sound to the Wandel Sea Basin in northeastern Greenland. In the latter, Dinantian coalbearing shales and sands, possibly accumulating in fault-controlled basins, are disconformably overlain by Westphalian–Stephanian marine sands and carbonates (Håkansson et al., 1981b; Håkansson and Stemmerik, 1984; Plate 22).

Svalbard and Barents Shelf

In the Svalbard archipelago, latest Silurian to earliest Devonian sediments are restricted to the Central Spitsbergen Billefjord Trough where 1500–2000 m thick Downtonian red beds were deposited on a late Caledonian basement complex (Plate 22). These strata were folded and thrusted during the early Gedinnian Haakonian deformation phase. Sedimentation resumed during the late Gedinnian with the accumulation of continental red beds; these grade upward into Middle Devonian marginally marine clastics. These strata, which attain a maximum thickness of 6000 m, were shed from southwestern and southeastern sources. Similar series may be present beneath the Permo-Carboniferous and Mesozoic cover of eastern Svalbard and the Svalbard Bank (Birkenmajer, 1981; Rønnevik and Beskow, 1983). During the Late Devonian, sedimentation was interrupted by the Svalbardian wrench deformations, which probably contributed significantly to the assembly of the western, central, and eastern parts of Svalbard to its present configuration (Harland et al., 1984; Plate 3). The accumulation of a continental coal-bearing series resumed during the latest Devonian under a transtensional setting (Steel and Worsley, 1984).

At the transition from the Visean to the Namurian, wrench movements, referred to as the Adriabukta phase, gave rise locally to important deformations and low-grade metamorphism (Birkenmajer, 1981). In the course of the Namurian, the area of sedimentation gradually increased while subsidence rates of the main grabens slowed down. A last phase of significant wrench deformations, causing partial basin inversion, occurred during the late Namurian (Worsley and Edwards, 1976; Gjelberg and Steel, 1981; Steel and Worsley, 1984).

For Bjørnøya a similar latest Devonian to Carboniferous basin evolution is indicated by Gjelberg and Steel (1983) (Plates 3, 4, 22).

From the above it can be concluded that wrench deformations related to the Arctic–North Atlantic megashear probably terminated during the latest Visean and that post-Visean deformations in the Svalbard area may have been induced by transform movements compensating for crustal extension in the Norwegian–Greenland Sea Rift.

The Devonian and Carboniferous evolutions of the Western Barents Sea has to be derived largely from reflection seismic data that are only partly calibrated by well data. Thus considerable uncertainty exists in the identification of the different Paleozoic seismic reflectors. Yet, these data suggest that Middle(?) to Late Devonian and Early Carboniferous rifting in the southwestern Barents Sea induced the development of a northeast–southwest-trending fault pattern controlling the subsidence of the Nordkapp Basin and possibly also of the Tromsø and Bjørnøya basins (Fig. 10). On the Svalbard Bank, northwest–southeast-striking dextral wrench faults are evident and are interpreted as a conjugate shear to the Arctic–North Atlantic megashear (Rønnevik and Beskow, 1983; Rønnevik and Jacobsen, 1984; Gabrielsen et al., 1984).

During the Devonian and Carboniferous, marine transgressions originating from the eastern Barents Sea advanced westward as illustrated by marginally marine Middle Devonian strata in central Spitsbergen and Bashkirian marine influences in Svalbard and Bjørnoya. The occurrence of suspected Early Carboniferous sandy carbonates in Andøya (West Norway) may be indicative of a major transgression (Sturt et al., 1979). Devonian and Early Carboniferous facies patterns as shown in Plates 2–4 are, however, poorly controlled, and it is uncertain whether evaporites accumulated during the Late Devonian in the Tromsø and Northkapp basins as suggested by Rønnevik and Beskow (1983) and Faleide et al. (1984).

In the coastal area of northernmost Norway, the east-west-trending dextral Trollfjord–Komagelv wrench fault was active during the Devonian and Early Carboniferous (Fig. 10; Johnson et al., 1978). Dyke intrusions associated with this fault are dated as f 355 Ma (Beckinsale et al., 1975).

Conclusions

In summary, there is ample geological evidence in support of the postulated Devonian to Early Carboniferous Arctic–North Atlantic sinistral megashear. The bulk of these movements occurred apparently during the Middle and Late Devonian with Early Carboniferous displacements being on a relatively minor scale. On the other hand, there is still considerable controversy about the paleomagnetic evidence supporting this concept, as well as about the timing and the magnitude of these movements (e.g., Esang and Piper, 1984; Scotese et al., 1984; Keppie et al., 1985).

