

39. CRUSTAL STRUCTURE ALONG THE LEG 152 DRILLING TRANSECT¹

Hans Christian Larsen,² Trine Dahl-Jensen,² and John R. Hopper²

ABSTRACT

Seismic data, drilling data, and gravity data define a 40-km-wide continent-ocean transition (COT) below the outer half of the 70- to 80-km-wide shelf along the Leg 152 drilling transect. The seismic data include a detailed grid of shallow high-resolution seismic data; a regional grid of conventional multichannel seismic data (including sonobuoy refraction data); and a deep crustal, vertical incidence multichannel seismic profile. Below the outer shelf, the top of the Precambrian crust off southeast Greenland dips through a normally faulted, seaward-facing flexure zone below the landward feather edge of the seaward-dipping reflector sequence (SDRS), which is subaerially erupted basaltic lavas. Associated with the flexure zone, the continental crust thins from approximately 28 km thick to almost zero over a maximum distance of 40 km. Variations in present-day crustal thickness are much less because of accretion of thick igneous crust during breakup. Most of the thinning of the continental crust takes place within only 25 km from the mid-shelf flexure zone. Within this zone, landward-dipping normal faults are present in the upper continental crust, indicating that thinning was accommodated by fault-related crustal extension. Lower crustal thinning appears to be of a different nature and may be related to "intrusive" underplating. The seismically defined COT is consistent with a distinct development from initially continental to oceanic volcanism observed in the boreholes. The new igneous SDRS crust seems to be 20–22 km thick, including at least 5 km of lava in the upper crust. Deep reflections located below the COT at 30–35 km depth may represent a pillow of thick (10 km or more) underplated material or intra-mantle reflections from a residual mantle.

INTRODUCTION

The southeast Greenland margin is characterized by a wide, margin-parallel zone with a seaward-dipping reflector sequence (SDRS; Larsen and Jakobsdóttir, 1988; Larsen, 1990). The SDRS is present below the middle to outer shelf and in the adjacent deep-water area (Figs. 1, 2). Larsen and Jakobsdóttir (1988) interpreted the main part of southeast Greenland SDRS as subaerially erupted lavas that emanated from a fairly narrow, Icelandic-type spreading center. This implies that the SDRS crust represents Icelandic-type oceanic (igneous) crust (Pálmason, 1986) with the continent-ocean transition (COT) located below the inner part of the SDRS. Testing these interpretations by borehole data was one of the major tectonic objectives of Ocean Drilling Program (ODP) Leg 152.

The Leg 152 drilling transect is the southernmost of two southeast Greenland margin transects studied by ODP and Deep Sea Drilling Project (DSDP), named the EG63 and EG66 transects, respectively (Fig. 1). Drilling during Leg 152 penetrated the feather edge of the SDRS and sampled the central part of the SDRS (Figs. 3, 4). The recovered rocks were exclusively subaerially erupted and mainly basaltic lavas. The SDRS comprises a lowermost (innermost) continentally contaminated lava sequence followed by picrites and tholeiites with much reduced or no continental contamination (Larsen, Saunders, Clift, et al., 1994; Fitton et al., Chaps. 28 and 29, this volume; Saunders et al., this volume). This evidence is in strong support of the interpretation of the SDRS as representing Icelandic-type crust, as well as for a location of the COT directly below the feather edge of the SDRS (middle outer shelf). However, because there are strong variations along the margin in the degree of Neogene shelf-edge progradation and in the width of the shelf, the position of the COT can vary from below the shelf slope to the inner middle shelf (see also Larsen, 1990).

The goal of the present paper is to provide a more detailed model for the crustal structure and crustal deformation across the COT along the EG63 transect. The model is supported by (1) Leg 152 drilling data; (2) high-resolution, conventional, and deep crustal multichannel seismic (MCS) data; and (3) gravity data. Acquisition of wide-angle seismic data for deep crustal velocity control is planned for 1996.

SETTING OF THE EG63 TRANSECT

In the northeast Atlantic Ocean, the oldest identified pair of seafloor spreading anomalies are C24n–C24r (Talwani and Eldholm, 1977; Srivastava and Tapscott, 1986; Larsen, 1988). Anomaly C24n (53 Ma; Cande and Kent, 1992) is developed off southeast Greenland as a double-peaked anomaly. Along the Leg 152 transect it is partly overlapped by the most seaward and youngest part of the SDRS; hence, it provides a minimum age for the SDRS (Fig. 2; Larsen and Jakobsdóttir, 1988). However, the SDRS extends seaward to chrons C22n–C21n (49–47 Ma) and may extend into even younger anomalies in the north near the Iceland-Greenland Ridge. Weak and semi-linear magnetic anomalies are present over the main SDRS along the Leg 152 transect, and they may represent either low amplitude anomalies older than C24n (e.g., C25n–C27n, 56–61 Ma) or short magnetic events within C24r (the cryptochrons of Cande and Kent, 1992; see also the discussion in the "Summary and Principal Results" chapter of Larsen, Saunders, Clift, et al., 1994; and Larsen and Saunders, this volume). Lavas from the Middle Series at Site 917, which are among the oldest part of the SDRS along this transect, have yielded ⁴⁰Ar–³⁹Ar ages of 61–62 Ma (Sinton et al., 1994; Sinton and Duncan, this volume; Werner et al., this volume).

The crustal structures landward of the East Greenland SDRS show considerable variation from south to north. A three-fold division of the margin structure along strike was defined by Larsen (1990). The transect studied in this paper is located within the southernmost segment of these three margin segments, which stretch from the southern tip of Greenland to the northern part of the Iceland-Greenland Ridge in the north, and consists almost entirely of high-

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²Danish Lithosphere Centre (DLC), Øster Voldgade 10, 1350 K, Copenhagen, Denmark. larsenhc@dlc.ggu.min.dk

