40. LATE PALEOCENE–EOCENE VOLCANIC EVENTS IN THE NORTHERN NORTH ATLANTIC OCEAN[1]

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ABSTRACT

In the North Altantic area, the thick basalts of Greenland, the Faeroes, and Scotland have been widely considered as a single province related to the early Tertiary rifting. The development of this province in time and space has largely remained unknown because of the lack of relevant geological and geophysical data offshore and the uncertainties of chronostratigraphic correlation between widely separated areas.

The Leg 81 results allow late Paleocene–Eocene volcanic events in the North Atlantic to be placed in a consistent spatial and temporal framework. The Leg 81 sites demonstrate the volcanic origin of the dipping reflectors and, using their seismically mapped distribution, allow the first definition of the true extent of the volcanic province associated with rifting. Secondly, the more precise stratigraphy established from the Leg 81 data allows a first correlation between events recorded onshore and offshore. The volcanism took place in two phases entirely during the Anomaly 24B/25 reversed polarity interval. The first phase comprised voluminous effusive volcanism (the dipping reflectors) and was succeeded by pyroclastic volcanism, recorded as ash-fall deposits in the Barents Sea, North Sea, Rockall Plateau, and Bay of Biscay, just prior to the onset of spreading in Anomaly-24B time.

INTRODUCTION

The common association of intense basic volcanism with rifting is well known and is illustrated by the comparatively uniform and voluminous flood basalts of the Columbia River, Deccan, Karroo, and Parana. (Cox, 1980).

In the North Atlantic area, the thick basalts of Greenland, the Faeroes, and Scotland have been commonly considered to be a single volcanic province whose development has been related in a general way to the breakup of the northern North Atlantic (Brooks, 1980). The development of this province in time and space has remained unknown and at best speculative, largely because of the lack of relevant offshore geological and geophysical studies and the uncertainties involved in correct chronostratigraphic correlation of widely separated areas such as Greenland and the North Sea.

In this respect, the Leg 81 results have considerable importance for two reasons. Firstly, the results demonstrate the volcanic origin of the dipping reflectors and, using their seismically mapped distribution, allow the very first definition of the true extent of the volcanic province associated with rifting. Secondly, the more precise stratigraphy established from the Leg 81 paleomagnetic and biostratigraphic results allows a first correlation between the events recorded onshore and offshore in the northern North Atlantic and therefore discussion of the development of the volcanic province in space and time.

We here present a discussion of late Paleocene–Eocene volcanic events in the North Atlantic in the context of the Leg 81 results.

DISTRIBUTION OF LATE PALEOCENE–EOCENE VOLCANICS AND PYROCLASTICS

Leg 81 results have demonstrated that the dipping reflectors observed on the southwest margin of Rockall Plateau consist primarily of basalts with interbedded pyroclastics and mineral sediments, and that the structure of the margin bears a number of striking similarities to East Greenland (Roberts et al., this volume). The lavas cored at the Leg 81 sites were erupted in the reverse polarity interval immediately preceding Anomaly 24B. Published and unpublished reflection profiles show coeval dipping reflectors across the margins of Greenland, Rockall Plateau, and the Norwegian Sea, and this allows direct mapping of the predrift extent of the volcanism (see Figs. 1, 2).

The characteristic dipping reflectors of Zone III can be mapped over the entire length of the west margin of Rockall Plateau and north of Hatton Bank to the Faeroes Plateau. The flat-lying basalts of Zone IV probably comprise the thick subaerially exposed basalts of the Faeroes. Unpublished reflection profiles west of the Faeroes (Roberts, 1978) subsequently published in part by Smythe (1983) show that the adjoining zone of dipping reflectors continues northward in the Norwegian Sea, where it lies landward of Anomaly 24B (Mutter et al., 1982).

Along the conjugate margin of East Greenland, a closely comparable suite of dipping reflectors landward of Anomaly 24B can be followed the length of the margin from Cape Farvel in the south to Scoresby Sund in the north (Featherstone et al., 1977; Larsen, 1983). The most landward parts of Zones III and IV make up the