INNUITIAN FOLD BELT

The late Paleozoic stratigraphic record of the Sverdrup Basin, Northern Ellesmere Island and of Northern Greenland, as summarized in Plate 22, reflects the evolution of the Innuitian fold belt.

In northern Greenland, the late Caledonian orogeny resulted in the emplacement of thin-skinned thrust sheets. This was accompanied by the deposition of some 3000 m of synorogenic Silurian flysch series. The youngest sediments involved in these thrust sheets are of Gedinnian age. Although younger Devonian strata are not preserved in the area, last thrust movements occurred probably during the Late Silurian–Early Devonian Vølvedale orogeny (Dawes and Peel, 1981; Håkansson and Pedersen, 1982; Pedersen, 1986).

Time-equivalent deformations are evident in northern Ellesmere Island where Early Devonian shallow marine clastics form part of a neoautochthonous sequence (Mayr, personal communication). In the Innuitian foredeep, the Franklinian Basin, synorogenic flysch series mainly derived from eastern sources continued to accumulate during the Early Devonian.

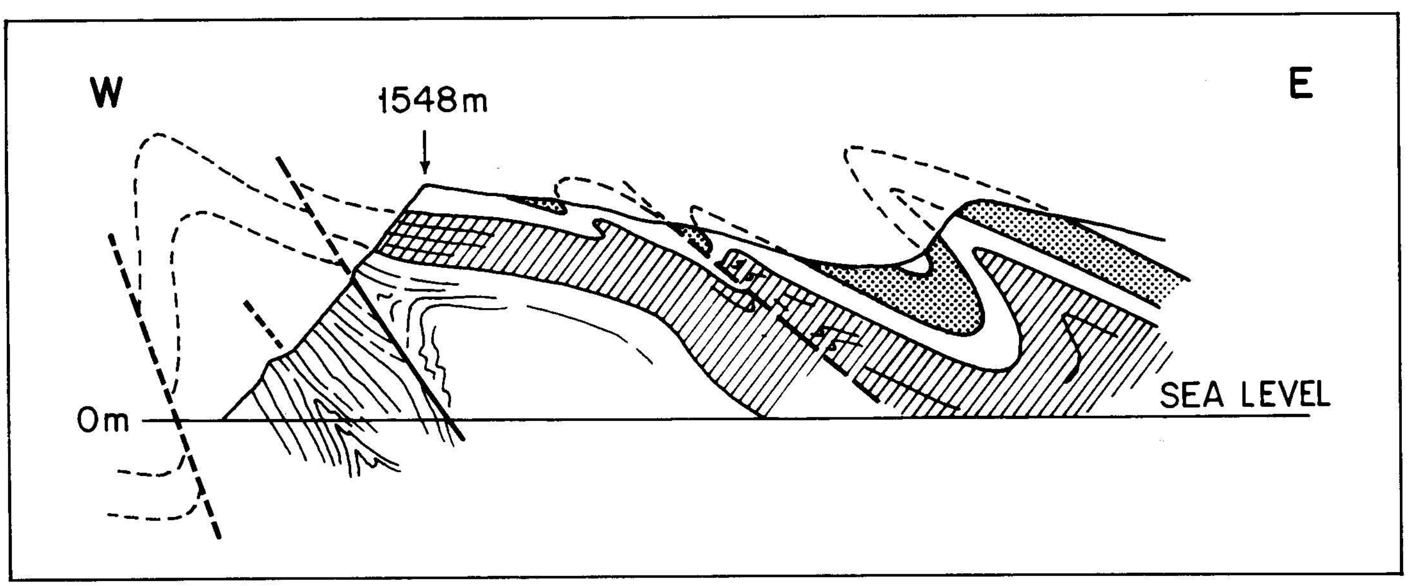

Figure 6—Structural cross section through Svedenborg Bjerg, Kong Oscar Fjord, Central East Greenland, showing Mid-Devonian Old Red clastics intensely folded and thrusted by Early Carboniferous movements. Modified after Bütler, in Haller (1971).

Carbonate shelves, occupying the distal southern margin of this foredeep basin, persisted into the Emsian and early Eifelian, by which time they became buried beneath deltaic, and later continental, clastics prograding from the northeast and east (Plates 2, 3). These molasse series, which extend into the Frasnian, were deposited during the Ellesmerian orogenic cycle terminating in the Early Carboniferous with the folding of the Franklinian Basin and the consolidation of the thin-skinned Parry Island fold belt that forms the externides of the Innuitian orogen (Plate 4; Fig. 9). Granitic intrusives in the northern part of the Innuitian fold belt are dated as 360 ± 25 Ma and 345 ± 15 Ma (Trettin and Balkwill, 1979; Kerr, 1981a; Balkwill and Fox, 1982; Smith and Stearn, 1982; Fox, 1985).