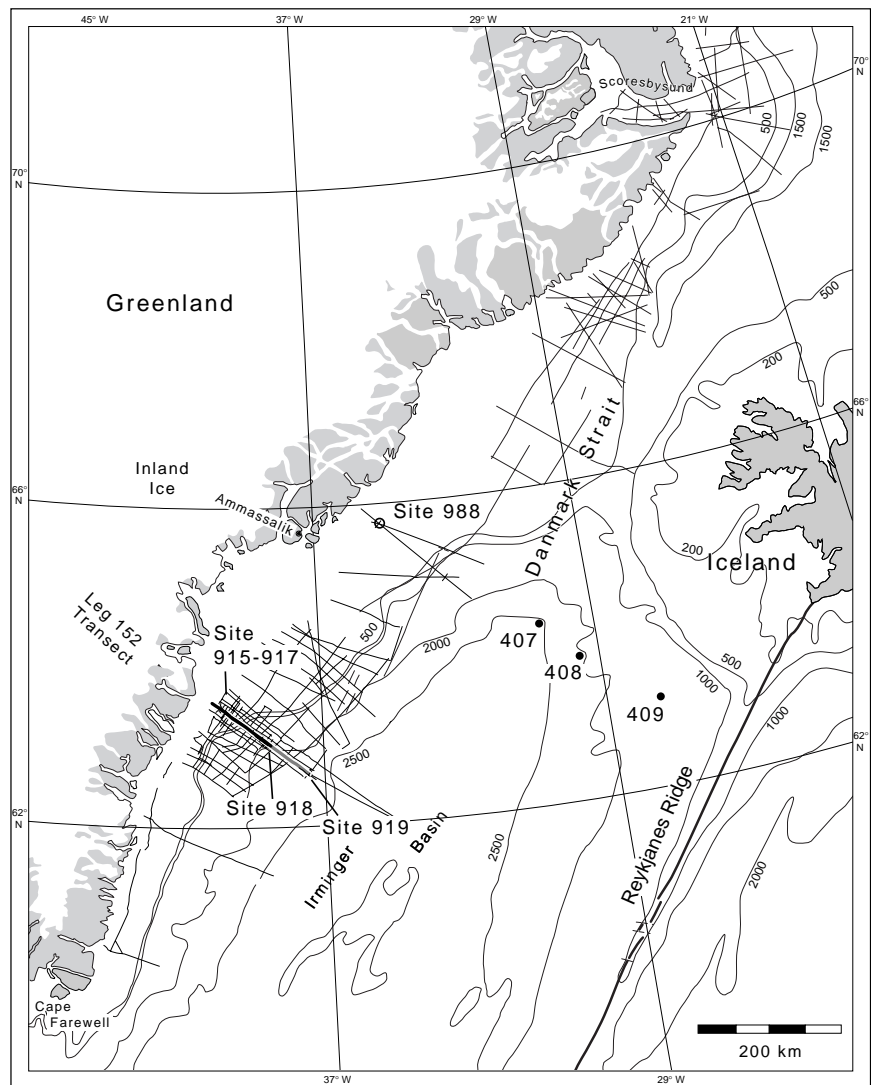


Figure 1. Study area, seismic coverage, and drill sites. Sites 914–919 defining the southern East Greenland transect were drilled during Leg 152. DSDP Sites 407–409 and Leg 163 Site 988 defining a northern transect also shown though not discussed here. Sites 915–917 were re-drilled during Leg 163 (Duncan, Larsen, Allan, et al., 1996). The detailed grid of seismic data are high-resolution, shallow seismic data (Larsen et al., 1994b). The more regional grid comprises conventional multichannel seismic data (Larsen, 1983, 1990). The deep seismic profile (heavy black line) covers approximately the landward half of the drilling transect.

grade basement rocks of Precambrian age (coastal zone and inner shelf) that become replaced seaward by breakup-related lavas and younger sediments (Figs. 1, 2). The rift lavas (the SDRS) therefore onlap the Precambrian (Archean to early Proterozoic) basement (Figs. 2–4). The onlap zone forms a crustal monocline, seaward flexure. This is different from the two northern margin segments where breakup took place within a complex area affected by the Caledonian orogeny and subsequent late Paleozoic and Mesozoic rift basin formation. Compared to other more complex settings, the development of the southeast Greenland margin as a single, Late Cretaceous to early Tertiary rift-drift event within a craton offers unique possibilities to isolate and image the deformation specifically associated with breakup. Furthermore, large amounts of uplift, glacial erosion, deep crustal exhumation, and the sediment-starved nature of the continental shelf (Larsen, 1990) all serve to provide unique coastal exposures close to the inferred COT. A coast parallel dike swarm and associated seaward flexure of the crust is found along the coast from the EG66 transect and northward. Gabbroic and syenitic intrusions are present within this zone (Fig. 2; Myers, 1980; Myers et al., 1993) and are locally associated with basaltic lavas overlying thin sediments (see Larsen, 1980; Nielsen et al., 1981; and see Brooks and Nielsen, 1982, for a review). These isolated exposures of basalt may well be outliers of the more extensive and thick flood basalt province farther north

(Fig. 2; Larsen et al., 1989). Based on aeromagnetic data, Larsen (1978) suggested that the coastal flexure and dike swarm have a wide offshore continuation to the south that intersects the landward parts of the two drilling transects (Figs. 1, 2).

GEOPHYSICAL DATA ALONG THE TRANSECT

The geophysical data along the transect comprise aeromagnetic data (Larsen and Thorning, 1980), 24-fold 7-s MCS data (with regional sonobuoy refraction data; Larsen and Jakobsdóttir, 1988), 6-fold 3-s high-resolution data (Larsen et al., 1994b), and deeper 60-fold 14-s MCS data (with some wide-angle data close to the shore). In addition, regional marine gravity, land-based gravity, and satellite gravity data exist and were used to construct a profile along the studied cross section.

The three different MCS data sets provide imaging at different crustal levels. The high-resolution data are clearly superior within the upper 500 meters below the seafloor and in places show significant reflectors down to 1–2 km depending on the water depth and seafloor multiple (Larsen, Saunders, Clift, et al., 1994; Lykke-Andersen, this volume). The conventional MCS data provide the optimum image of the upper crust between 4- and 8-km depth (Fig. 3). The deep MCS