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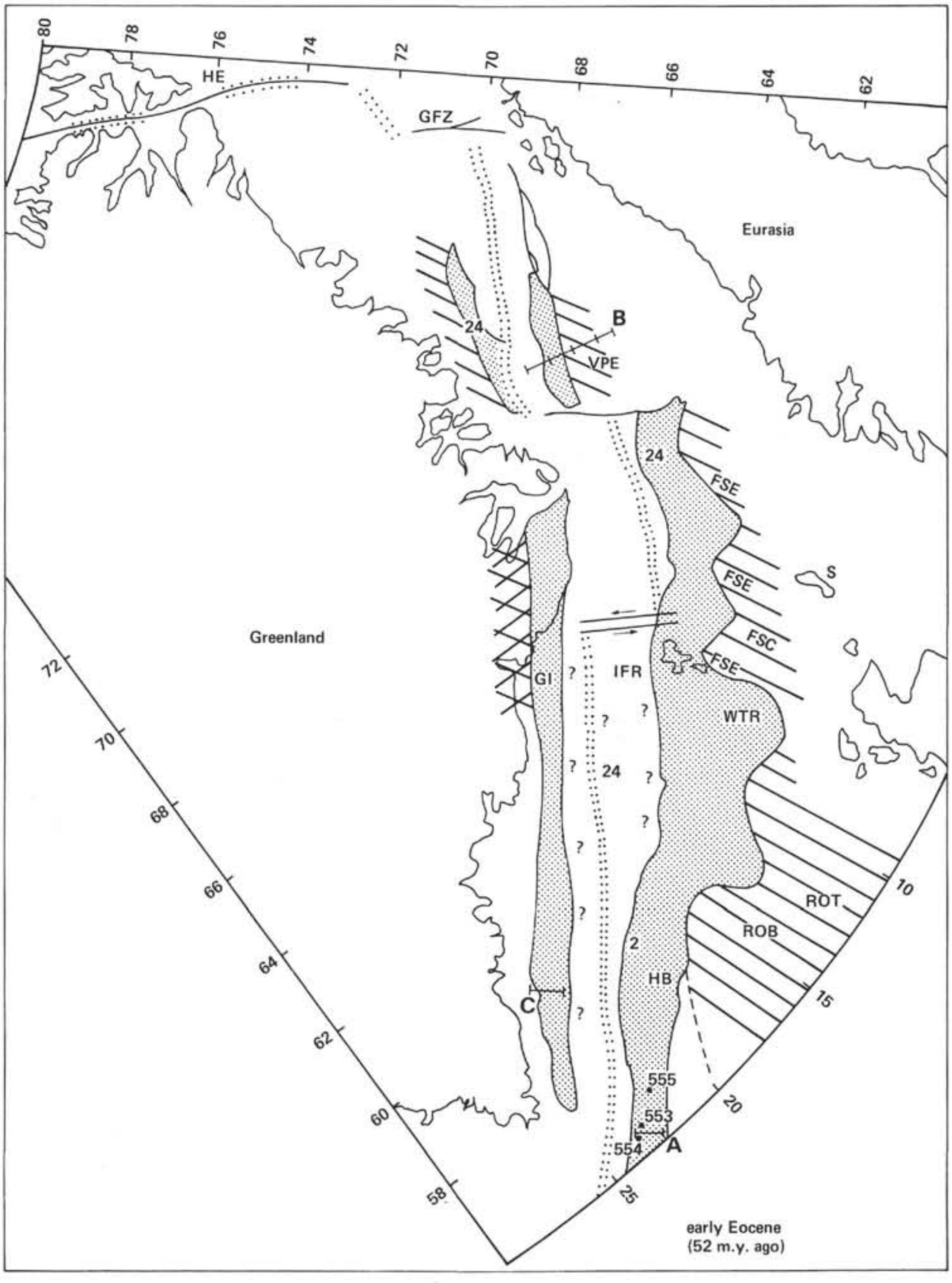

Figure 1. Distribution of dipping reflectors in the North Atlantic and Norwegian-Greenland Sea at Anomaly 24 time. Dipping reflectors shown as stippled area; area of known sills and/or flows by diagonal hatching; major plutons have been omitted. A, B, and C are lines of section shown in Figure 2.

thick late Paleocene basalts of the Blosseville Kyst (Roberts et al., this volume).

In the Norwegian Sea (Figs. 1,2), the distribution of dipping reflectors is best known on the east margin. From the Faeroes, the zone can be followed continuously as an approximately 50-100 km wide belt between the Faeroe-Shetland Escarpment of Talwani and Eldholm (1977) and Anomaly 24B to the Vøring Plateau (Mutter et al., 1982). In the latter area, the dipping reflectors lie west of the Vøring Plateau escarpment and are partly overlapped by Anomaly 24B (Mutter et al., 1982; Hagevang et al., 1982). Drilling during Leg 38 established the subaerial origin of the basalts. Further north, the dipping reflectors are observed in the same position in the greater water depths of the Lofoten Basin (Hinz and Weber, 1976). The occurrence and distribution of dipping reflectors beneath the west margins of the Norwegian Sea are less well known in part because of the complex history of spreading (Nunns, 1983). Dipping reflectors may be present beneath the Jan-Mayen Plateau (Talwani and Udintsev, 1976) and possibly beneath the east margin of Greenland north of the San Mayen Frac-