A palinspastic restoration of the Arctic area, prior to the opening of the Canada Basin, suggests that the direct western extension of the Innuitian fold belt is formed by the Alaska North Slope and the area of the Brooks Ranges and British Mountains of Ellesmerian deformation from where granitic intrusives ranging in age from 430 to 330 Ma have been reported (Fig. 7; Ziegler, 1969; Bird et al., 1978; Metz et al., 1982; Dutro, 1981; Hubbard et al., 1987). In the northernmost Yukon, the Innuitian deformation front is evident in the northern Richardson Mountains from where it trends in a southwestern direction (Bell, 1973).

The folds and thrusts of the Innuitian orogenic belt are clearly southeast vergent (Kerr, 1981a). South-vergent folds and thrusts of the north Greenland belt, which in part overprint the earlier west-vergent Vølvedal structures, are also attributed to the Ellesmerian orogeny. Because it is impossible to stratigraphically date these deformations more accurately than post-Late

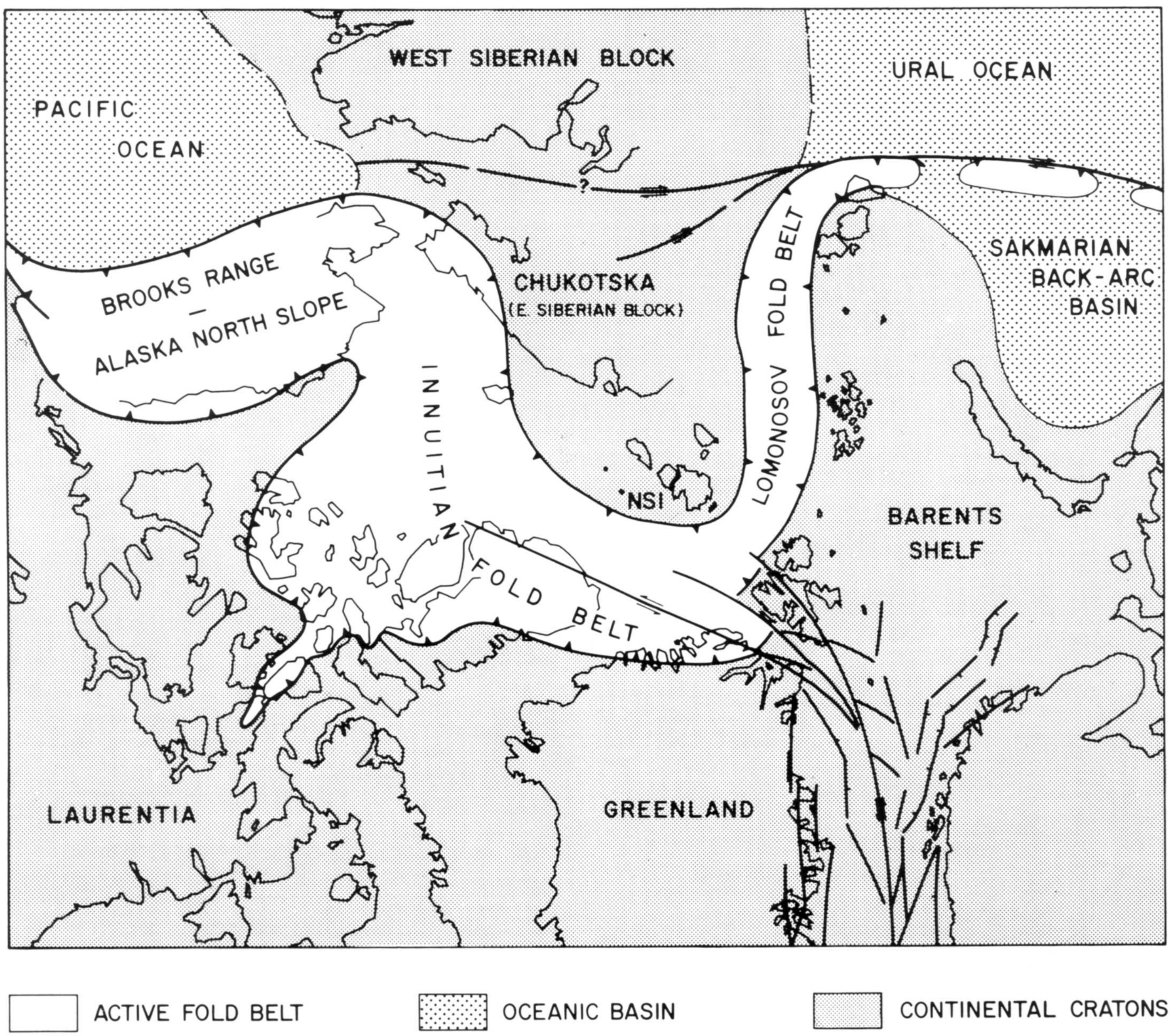

Figure 7—Conceptual Early Carboniferous tectonic framework of the Arctic. NSI = New Siberian Islands.

Silurian, they may actually be of an Early to Middle Devonian age (Dawes and Peel, 1981; Pedersen, 1986). This would be in keeping with the Emsian influx of deltaic clastics into the Innuitian foredeep.

In northern Ellesmere Island, sinistral wrench deformations are thought to have induced the Early to Middle Devonian intrusion of granodiorites and basic plutons and the development of intramontane basins in which thick Late Devonian and also Early Carboniferous continental clastics accumulated (Trettin and Balkwill, 1979; Trettin, 1987; Trettin et al., 1987; Mayr, personal communication). Also in the Brooks Range of Northern Alaska, synorogenic Late Devonian clastics accumulated in intramontane basins (Moore and Nilsen, 1984; Hubbard et al., 1987). In the North Greenland fold belt, wrench deformations probably induced the intrusion of the post-orogenic Middle Devonian Midtkap igneous complex and a low-grade metamorphism straddling the Early to Late Carboniferous boundary (Håkansson and Pedersen, 1982; Pederson and Holm, 1983). The latter is coeval with the the Adriabukta deformation phase in Svalbard that gave rise to local metamorphism (Birkenmajer, 1981). This suggests that the Late Devonian to Early Carboniferous consolidation of the Innuitian fold belt was accompanied by major sinistral wrench deformations related to the Arctic–North Atlantic megashear.