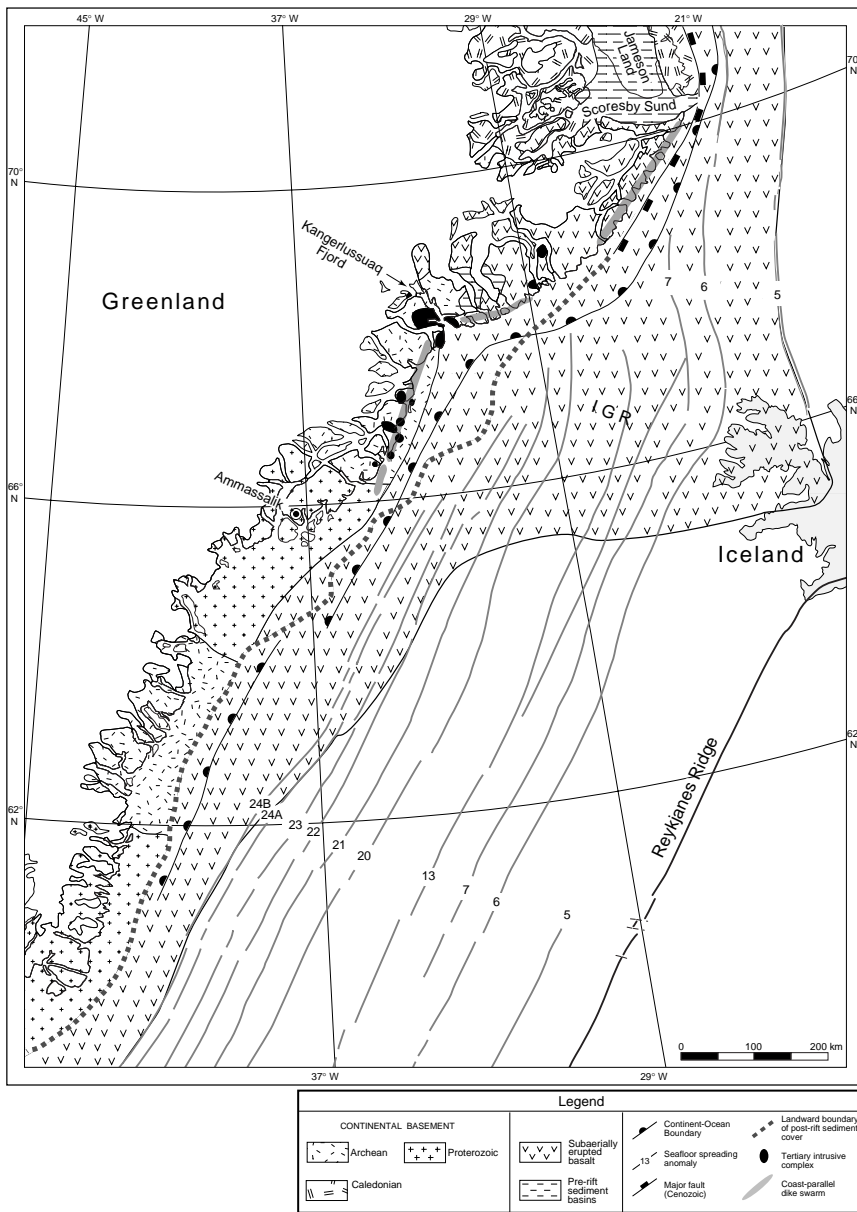


Figure 2. Geological map of the study area. The Leg 152 study is located within the northern part of the southeast Greenland Archean block and the igneous crust seaward of this (see Fig. 1 for precise location). In the north, the Precambrian crust is replaced by the East Greenland Caledonides. Extensive late Paleozoic and Mesozoic rifting and basin formation preceded continental breakup. South of the Iceland-Greenland Ridge (IGR), the extrusive part of the new igneous (oceanic) crust onlaps the Precambrian crust. The subaerially erupted lavas, which are in places at least 5 km thick, are widely distributed over both continental crust, where they form plateau basalts, and within the oldest oceanic crust seaward of the continent/ocean boundary, where they form the seaward-dipping reflector sequence (SDRS). Larsen (1978) suggested that the coast-parallel dike swarm and flexure extend offshore and southward and cross the inner part of the Leg 152 transect. Note how the seaward termination of the SDRS is diachronous. The oldest seafloor spreading anomaly identified in the northeast Atlantic is the pair 24A and 24B (representing magnetochrons C24n.1–C24n.3; Cande and Kent, 1992).

data take over from this level and include Moho and possibly some sub-Moho reflections (Fig. 5).

Oceanward, the aeromagnetic data define the seafloor-spreading anomalies (Figs. 2, 3). Landward, the magnetic data show a near-surface magnetic basement with large amplitude anomalies of similar nature to those associated with the Precambrian basement that outcrops along the coast and below the inner shelf (see also the regional magnetic map by McNab et al., 1995). The middle shelf is characterized by large amplitude magnetic anomalies that Larsen (1978, 1990) correlated with the East Greenland coastal dike swarm and flexure zone.

Seismic data from the southeast Greenland shelf and slope show the presence of strong seafloor-generated multiples, in particular below the outer shelf banks. The first seafloor multiple effectively defines the depth of penetration on the shallow seismic data. Frequency-wavenumber (F–K) filtering followed by normal moveout (NMO) corrected stacking of common depth point (CDP) gathers with a carefully picked inner trace mute is fairly effective in removing multiple energy on the longer offset, 7- and 14-s data (2400 m and 4600 m

streamer, respectively). The 14-s data were also processed with wave-equation demultiple-processing. Nonetheless, any reflections at depth that mimic the slope of the seafloor must be treated with great caution.

UPPER CRUSTAL STRUCTURE AND THE NATURE AND TIMING OF VOLCANISM

The Precambrian basement exposed along the southeast Greenland coast extends below the inner shelf and its upper surface dips, through a margin parallel flexure, below the landward feather edge of the SDRS. Seaward of this flexure, Tertiary and Quaternary sediments rest on top of the SDRS and thicken below the outer shelf (Larsen, 1990). Just over 1 km of sediment (mainly glaciomarine deposits) is present below the outer shelf at the EG63 transect (Figs. 3, 4). The SDRS also thickens seaward from about 1 km within the flexure zone to 5–6 km below the continental slope and rise (Fig. 4; Larsen and Jakobsdóttir, 1988).