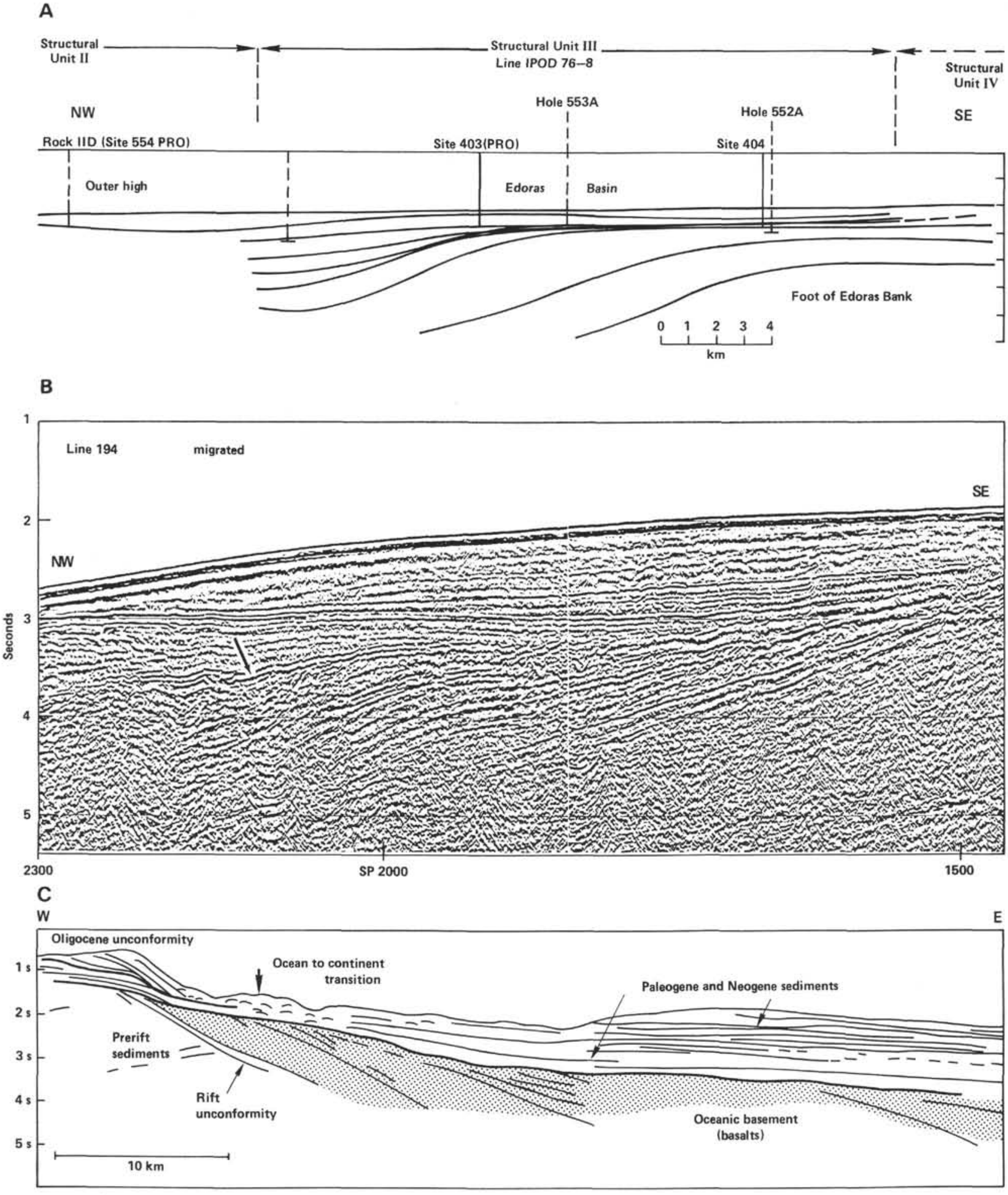

Figure 2. A. Section across dipping reflectors of southwest Rockall Plateau, structural zonation scheme used by Roberts et al., this volume. Location A is shown on Figure 1. B. Migrated multichannel seismic section across the Vøring Plateau (from Mutter, 1982). Location B is shown on Figure 1. C. Line drawing of multichannel seismic section across southeast Greenland margin (from Larsen, 1983). Location C is shown on Figure 1.

ture Zone (Mutter, pers. comm., 1983). The distribution of dipping reflectors in the northern Greenland Sea is unknown.

In addition to the inferences drawn from the distribution of the dipping reflectors, information on the extent of the province can be drawn from previous DSDP coring of coeval lavas and seismic interpretation of the distribution of lavas. On Rockall Bank, drilling at Site 117 cored NP10 sediments above basalt (Laughton, Berggren et al., 1972) which may form part of a more extensive outcrop of basalts mapped on Rockall Bank by Roberts and Jones (1975) using shipboard magnetic data. In this area, the basalts overlie proven pre-Cambrian basalt (Roberts et al., 1972). On the Rockall Plateau, high-amplitude magnetic anomalies, gravity anomalies, and subbasement reflectors suggest the presence of intrusive centers, widespread sills, and/or flows (Roberts, 1971; Scrutton, 1972; Roberts, 1975; Vogt and Avery, 1974) confirmed by cores taken on the east margin of George Bligh Bank (Ferragne et al., in press) (Fig. 3).

Between Rockall Bank and the Faeroes (Fig. 3), several large volcanic seamounts with intrusive centers are present. Within the Rockall Trough, basaltic flows and sills have been mapped seismically as far south as 54°N to just south of the Wyville–Thomson Ridge (Roberts et al., 1981; Roberts et al., 1983). In the latter area, the lava flows thicken abruptly northward of Rosemary Bank seamount to form the volcanic edifice of the Wyville-