The Innuitian fold belt presumably forms the megasuture between the Siberian Block and Laurentia (Fig. 7; see also Scotese et al., 1979). It is, however, uncertain whether the collision and suturing of these cratons was governed by a south-plunging subduction system located along the northwestern margin of Laurentia–Greenland or by a north-plunging subduction system paralleling the leading edge of the Siberian block (Trettin and Balkwill, 1979). The rather moderate expression of a Late Devonian to Early Carboniferous calc-alkaline magmatism in northern Alaska and Ellesmere and Axel Heiberg islands, as well as their absence in the North Greenland fold belt, cannot be considered as an argument for or against either of the two hypotheses. Yet, the Late Carboniferous development of the Sverdrup Basin, which may be related to back-arc rifting, would be more easily explained if the first alternative would apply (see Chapter 4).

Assuming that the Innuitian fold belt indeed represents the megasuture between Laurentia–Greenland and the Siberian Craton, and accepting the validity of the Devono-Carboniferous Arctic–North Atlantic megashear, and taking into account the indicated contemporaneous wrench deformations in the internal parts of the Innuitian and North Greenland fold belts, it is likely that the northern margin of the Barents Shelf collided during the Late Devonian to Early Carboniferous with the southeastern margin of the Siberian Craton. Unfortunately, critical areas, including the Lomonosov Ridge, are not accessible to geological investigation. On Franz Josef Land Early(?) Carboniferous paralic clastics overlay late Precambrian (Wendian) metamorphics. On the North-Islands (Severnaya Zemya), Old Red-type Late Devonian sandstones overlay Siluro-Ordovician marine platform series. The latter appear to have been affected by folding that is suggested to be of a "Caledonian" age, and there is apparently evidence for post-Late Devonian folding. Whether these late deformations can be attributed to the Ellesmerian orogeny or to the Permian Uralian orogeny cannot be resolved on the basis of the available literature (Churkin et al., 1981; Karasik et al., 1984). "Caledonian" deformations are furthermore reported from the coastal areas of Taimyr Peninsula