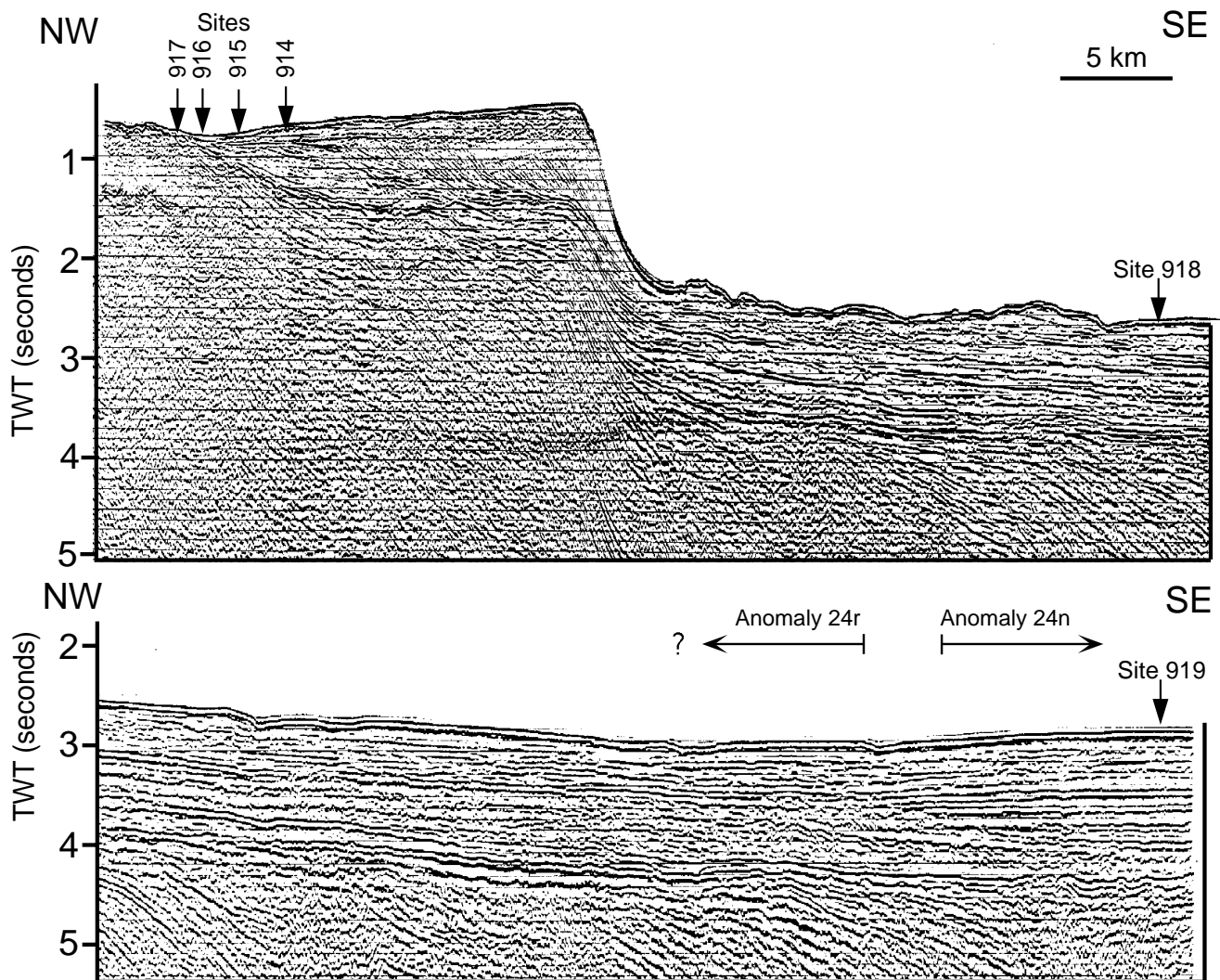


Figure 3. Seismic Line GGU81-08 (migrated multichannel data) along the Leg 152 transect. The deep seismic profile (Fig. 5) extends from slightly landward (NW) of this profile and close to Site 918. Note how the SDRS is better imaged in the deep water region. To the south of the profile and the Leg 152 transect, progradation of the shelf (mainly by glacial processes) is less, and the SDRS can be followed more landward.

Coast-Parallel Crustal Flexure Zone

The margin-parallel flexure zone is close to the coast from about 66° to 68° N where it parallels the East Greenland coastal dike swarm (Wager and Deer, 1938; Nielsen, 1978; Larsen, 1978; Myers, 1980; Karson et al., 1994). Shallow seismic data along the EG63 transect indicate the presence of landward-dipping normal faults bounding blocks of seaward-dipping reflectors (lavas) within the flexure zone. One fault was sampled by Sites 916 and 917 (Fig. 4; Larsen, Saunders, Clift, et al., 1994). In general, correlation of fault traces on the seismic grid is difficult, but a roughly margin parallel to perhaps slightly margin oblique strike is indicated (i.e., between north-northeast-south-southwest and northeast-southwest). This is consistent with the strike and dip direction of the lavas. The true dip of the landward-dipping normal fault penetrated by Sites 916 and 917 depends on the details of the interpretation of the multiple fault traces within the fault zone, and on the exact strike of the fault. Within the faulted zone, fault planes dipping as steep as 50° or more seem present, but the overall dip of the fault zone may not be more than 30° – 40° . Using Formation MicroScanner borehole data, Cambay (this volume) interprets the seaward dip of the basaltic units to steepen from about 10° – 15° at the top to about 20° – 25° at approximately 800 m below the seafloor. These dips are about equal to or slightly greater than the

dips indicated by the seismic data, and conflicting dip directions are present at shallow levels. The seismic data, in particular in the upper part of the lava pile, are considered to provide the most reliable measure of regional dip.

Eocene sediments show onlap to the surface of the tilted lavas as well as some seaward dip and minor faulting (see “Summary and Principal Results” chapter in Larsen, Saunders, Clift, et al., 1994). Thus, about 5° – 10° of the dip within the lavas is due to later flexural bending of the faulted zone related to post-volcanic cooling and subsidence of the new igneous crust seaward of the faulted zone. During this late tilting only very minor fault movements took place compared to the several hundred meters of displacement that occurred between the deposition of the oldest lavas (Paleocene, 62–61 Ma) and the overlying (lower?) to middle Eocene sediments. During this time of active fault movements (and possible fault activity prior to volcanism), the dip of the normal fault was more than the present dip, perhaps as great as 50° – 60° .

Prebasaltic Sediments

Meta-sediments of unknown age (Cretaceous?) and with a steep to subvertical orientation were discovered at Site 917. These lie unconformably (40° – 50° unconformity) below the basalts and a thin,