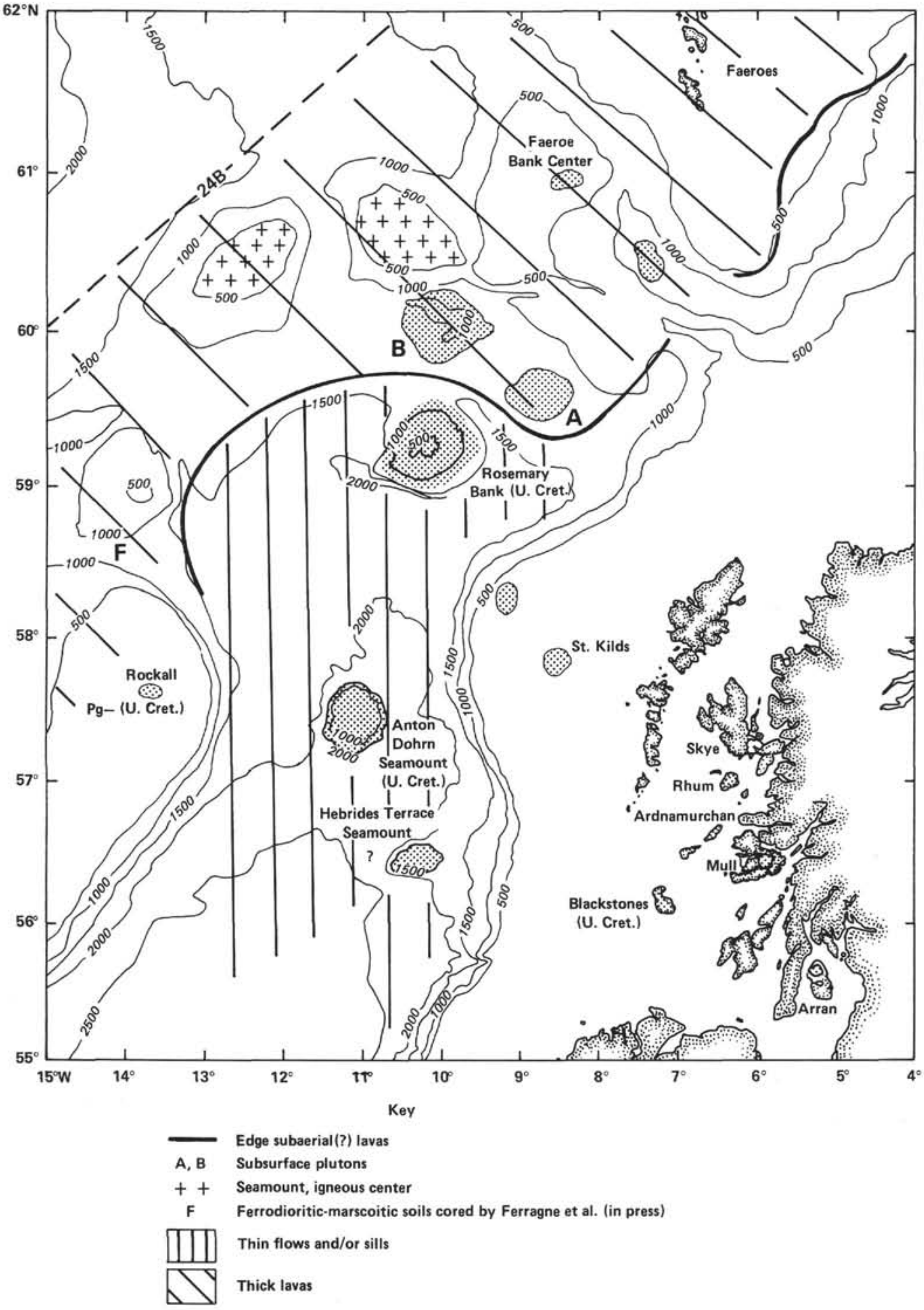

Figure 3. Distribution of lava flows, sills, etc., in Northern Rockall Trough and Wyville–Thomson Ridge area (modified from Roberts et al., 1983).

Thomson Ridge (Roberts et al., 1983). In the Faeroes-Shetland Channel a comparable distribution of thick lavas and laterally equivalent flows or sills is present. The thick lava piles of the Faeroes Plateau and Wyville-Thomson Ridge thin abruptly into the Faeroe-Shetland Channel, passing into flows and sills that extend the width of the channel. The edge of the thick pile is defined by the Faeroe-Shetland escarpment of Talwani and Eldholm (1977). The latter feature can be mapped northward as a sinuous feature from the Faeroe-Shetland Channel into the contiguous Møre Basin (Smythe, 1983; Price and Rattey, in press) where abundant sills and/or flows are present. In the adjacent Vøring Plateau Basin, the thick lavas of Zone II lying west of the Vøring Plateau escarpment pass into widespread sills or lavas that pervade the Paleocene-Eocene and earlier Mesozoic succession. Although the offshore geology of the conjugate East Greenland margin is less well known, aeromagnetic data suggest the presence of extrusive and/or intrusive bodies (Larsen, 1983). On shore, early Tertiary igneous rocks intrude and overlie the exposed Mesozoic strata of Jamieson Land.

Constraints on the age of the flows discussed above are provided by their nearly ubiquitous occurrence below the ash-marker horizon, an interregional seismic reflector of late Paleocene age clearly correlatable throughout much of the North Sea, Faeroe-Shetland Channel, Møre and Vøring basins, and the Rockall Trough. The reflection arises from a series of air-fall tuffs that occur widely throughout these areas and are known to occur in the northern Rockall Trough (Jones and Ramsay, 1982) and in the Bay of Biscay (Knox, in press). These data constrain the end of the volcanic activity. In our subsequent discussion of the chronology of these events, we have excluded the widespread Hebridean basalts which, although belonging to this province (*sensu lato*), are substantially older (early-mid Paleocene according to Curry et al. (1978).

SOUTHWEST ROCKALL PLATEAU

Three phases of volcanism have been recognized at the Leg 48 and 81 sites on southwest Rockall Plateau (Montadert, Roberts et al., 1979; Morton and Keene, this volume). The entire volcanism probably took place in the reverse polarity interval preceding Anomaly 24B (Fig. 4). In the first phase, voluminous eruption of tholeiitic basalts of MORB affinity produced the characteristic dipping reflectors of Zone III and terminated close to the NP9/10 boundary at Site 553 (Roberts et al., this volume; Harrison and Merriman, this volume; Richardson et al., this volume; Macintyre and Hamilton, this volume). Seismic profiles (Roberts et al., this volume) show that the basalts of Zone III are at least 6 km thick and that the average width of the zone is c. 40 km. In Zone IV, the equivalent sequence may have been drilled at Site 555, but it is thinner and consists of 200 m of interbedded pillow lavas, hyaloclastites, and thin marine sediments. The uppermost basalts have been dated at 53.5 ± 1.9 m.y. (Macintyre and Hamilton, this volume) and the interbedded sediments are of NP9 age (Roberts et al.; Site 555 chapter, this volume).

Phase 2 is dominated by explosive volcanism illustrated by tuffs and lapilli tuffs of basaltic composition. At Site 553, the tuffs are of tholeiitic composition but show evi-