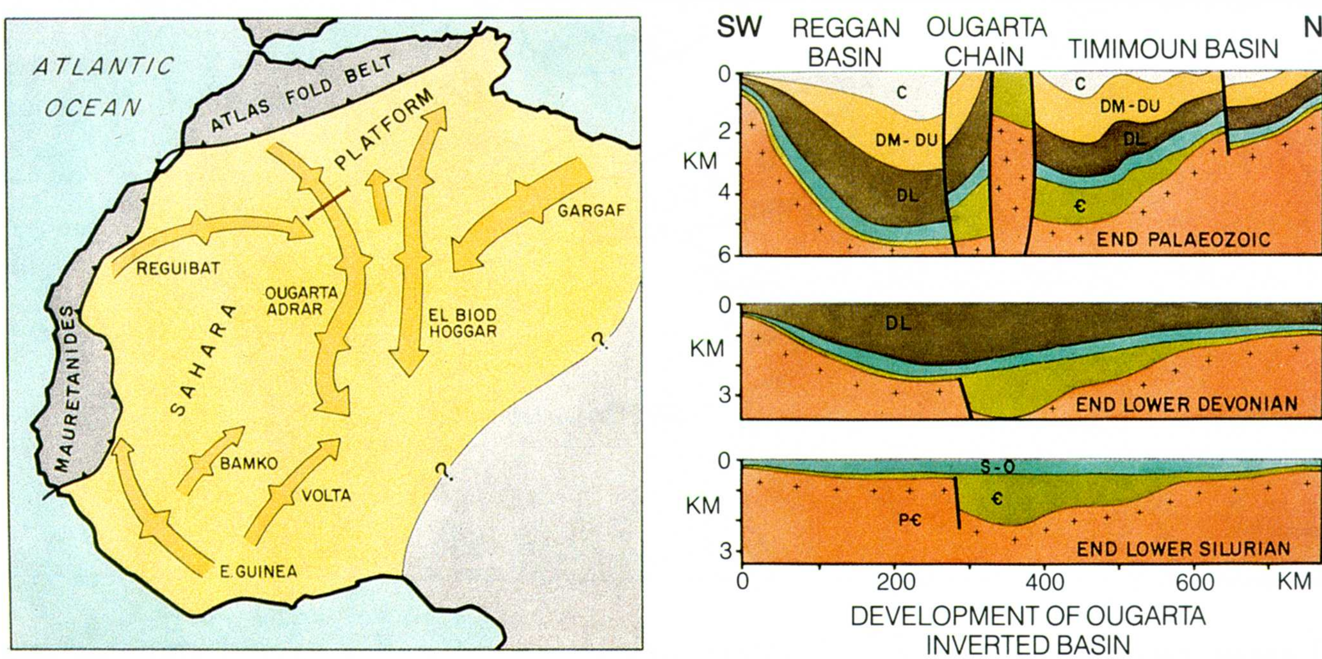

Figure 8—Carboniferous intraplate deformation of the Sahara Platform and evolutionary diagram of the Ougarta Trough. Arrows indicate axes of major arches.