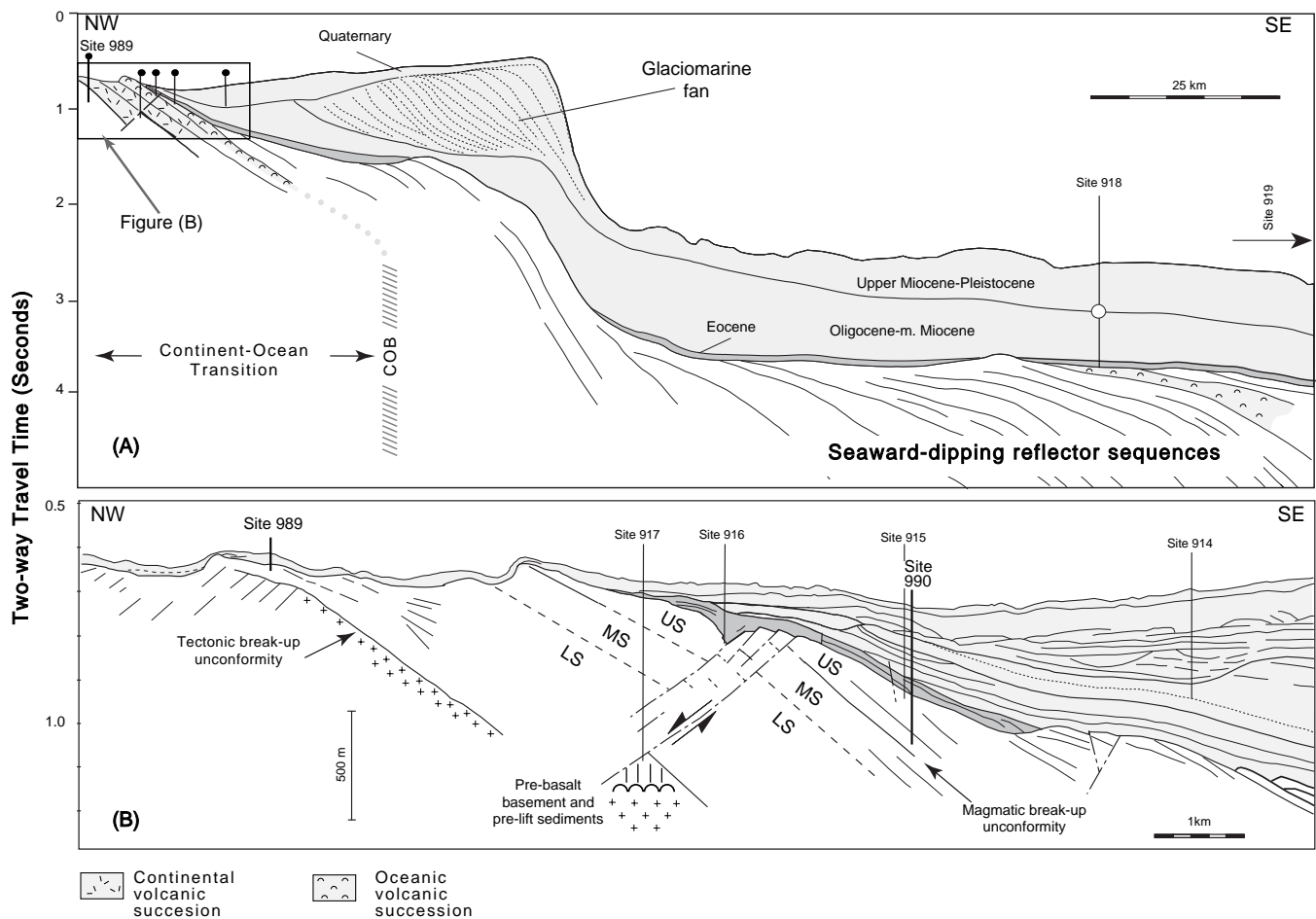


Figure 4. Interpreted section based on data from Figure 3 and shallow seismic data (Larsen, Saunders, Clift, et al., 1994; see also Clausen, and Lykke-Andersen, both this volume). Note that the volcanic sequence in (A) is divided into a lower continental succession and an upper oceanic succession. The continental succession comprises the Lower Series (LS) and Middle Series (MS). The oceanic succession comprises the picrite bearing Upper Series (US) and what seems to be fairly uniform, Icelandic-type tholeiites that are found at Sites 915 and 918. Leg 152 drill sites (914–919) and Leg 163 drill sites (989, 990; Duncan, Larsen, Allan, et al., 1996) are shown. The presence of steeply dipping pre-rift sediments at the bottom of Hole 917A and the indication from seismic data of pre-rift sediments being present below the breakup unconformity at the Site 989 area suggests that some rifting and sedimentation took place prior to volcanism and breakup (see also Fig. 6, and Vallier et al., this volume).

fluvial quartzitic sediment bed at the base of the lavas (for discussion of possible age and origin of these sediments, see also Vallier et al., this volume, and Larsen and Saunders, this volume). Except for the uppermost 1–2 m close to the fault zone, these sediments show a fairly constant dip over the 50-m interval cored (Larsen, Saunders, Clift, et al., 1994). This favors an interpretation of the steep dip as representing significant tectonic tilting of the sediments prior to volcanism rather than soft sediment deformation such as syndepositional slumping (see also Larsen and Saunders, this volume). Tectonic rotation of these strata to their present steep dip along one single set of faults is an unlikely interpretation, and we therefore suggest that at least one episode of early deformation preceded the faults we observe in the seismic data and in the core (Fig. 6). It is possible that similar pre-rift sediments, although with variable and much less dip, also occur up to 10 km landward of Site 917, where seismic data show a bedded nature of the upper crust to be present below the lavas and breakup unconformity (Fig. 4). However, Leg 163 (Site 989) failed to penetrate these reflectors and their nature is not known (Duncan, Larsen, Allan, et al., 1996).

The presence of pre-rift sediments at Site 917 in the context of other—albeit scarce—occurrences of pre-rift sediments along the southeast Greenland margin suggests that along the proto-rift a narrow basin formed prior to breakup and became inverted and tecton-

ized prior to initial volcanism. Pre-rift sediments exposed onshore farther north indicate filling of a rift-related basin from the late Early Cretaceous to the early Paleocene, after which it became flooded with basaltic lavas (Nielsen et al., 1981; M. Larsen, unpubl. DLC data). There are many striking similarities between the geological development of the two areas (see also Larsen and Saunders, this volume).

Volcanic Stratigraphy

The volcanic stratigraphy within the faulted and flexed zone has been broadly divided into two successions on the basis of the ship-board and shore-based studies of the cores from Sites 917, 915, and 918. We have named these the “continental” and “oceanic” successions. The former includes the picritic, basaltic, and dacitic lavas of the Lower and Middle Series from Site 917, many of which have been contaminated during ascent through continental lithosphere (Fitton et al., Chaps. 28 and 29, this volume). Upward and seaward, this succession is replaced by an oceanic succession with rapidly decreasing continental contamination (Site 917 Upper Series and Sites 915, 918 lavas; Fitton et al., Chaps. 28 and 29, this volume; Saunders et al., this volume). A thin sediment horizon marks the boundary between these two distinct successions. The oceanic succession makes up the main part of the SDRS and apparently was extruded directly