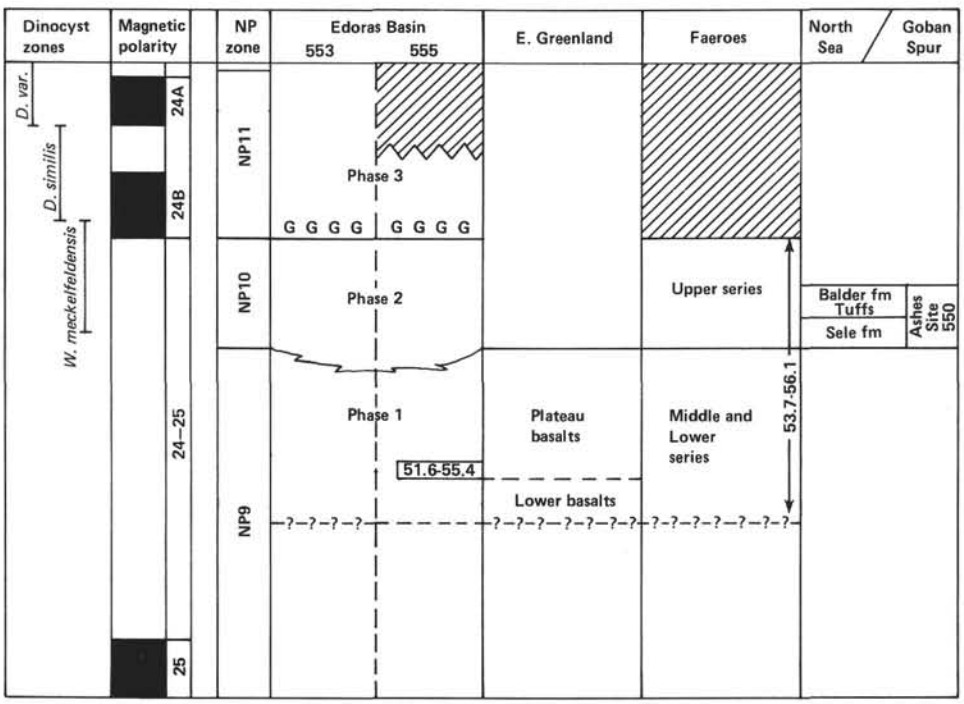

Figure 4. Tentative correlation of volcanic events in the northeast Atlantic and northwest Europe. (Correlation of dinocyst, nannoplankton zones, and magnetostratigraphy from Berggren et al., in press, and Backman and Shackleton, in press. Radiometric dates from Macintyre and Hamilton [this volume] and Fitch et al., 1978.)

dence of high-temperature alteration at source as a result of the interaction of seawater with the tholeiitic parent magma. Tholeiitic tuffs of Phase 2 occur at Site 555, but some alkali-basalt tuffs are also present.

The explosive volcanism of phase 2 was ended abruptly at or close to the NP 10/11 boundary and immediately prior to Anomaly 24B by a major transgression marked by glauconite-rich sediments. Dinoflagellate cyst analysis (Costa and Downie, 1979; Brown and Downie, this volume) places the horizon within Zone Ib of Costa and Downie (1979) and prior to the appearance of *Dracodinium similis*. The analysis thus assigns the horizon to the *Wetzeliella meckelfeldensis* Zone of Costa and Downie (1976). Studies of heavy mineral composition above and below this horizon (Morton, this volume) show that it corresponds to a cutoff in heavy minerals derived from Greenland that occurred just prior to Anomaly 24B and therefore marks the final separation of Greenland from Rockall Plateau. Above this level, volcanism is minor and represented by submarine basaltic flows and thin distally derived tuffs.

EAST GREENLAND

The thick basalts of the Blosseville Kyst of East Greenland represent the on-strike equivalents of the dipping reflectors now deeply subsided beneath the slope to the southwest (Roberts et al., this volume). In the coastal area, the basalts are thickest and penetrated pervasively by dykes (Zone II) but thin rapidly inland (Zone IV equivalent) (Water 1934; Nielsen and Brooks, 1981). The thin (c. 1 km) plateau basalts occurring north of Scoresby Sund (Upton et al., 1981) may represent Zone IV basalts separated from the Zone III dipping reflectors of the Jan Mayen Ridge by the Oligocene jump in spreading (Nunns, 1983).

Major rifting associated with the indication of basaltic volcanism in East Greenland has not been documented (Brooks, 1980; Brooks and Nielsen, 1982). The first volcanics are a complex series of basaltic flows of variable composition, including picrites (Brooks and Nielsen, 1982), interbedded with the hyaloclastites, tuffs, and tuffaceous shales of the Vandfaldsdalen, Mikis, and Hjaengefjeldet formations (Nielsen et al., 1981) of some 2 km thickness at Kangerdlugssuaq (Nielsen and Brooks, 1981). These are overlain by the subaerial Plateau Basalts which are 5 km thick at Kangerdlugssuaq and consist of uniform Fe–Ti rhoeiites (Brooks et al., 1978; Nielsen and Brooks, 1981; Brooks and Nielsen, 1982). The coast parallel THOL-I dyke swarm of Nielsen (1978) fed the Plateau Basalts, and individual dykes die out upward within the basalts. Within the Plateau Basalts, alkaline basalts are absent. However, alkali basalts do occur near the landward and presumably later extremities of the lava pile at Prinsen of Wales Bjerge (Anwar, 1955) and also in the Triangular Nunataks area, where they clearly overlie the Plateau Basalts (Fawcett et al., 1982). Alkali basalts also compose a small part of the Plateau Basalts (Zone IV) outcropping north of Scoresby Sund (Upton et al., 1981). The East Greenland succession offers a compelling analog to the volcanic sequence of southwest Rockall Plateau. In both cases the main tholeiitic basalts of MORB affinity are succeeded by minor alkali-basalts erupted in an off-axis position. The occurrence of a similar association in Iceland (Jakobbson, 1972) indicates that the succession can occur in both oceanic and continental settings. It is striking, however, that the Greenland succession is characterized by the *absence* of explosive basaltic volcanism. The hyaloclastites were formed by brecciation of submarine basalt flows (cf. Site 555 chapter, this volume), and the associated tuffs represent the products of erosion of adjacent subaerial flows (Nielsen et al., 1981; Upton et al., 1981) in contrast to the tuffs found at Sites 553 and 555.