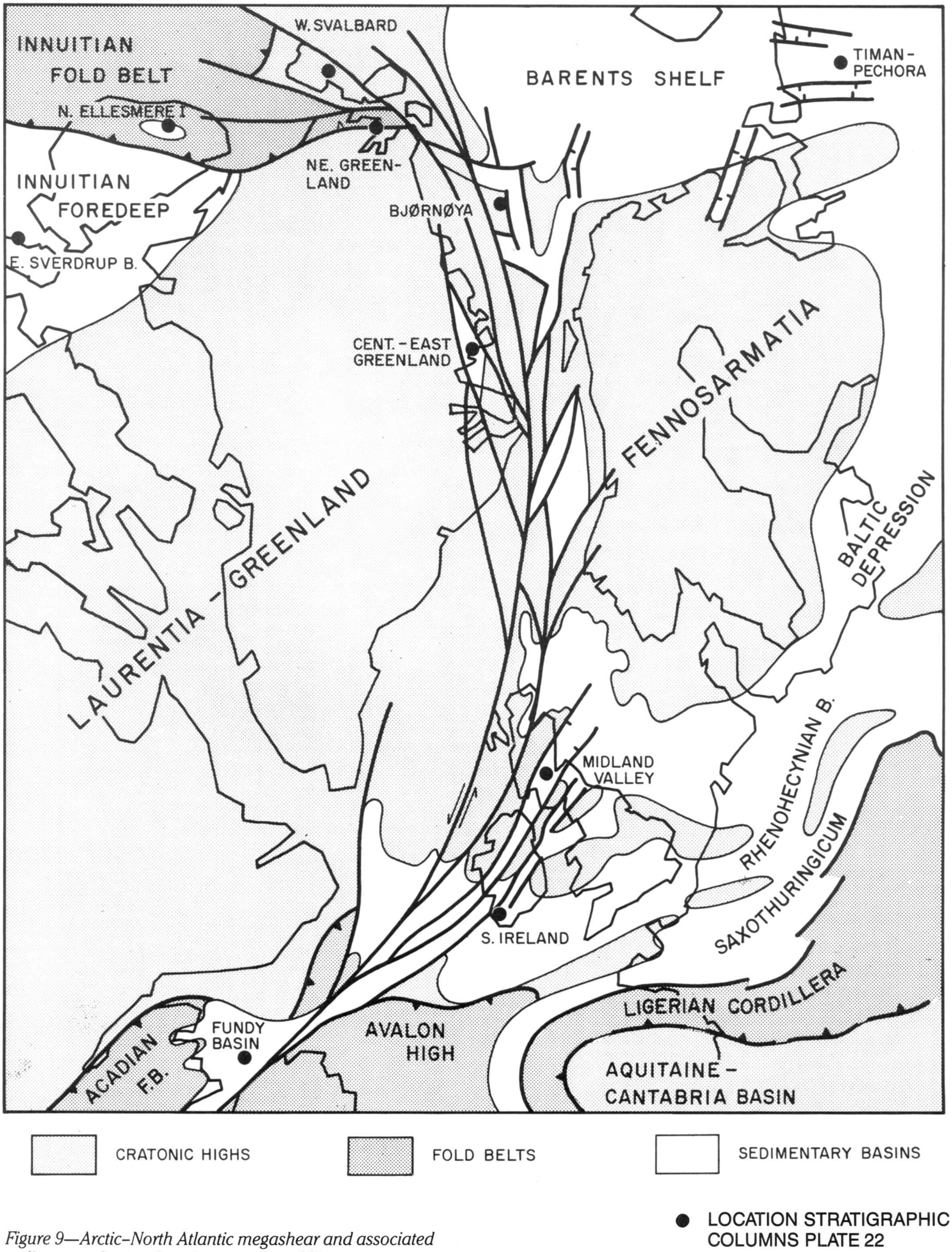

Figure 9—Arctic–North Atlantic megashear and associated sedimentary basins (location map Plate 22).