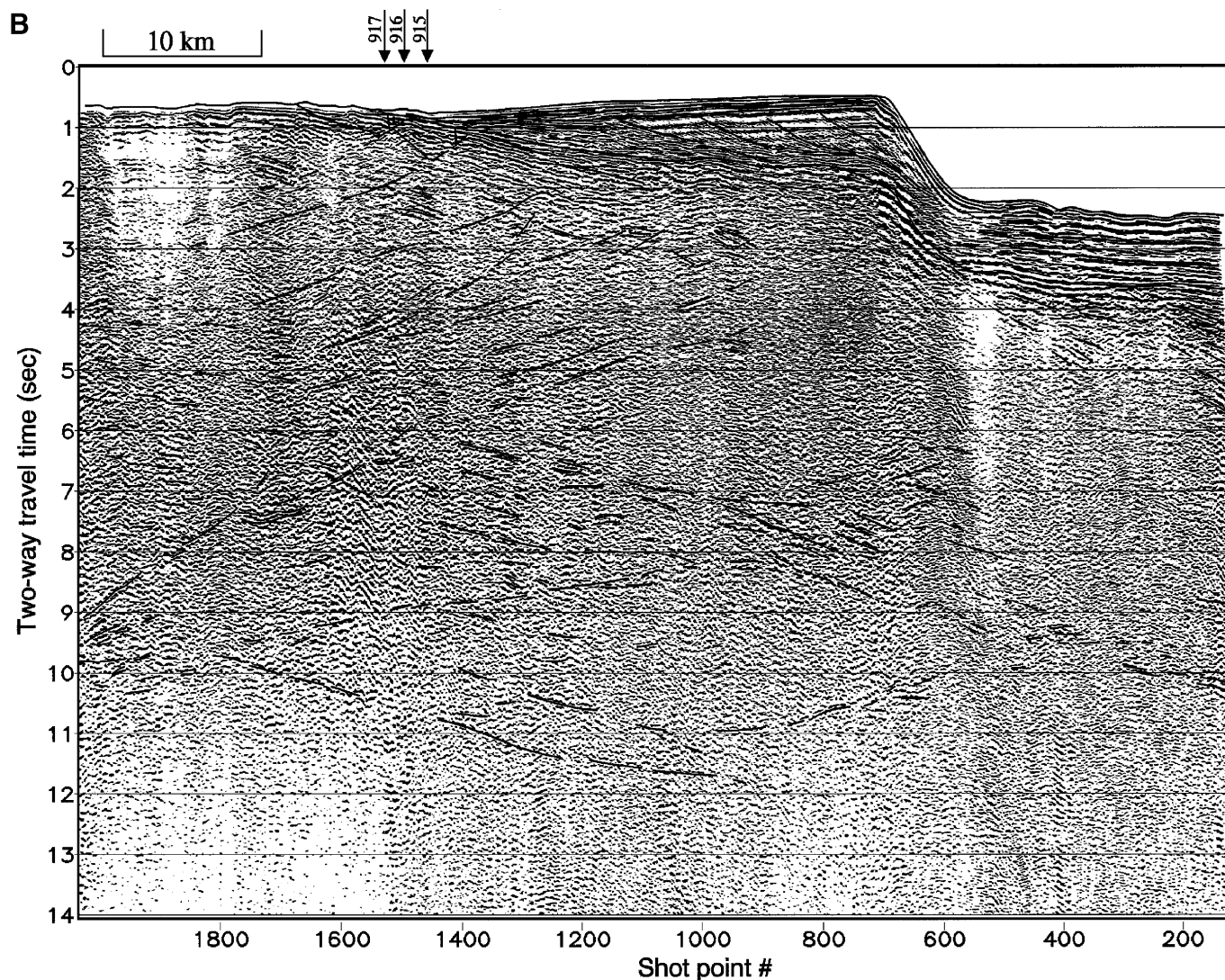


Figure 5. New deep multichannel seismic data collected along the Leg 152 drilling transect in 1994. Leg 152 Sites 915, 916, and 917 are shown along with Leg 163 Site 989. The data processing sequence is: trace mixing of same channels on adjacent shots, shot space wave-equation demultiple processing, CDP space frequency/wavenumber domain velocity filtering, NMO stacking using a carefully picked inner trace mute, post-stack minimum phase predictive deconvolution, coherency filtering, and tracing mixing along the stack. No automatic gain control is used, but a trace-to-trace amplitude equalization was applied prior to plotting. **A.** (foldout in back pocket of volume) Large-scale wiggle-trace/variable-area plot of the data. At the landward and seaward ends of the profile, the major crustal regions that are discussed in the text are marked. White dots outline deep reflectors that define Moho and possibly events in the upper mantle. **B.** Small-scale variable area plot of the same data with our interpreted line drawing superimposed. **C.** Depth-converted version of the line drawing shown in (B). Horizontal scale for both (B) and (C) is the same.

from an asthenospheric reservoir below a subaerial seafloor spreading center, akin to Iceland.

Transition Into Oceanic Crust

The implication of the observed lava stratigraphy is that formation of Icelandic type oceanic crust from a subaerial spreading center was established at the time of the extrusion of the lower part of the oceanic succession, and that this volcanic rift was located within the source region of these lavas. This provides us with the possibility to estimate the position of the continent-ocean boundary within the upper crust (i.e., the outer limit of the COT). Application of the kinematic model for crustal accretion in Iceland (Pálmason, 1986) to the southeast Greenland SDRS shows that lavas now located shallow in the crust stem from a fairly narrow rift zone 4–10 km wide and located around 25 km in the down-dip direction (Larsen and Jakobsdóttir, 1988). A

more distal origin would, for simple geometric reasons, require continuation of the lavas to unrealistically great depths (see Larsen and Jakobsdóttir, 1988, and Larsen and Saunders, this volume). The implication of this is that the early oceanic volcanism (Site 915 lavas) emanated from a rift zone located below the present-day outer shelf (Fig. 4). It also follows from this that if deep erosion of the lava cover has taken place at Site 915, we would have overestimated the distance to the eruption site. However, only limited erosion is indicated (Larsen, Saunders, Clift, et al., 1994), and we therefore consider the location of the source region (i.e., seaward end of COT in Fig. 4) to be fairly precise.

The original eruption site for the continental succession (Lower and Middle Series at Site 917) is not as well constrained. According to the Iceland model, the eruption site would be fairly close down-dip (because these lavas have a significant dip and represent a deeper crustal level). However, the continental succession is not necessarily