The eruption of the basalts was closely followed by major igneous intrusion represented by the classic basic plutons of the Skaergaard and Kap Edvard Holm and the related THOL-2 dyke swarms (Nielsen, 1978), major salic plutons such as Kangerdlugssuaq, and alkaline dykes and the development of the coastal flexure.

No comprehensive magnetostratigraphic study of the East Greenland succession pile has been attempted although all published work suggests that the entire sequence is reversely magnetized (Tarling, 1967; Watt and Watt, 1971; Hailwood et al., 1973; Faller, 1975; Nielsen et al., 1981). By contrast, biostratigraphic studies have focused on dinoflagellate cyst analyses of interbasaltic sedimentary horizons in the key basal and upper parts of the succession (Soper and Costa, 1976; Soper et al., 1974, 1976). Sediments from the Vandsfaldjelen Formation near the base of the succession have yielded a very sparse flora consisting only of *Apectodinium homomorphum* and *Acritarch* sp. nov.

The Sparnacian age proposed originally by Soper et al. (1976) is now questionable because Harland (1980) has demonstrated that the association of forms (called by him the *Apectodinium homomorphum* plexus) is not age diagnostic; the occurrence of the association at any level in the Ypresian rather reflects adverse environmental conditions. The extremely low diversity of the sparse flora obtained from the Vandsfaldjelen Formation clearly suggests such conditions. The important point is that the onset of basic volcanism cannot be precisely determined from the available biostratigraphic data although it must have occurred in Sparnacian–Ypresian time and presumably during the reverse polarity interval immediately prior to Anomaly 24B.

The termination of basaltic volcanism can however be more precisely constrained by dinocyst biostratigraphy. Samples from an interbasaltic horizon near the top of the lava pile at Kap Dalton yielded a diverse dinocyst assemblage allowing reasonable assignment to the *Wetzeliella meckelfendensis* Zone (Soper et al., 1974) of Costa and Downie (1976). Although the sediments overlying the basalts at Kap Dalton have less rich assemblages (Soper and Costa, 1976), their diversity is sufficient to allow assignment to the *Drawdinium varielongitudum* Zone (Soper et al., 1976). These results show that basaltic volcanism ended during the interval between the *W. meckelfeldensis* and *D. varielongitudum* zones. These results by comparison with the biostratigraphy established on the Rockall Plateau (see Backman et al., this volume) and the reversely magnetized basalts of Greenland indicate cessation of basic volcanism during the later part of the reverse polarity interval prior to Anomaly 24B (Fig. 4).

The assignment of the East Greenland basalts to this reverse polarity interval shows a direct correlation with the volcanics of southwest Rockall Plateau. Radiometric and fission track ages are generally compatible with this correlation (Fig. 4). In East Greenland, Brooks and Gleadow (1979) show that the Skaergaard intrusion was emplaced at 54.6 ± 1.7 m.y., towards the end of the period of basaltic eruption at 54.5 ± 0.5 m.y. favored by Fitch et al. (1978) using samples close to the top of the basaltic pile. At Site 555, fission track ages were obtained from sediment interbedded with the basalts (Duddy et al., this volume); the smallest and youngest ranges of crystal ages yielded an average age of 54 ± 3 m.y. Macintyre and Hamilton (this volume) have derived a similar age of 53.5 ± 1.9 m.y. for the top of the basalts at Site 555. It should be noted that a younger age range of 48.6 to 53.9 m.y for East Greenland basaltic volcanism has been proposed by Odin and Mitchell (1983). These ages do not lie within the 56.1–58.6 m.y. age range for the reverse polarity interval between Anomalies 24B and 25 recently proposed by Berggren et al. (in press). Notwithstanding the small difference with the Berggren et al. (in press) time scale, the Rockall and Greenland ages demonstrate a close coincidence in the cessation of basaltic volcanism.

FAEROE ISLANDS

Correlation of the Faeroes basalt succession with East Greenland, Rockall, and elsewhere is more difficult because marine palynomorphs are absent. Nonetheless, the available evidence shows that the Faeroese succession is correlative with the latter areas.

Rasmussen and Noe-Nygaard (1970) have subdivided the Faeroese basalt pile into a Lower, Middle, and Upper Series (Fig. 4). The Lower and Middle Series are divided by a coal unit up to 15 m thick overlain by tuffs and agglomerates of explosive origin. Tarling and Gale (1968) have shown that the sequence is predominantly reversely magnetized and identify two short normal-polarity intervals in the Lower Series. These short polarity intervals probably represent minor excursions detected only because of the high rate of basalt accumulation (Krumsiek and Roberts, this volume). Similar short polarity intervals in the late Paleocene have been detected at Sites 553 and 555 (Krumsiek and Roberts, this volume) and also in the North Sea and southeast England (Hailwood, pers. comm.) although further work is needed to confirm their significance.

The Lower and most of the Middle Series form a chemically uniform group comparable with the East Greenland basalts (Noe-Nygaard and Rasmussen, 1968; Brooks et al., 1976). However, the late Middle Series and Upper Series basalts show a marked decline in TiO_2, Fe/Mg, and $(La/Sm)_{EF}$ values attributed to a change from plume-related magmatism to more typical mid-ocean ridge magmatism (Schilling and Noe-Nygaard, 1974). The apparent absence of a comparable change in East Greenland or Rockall may suggest that the Upper Series is younger but the close overlap in ages discussed below suggests that lateral variations in magma composition in the rift system may be a more plausible model.