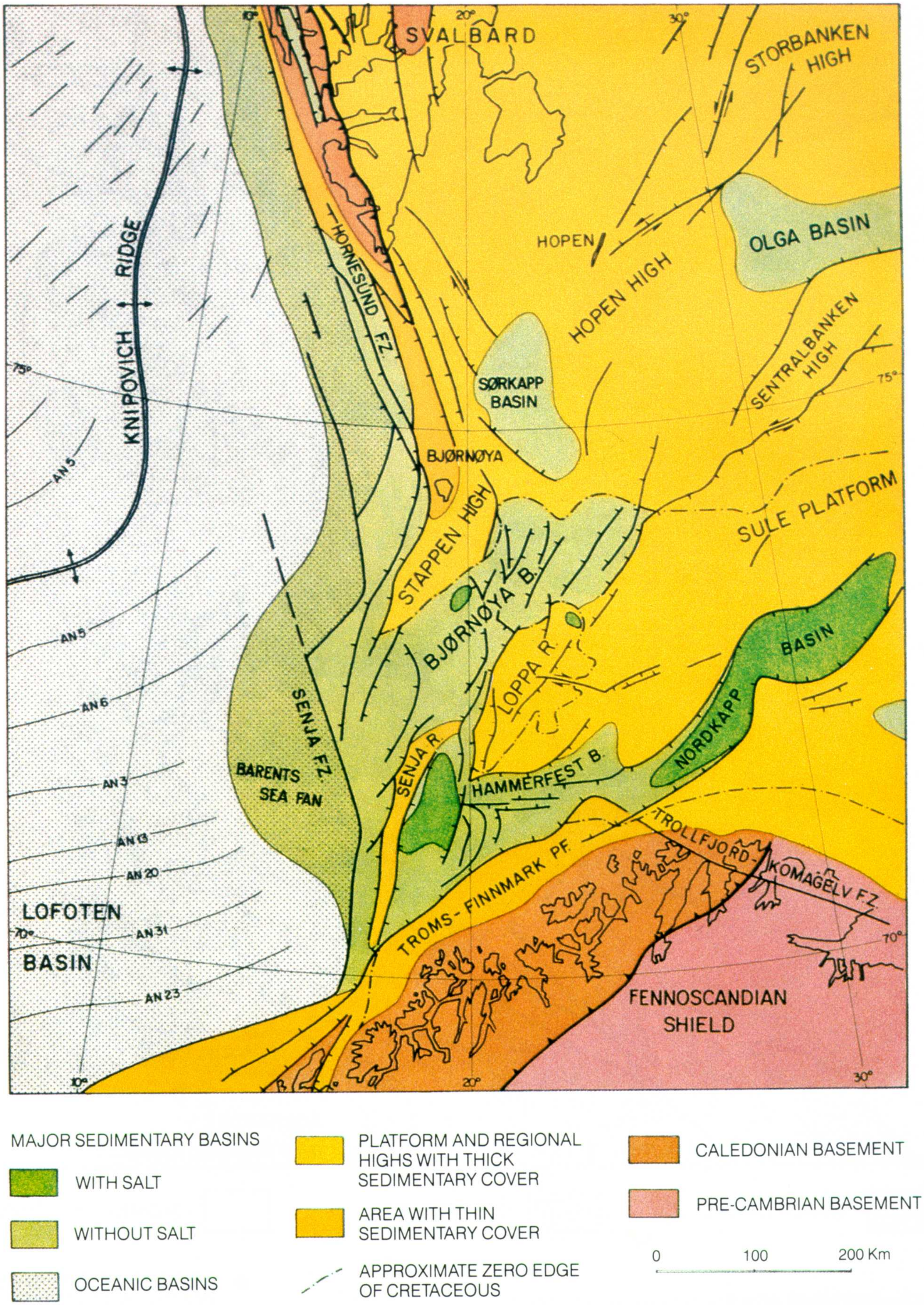

Figure 10—Tectonic map of the Western Barents Shelf. After Norske Shell.

(Hamilton, 1970; Churkin and Trexler, 1981; Pogrebitsky, 1982).

In short, the hypothesis that the Innuitian fold belt finds its eastward extension in the Lomonosov fold belt remains a postulate that begs further analysis.

EASTERN MARGIN OF FENNOSARMATIA AND THE PRE-URALIAN ARC-TRENCH SYSTEM

Following the earliest Devonian eustatic low stand of sea level, sea levels rose cyclically during the Devonian and Early Carboniferous (Vail et al., 1977; House, 1983; Johnson et al., 1985).

This is reflected on the stable eastern shelves of Fennosarmatia by cyclical late Middle and Late Devonian transgressions inducing the development of vast carbonate, carbonate-evaporite, and mixed carbonate-clastic platforms (Plates 2, 3). These encroached progressively onto the Fennoscandian-Baltic Shield. Similar carbonate platforms characterized the Early Carboniferous setting of the Moscow Platform and probably also of the eastern parts of the Barents Shelf (Vinogradov, 1969).

During the late Eifelian to early Dinantian, the Kola Peninsula, the Timan–Pechora area, and the northern parts of the Moscow Platform became affected by tensional tectonics, giving rise to the differential subsidence of the Kontozero, Pechora, and Kolva grabens and of the Vyatka rift. In the Timan–Pechora area and on the Kola Peninsula, the development of these grabens was accompanied by an alkaline volcanism (Tszyn, 1967; Churkin et al., 1981; Ulmishek, 1982; Gortunov et al., 1984; Plate 22). Whether similar grabens occur in the eastern part of the Barents Shelf is unknown. Moreover, the nature of the geodynamic processes that governed the development of these rifts is not clear. Could they be related to the Arctic–North Atlantic shear via, for instance, the Trollfjord–Kolmagelv fault, or is their development linked with the evolution of the Sakmarian–Magnitogrosk arc-trench and back-arc extension system?

According to Zonenshain et al. (1984), the late Early Devonian to Eifelian reactivation of subduction processes along the Sakmarian–Magnitogorsk arc-trench system was accompanied by the submarine obduction of oceanic crust onto the lower parts of the Fennosarmatian margin (Ruzencev and Samygin, 1979). Emplacement of these nappes was, however, not associated with intense subsidence of the adjacent shelf area (Artyushkov and Baer, 1983). Back-arc compression ceased and gave way to back-arc extension during the late Eifelian, and new oceanic crust was formed during the Givetian, at least in the southern parts of the Sakmarian Basin. The beginning of back-arc extension in the Sakmarian Basin coincided with the onset of rifting in the Timan–Pechora area.