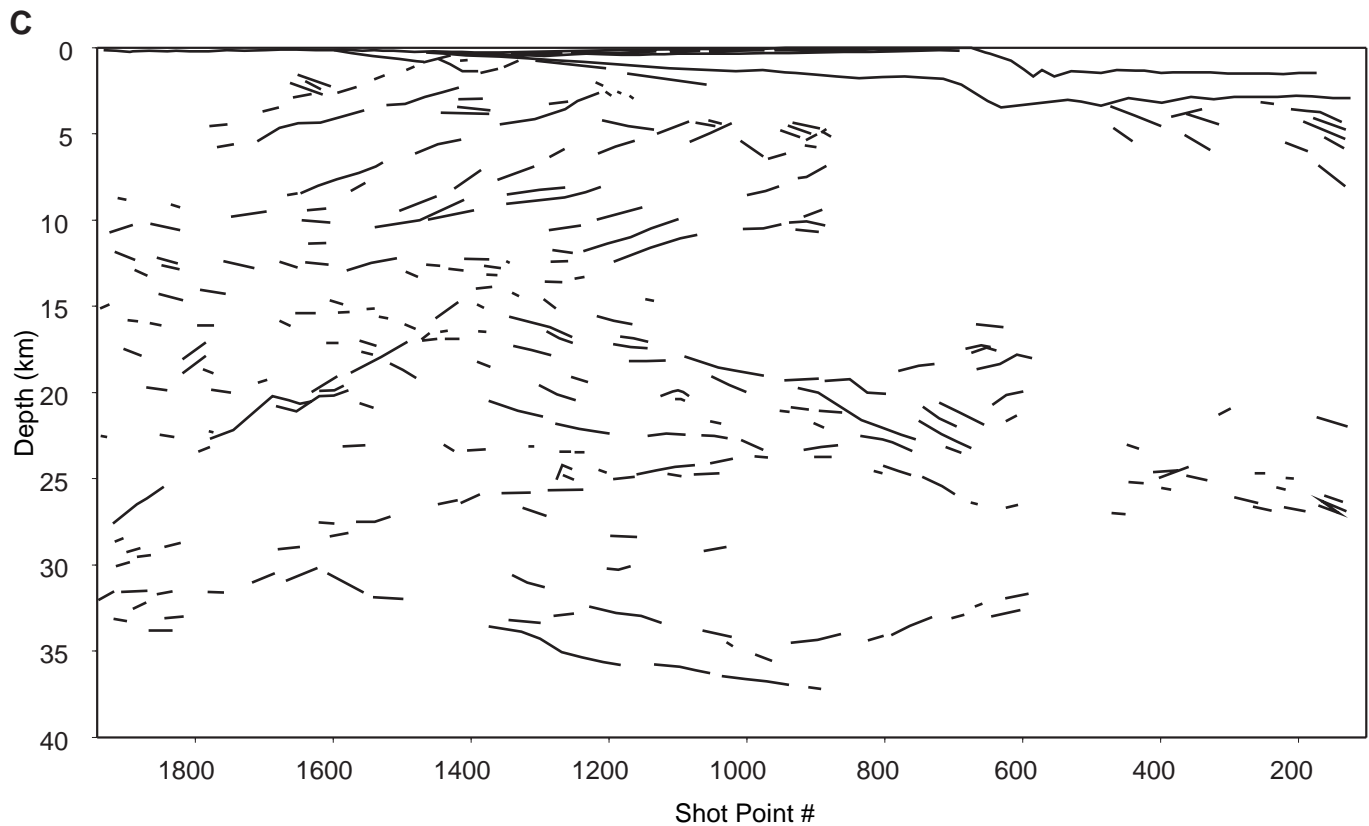


Figure 5 (continued).

part of the steady-state SDRS crust. Instead it could be part of the sheet cover that was tilted prior to the formation of, and later overlapped by, the main SDRS (see also Cambray, this volume, and Larsen and Saunders, this volume). This uncertainty with regard to the eruption site for the continental succession is illustrated by an on-shore, coast parallel dike that has a bulk composition and age similar to lavas from the continental succession (M. Storey, pers. comm., 1996).

The age of the continental suite is about 61 Ma (Chron C26r?; Sinton et al., 1994; Sinton and Duncan, this volume). This is between 4 and 5 m.y. older than the age indicated by magnetic anomalies in the oceanic succession of the SDRS (Chron C24r; 53–57 Ma). Therefore, a hiatus of approximately this length (or at least a highly condensed section) may be present between the continental succession and the onset of steady-state SDRS formation represented by the lower part of the oceanic succession (i.e., Site 917 Upper Series and Site 915 lavas). Alternatively, SDRS formation and spreading between the conjugate southeast Greenland and Rockall-Hatton Margins may have started significantly earlier than the generally accepted timing for the breakup of Greenland from Northwestern Europe (for further discussion, see Larsen and Saunders, this volume).

INTERPRETATION OF THE DEEP SEISMIC DATA

The deep 14-s data are shown at large scale in Figure 5A (back pocket). Our interpretation of the data is shown as a line drawing superimposed on a small-scale variable-area plot in Figure 5B. For additional reference, Figure 5C shows the same line drawing, only here it has been depth-converted using the velocities of Table 1. In general the signal-to-noise ratio is low, and most reflections are relatively weak and discontinuous. However, as discussed below, some first-order results emerge from the pattern defined by the line drawing.

The interpretation is further constrained by the shallow and conventional MCS data and drilling data, all of which have been transferred or projected onto the interpretation. In particular, the presence of typical SDRS below the outer shelf is not well imaged in Figure 5A (back pocket), but is based on a larger data set from Larsen and Jakobsdóttir (1988). Likewise, the interpretation of shallow crustal faults (drilled at Sites 916–917 and imaged by the high-resolution seismic data) is transferred to this interpretation. In addition, results from sonobuoy refraction and wide-angle experiments have been used to limit the possible range of the crustal thickness at both the seaward end and the landward end of the vertical incidence profile.

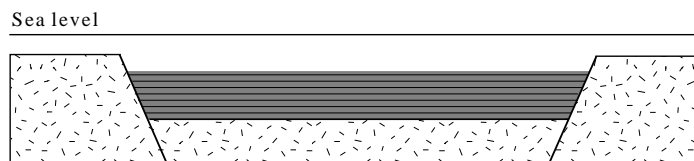
In the following sections we describe the interpretation of the main features of the data sets. From this interpretation a crustal model of the COT is developed. All times are reported as two-way travel-times, and the corresponding thickness and depth estimates are calculated using the velocities in Table 1.

Upper Crust

A group of curved, landward-dipping reflectors is present down to about 5–5.5 s (12–14 km) below the fault blocks within the feather edge of the SDRS (fault blocks projected from other sources). The curvature of these reflectors on seismic sections is reduced on the depth-converted line drawing (Fig. 5C) and may well represent more or less planar surfaces. We interpret these reflectors as the deeper parts of the normal faults imaged by the shallower data and drilled at Sites 916 and 917. Between these interpreted faults, some fairly strong, but discontinuous, linear and seaward-dipping reflectors occur at 1.5–2.5 s depth below the Site 917 area. These are clearly located deeper than the breakup unconformity as identified by drilling (Larsen, Saunders, Clift, et al., 1994), although they seem to have been rotated in a similar way. Older fault planes, such as those envis-

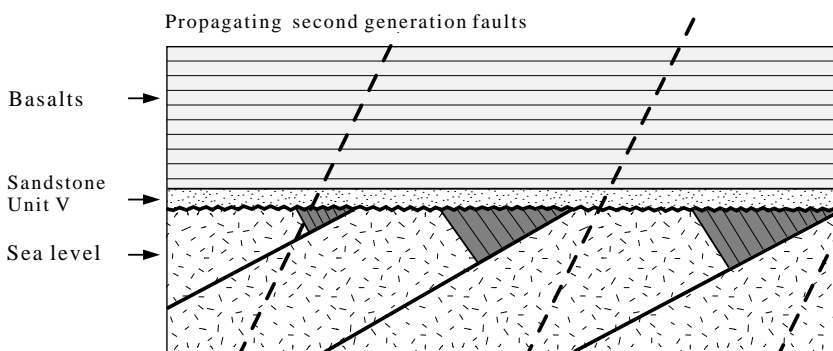
(1) Paleocene, Late Cretaceous ? - 62 Ma ?