Radiometric age determinations on the Faeroese basalts show a wide range of dates from 50.3 to 63.1 m.y. (Tarling and Gale, 1968), recalculated using ICC constants. However Fitch et al. (1978), using a regression analysis technique, narrowed the range to 54.5 ± 1.2 to 55.1 ± 1.0 m.y. These dates overlap with those quoted previously in this chapter and place the Faeroese basalts in the Anomaly 24/25 reverse polarity interval (Fig. 4). Although less precise, a late Paleocene age is given by study of pollen and spores from the intrabasaltic coal (Lund, 1983). Smythe et al. (1983) have proposed that the thin overlying tuff agglomerate sequence is equivalent to the Balder Formation tuffs of the North Sea. The regional significance of this proposal is, however, questionable, since the dissimilarity of the Balder Formation tuffs and Faeroese basalts (Morton and Knox, pers. comm.) does not suggest a common source. The Faeroese tuff conglomerate can therefore be regarded as a local event caused by subaqueous basaltic extrusion in the swampy conditions associated with the underlying coals that ceased once the pile had become subaerial.

In the area south of the Faeroes, Jones and Ramsay (1982) have dredged tuffs from the flank of the circular volcano forming the Faeroes Bank seamount. Although the origin of the material is unknown, the biostratigraphic age (NP11/NP12) suggests that these beds postdate the main volcanic phases discussed earlier in this chapter. Further south, Ferragne et al. (in press) have cored subaerial clays of lower Eocene age that represent the subaerial weathering products of ferrodioritic-marscoitic rocks.

VØRING PLATEAU AND THE NORWEGIAN SEA

The occurrence and distribution of dipping reflectors on the Vøring Plateau and in the Norwegian Sea was discussed earlier in this chapter. Only limited drilling has taken place on the Vøring Plateau, but the relevant DSDP sites (Sites 338, 342 and 343) proved to be early Eocene sediments overlying basalts cored at TD (Talwani and Udintsev, 1976). Radiometric dating has yielded ages incompatible with the well-constrained biostratigraphic age of the overlying sediments (Talwani, Udintsev, et al., 1976). Age relationships within the Vøring Plateau will require further drilling of the type proposed for the Advanced Ocean Drilling Program. It is nonetheless clear from the relationship of the basalts to the magnetic anomalies (Talwani et al., 1981; Hagevang et al., 1982) that the basalts cannot be younger than the end of the Anomaly 24B normal polarity interval.

NORTH SEA AND ADJACENT AREAS

Early Tertiary sediments of the North Sea Basin over an area extending from north of the Shetlands, Southeast England, Netherlands, Northwest Germany, and Denmark record a major phase of air-fall pyroclastic activity (Jacque and Thouvenin, 1975; Knox and Morton, 1983). Where complete, the sequence contains more than 200 tuffs. The lower unit of the sequence contains sporadic and chemically heterogeneous ash layers; this lower unit is the Sele Formation of Deegan and Scull (1977) and is equivalent to the lower part of the "negative series" of the Danish mo-clay. The upper unit contains by contrast abundant tuffs almost wholly of tholeiitic composition (Morton and Knox, pers. comm.); the upper unit is equivalent to the lower part of the Balder Formation of Deegan and Scull (1977) and to the upper part of the "negative series" and the entire "positive series" of Denmark. To the north and west of the North Sea, the ash sequence has been followed seismically by well log correlation into the West Shetland Basin (Ridd, 1983; Smythe et al., 1983) and along the Norwegian continental margin as far north as Troms (Jakobssen et al., pers. comm.). Throughout this area, the sequence is characteristically barren of coccoliths. A direct correlation to the standard nannoplankton scheme is therefore impossible. Recently, however, Knox (in press, a and b) has detected the tuff sequence close to the detrital magnetic event near Anomaly 24B observed by Montadert, Roberts et al. (1979) in the Bay of Biscay and also at Leg 80 sites on the Goban Spur. At DSDP Sites 401 and 550, about 40 thin fine-grained bentonite layers occur in the lower Eocene. These bentonites show identical compositional trends to the North Sea tuffs (a lower group of variable composition and an upper group of uniform tholeiitic composition) and are thus believed to be their distal equivalents. Unlike the North Sea, indigenous coccoliths are present and show that the bentonites lie wholly within the lower part of NP10 in sediments containing *Tribrachiatus nunnii* but not *T. contortus* (Müller, in press). The tuffs of the Balder and Sele Formations in the North Sea and West Shetland Basin can therefore be placed in the lower parts of NP10 and can be correlated with the lower part of the pyroclastic phase succeeding the basalts at Sites 553 and 555 (Fig. 4). This correlation has been verified by paleomagnetic studies made by Hailwood (pers. comm., 1983) which show that the North Sea tuffs are reversely magnetized and that they fall into the Anomaly 24–25 reverse polarity interval.

DISCUSSION

Several lines of evidence document the hitherto understated importance of voluminous volcanism during the short reversed polarity between Anomalies 24 and 25. These lines of evidence include the recognition of coeval suites of dipping reflectors of volcanic origin on the conjugate margins of Greenland, Rockall, and the Norwegian Sea from Leg 81 results; the correlation of events with effusive and pyroclastic volcanic activity over a wider area extending from Greenland via the North Sea to Denmark and from the Bay of Biscay in the south to Troms in the north. The total extent of this province is shown in Figure 5. The area of known basalts is clearly comparable to the great flood basalt provinces of the Deccan and Parana. A qualitative estimate of the volume of basalt is obviously difficult. However, a minimum estimate can be gained from the known area (100 $\times$ 2000 km) and an assumed average thickness (4 km) of the dipping reflectors and yields a very approximate volume of about 1 million km^3 exclusive of the undoubtedly large volume of basalts known to occur beyond the dipping reflectors; a total of about 2 million km^3 would not seem unreasonable and compares well with figures of about 3 million km^3 for the Deccan. The Leg 81 results suggest that at least there the entire dipping reflector sequence (at least 4 km thick) was erupted during the very short period of time (order 10^5 yr.) between Anomaly 25 and the NP9/10 boundary. Although a somewhat longer time interval is suggested by the data from East Greenland and the Vøring Plateau, it is incontestable that this substantial volume of basalt could not have been erupted in a period longer than the Anomaly 24–25 reversed polarity interval and was possibly erupted in a good deal less time.