Continued convergence between the Ural oceanic plate and the Sakmarian Magnitogorsk arc-trench system was accompanied by the Famennian collision of the Mugodjarian microcontinent with the latter (Plate 4). Famennian to Tournaisian back-thrusting of the arc induced its uplift and the shedding of clastics into the back-arc basin, where they were deposited as deep sea fans on oceanic crust. These clastics reached the toe of

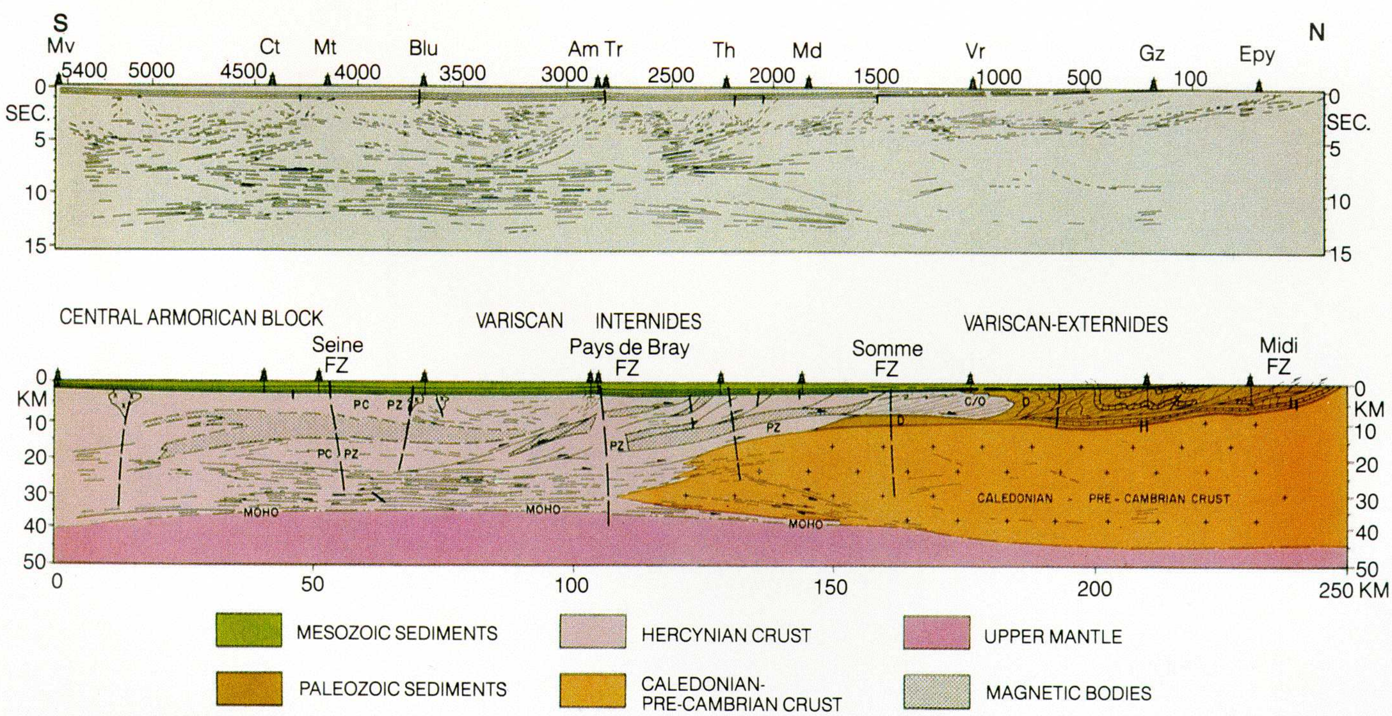

Figure 11—Crustal structure of the Paris Basin. After Cazes et al. (1985).

the Fennosarmatian passive margin. During the Tournaisan and early Visean, the Sakmarian–Magnitogorsk arc was characterized by a synorogenic acidic volcanism, which was followed by the middle Visean intrusions of granitic plutons (Plate 4). Subduction processes apparently came to a halt during the middle Visean (Zonenshain et al., 1984).

The evolution of the Pechora and Kolva rifts, on the other hand, suggests that back-arc extension may have persisted into the Tournaisian (Plate 22).

The subsequent evolution of the Uralian orogenic system was governed by a new B-subduction system that developed along the leading edges of the Kazakhstan and the Siberian cratons. The latter apparently became separated from the northern margin of Laurentia (Alaska–Chuckhi block) during the late Visean–early Namurian and began to converge with the now defunct Sakmarian–Magnitogorsk arc and the eastern margin of Fennosarmatia. This marked the onset of the actual Uralian orogenic cycle.

It is unknown how wide the oceanic Sakmarian back-arc basin was at the beginning of the Late Carboniferous. In view of this, the situation shown in Plate 4 should be regarded as conceptual.