It would be useful to be able to correlate the volcanic sequences of East Greenland, Southwest Rockall, and the other areas in more detail but the correlation remains constrained by the limitations of biostratigraphic radiometric dating techniques in resolving closely spaced events in an interval as short as the Anomaly 24-25 reversed polarity interval. Systematic magnetostratigraphy studies will be the future key to this problem. The difficulty of timing the onset of basaltic volcanism accurately has already been noted but the magnetostratigraphic and magnetic anomaly data show it cannot be earlier than the base of the 24–25 reversed polarity interval. However, as rifting rapidly progressed northward, the development of basaltic volcanism within the short Anomaly 24–25 interval may have been diachronous. The same consideration might also apply to a sought-for correlation between the eruption of basalts in flows and the eruption of tuffs. However, a simple model is that tectonic subsidence related to the terminal phases of rifting allowed the sea to invade the rift, thereby producing shallow marine eruptions and associated pyroclastics that were then distributed over the whole area observed. Evidence in support of such a model is given by similar age (base NP10) of the tuffs observed in Biscay, Rockall, and by inference in the North Sea, suggesting a widespread pyroclastic event. A sudden and rapid fall in activity towards the top of NP10 and just before Anomaly 24B coupled with the marine transgression suggests that the cessation of activity may be related to the transition from rifting to spreading (greater water depths?) between Greenland and Rockall. The waning of basaltic eruption recorded at the *W. meckelfeldensis* zone in Greenland and the change in the chemical character of the Faeroes, basalts may mark the same event.

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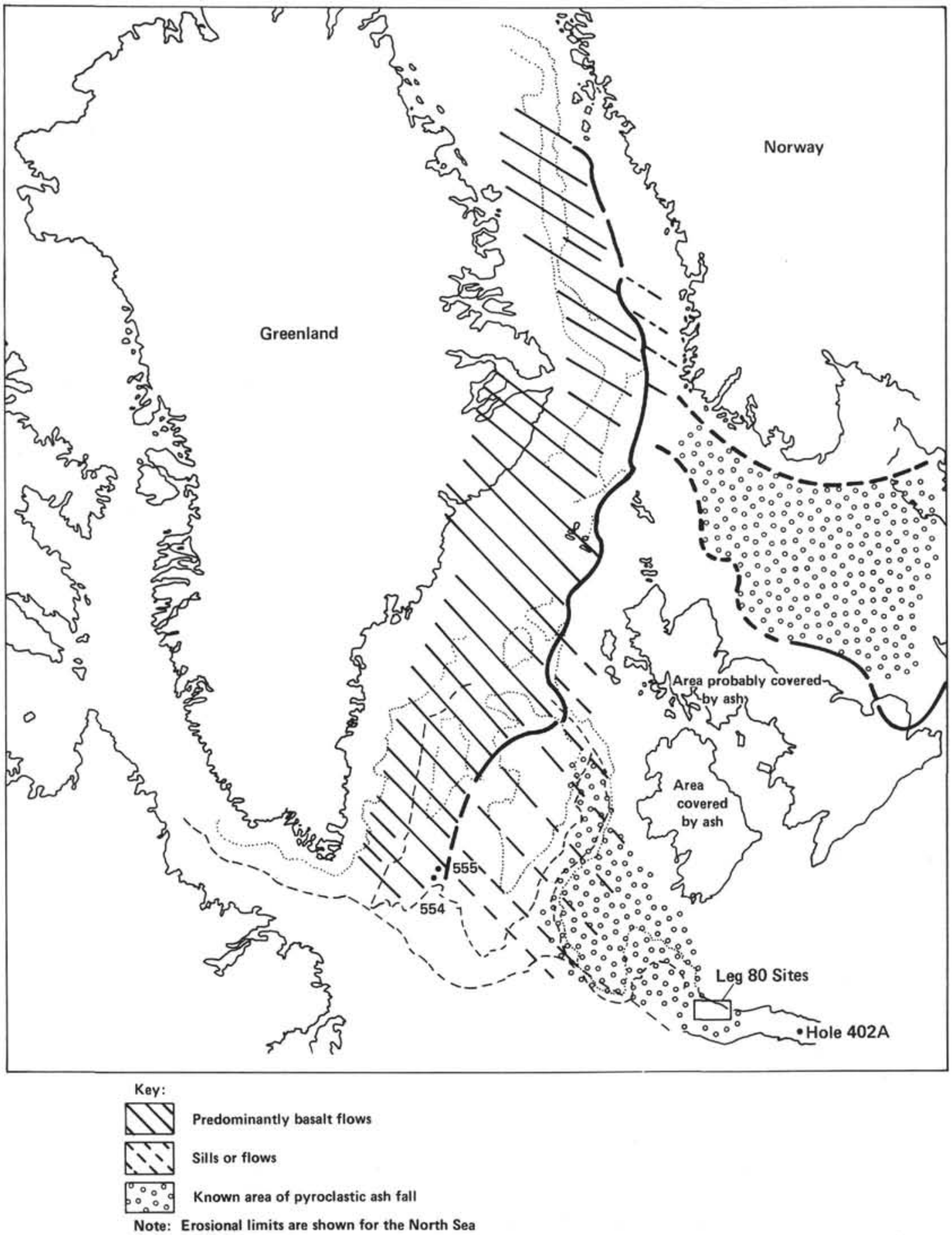

Figure 5. Distribution of Paleocene–Eocene basalts and pyroclastic deposits in Greenland and Europe. Sources discussed in the text. Dotted area = known area of pyroclastic ash fall. Note that erosional limits are shown for the North Sea.

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