Unit VI: marine shales deposited



(2) Paleocene; 62 - 56 Ma

Continental basement extended, uplifted and eroded.
Unit V fluvial sandstone deposited.
Basalts at the SDRS erupted over subaerial surface.



(3) Early Eocene; 50 Ma

Thermal cooling of continental margin, sag and faulting

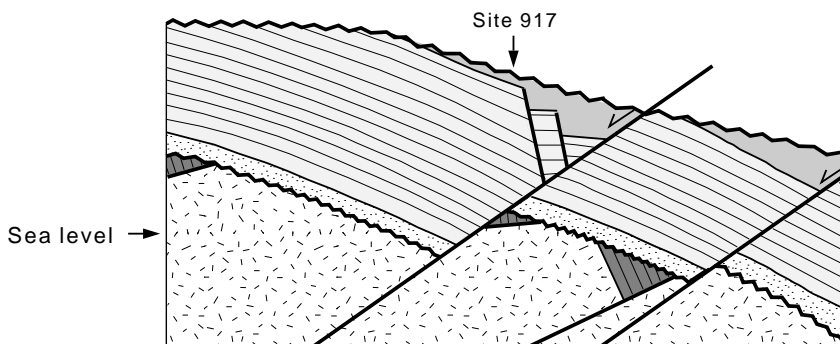


Figure 6. Proposed faulting history. At least two stages of faulting and deformation, including seaward rotation of beds, seem to be required to achieve the steep to sub-vertical dip of the pre-rift sediments at Site 917. If the landward-dipping reflectors below the basalts at Site 989 (Fig. 4) represent the meta-sediments drilled at Site 917, opposing dip directions are present. This could possibly be related to uplift and erosion prior to volcanism. Modified from Larsen, Saunders, Clift, et al., 1994.

aged in Figure 6, sill-like intrusions, or simply some entirely pre-rift (Precambrian?) structures are possible interpretations.

The upper crust that is characterized by landward-dipping reflectors (faults?) forms a wedge that thins and tapers in the seaward direction and eventually disappears below the outer shelf at around shot point 800. Possible reflections from the breakup unconformity are seen at a depth of 3 s (6 km) below the middle outer shelf around shot points 1000–800. The thinning of the continental wedge is coincident with a seaward thickening of the overlying feather edge of the SDRS, which attains a thickness of 2–2.5 s (4.5–6 km) below the outer shelf (shot point 720). In the deep water area below the shelf edge, the SDRS can be followed to similar depths. Larsen and Jakobsdóttir (1988) reported the SDRS to continue with similar or slightly increased thickness for about 80 km seaward of the profile (see also Fig. 3).

Middle Crust

At the landward end of the profile (shot point 600), a band of relatively continuous, seaward-dipping, mid-crustal reflectors can be traced from 5- to 6-s depth (13–14 km) to near the interpreted base of the crust (9- to 10-s depth, 22–23 km). Between this reflector band and the overlying wedge of upper continental crust and SDRS, an acoustically transparent region with poor signal-to-noise ratio is present. This nonreflective, mid-crustal zone is apparent from below the middle outer shelf and extends in the seaward direction below the lower continental slope and upper rise. One interpretation of this generally featureless zone is that it represents magmatic underplating; this is further discussed with the gravity modeling in a later section.

Within the landward end of the profile the band of mid-crustal reflectors may represent the poorly understood mid- to lower-crustal

reflectivity, which is often present in continental crust (e.g., Marillier et al., 1994). A well-developed band of such mid-crustal reflectors at approximately the same depth is reported from northeast Greenland (Larsen and Marcussen, 1992, and unpubl. data).

The apparent seaward continuation of the mid-crustal reflector band along the EG63 transect suggests a seaward continuation of continental crust at this mid- to lower-crustal level. Alternatively, the seaward part of this reflector band may relate to the formation of new igneous crust (i.e., magmatic underplating), a hypothesis that is discussed further below.

Lower Crust and Base of the Crust

We first discuss the terminology we use in this section. By base crust we mean the geophysically defined Moho and not the "Petrologic Moho" or other definitions. Likewise, by igneous or magmatic underplating we mean igneous material that has been added to the pre-existing continental crust. For convenience, we also refer to igneous underplating within the deep, thick crustal root of the entirely new igneous crust represented by the oceanic SDRS crust. In the latter case, this "underplated" material is merely just an expanded version of normal lower oceanic crust (layer 3), although it is known to have anomalously high velocities and probably also high densities (e.g., Holbrook and Kelemen, 1993; Kelemen and Holbrook, 1995). The latter is probably because of the excessive degree of melting that takes place during formation of the SDRS crust and drives the average composition of the generated magmatic liquids toward more basic, mantle-like compositions. The heavier fractions of this material never (or very rarely, e.g., the Site 917 picrites) rise to the surface, but form cumulate rocks, which are physically (seismic velocity and density) midway between typical lower crust material and normal mantle material. The mantle residue, from which magmatic melts have been derived, we call residual mantle, and this might well have undergone significant physical changes during melt extraction. We suppose that the geophysically defined Moho is located close to or slightly above the top of the residual mantle. This is because we see no mechanism for larger scale lowering of the seismic velocity of the mantle by melt extraction, but we could envisage that seismic velocities of cumulate rocks that form from settling in the melt column might closely approach mantle velocities.

In the following we mainly discuss the reflection seismic Moho, simply because that is what the data offer. We explain in each case how the reflection Moho relates to the concepts discussed above.

The base of the crust is defined on the basis of reflections from or around the Moho, and a downward decrease in reflectivity within the underlying mantle. Wide-angle data obtained on a coast parallel profile (DLC, unpublished) south of the Leg 152 transect suggest that the base of underplated igneous crust (i.e., the new Moho established during breakup) is little reflective and can be located considerably below the "old" continental Moho. We interpret reflection Moho to be present at about 10.5 s (31–32 km) toward the landward end of the profile and at around 10 s (21–22 km) toward the seaward end of the profile.

The landward Moho reflection at 10.5 s is the base of present-day continental crust. We have also marked a reflection at ~9.5 s that we interpret as the base of the original continental crust. Thus, the original crust is at present 27–28 km thick. If this interpretation is correct, then there is about 3–4 km of underplating below the most landward part of the profile and a total crustal thickness 31–32 km. The interpreted thickness of the continental crust in this position is consistent with preliminary interpretations of wide angle and vertical incidence data along a coast-parallel line south of the Leg 152 transect (DLC unpubl. data), and is also consistent with data from the southwest Greenland shelf showing that the crustal thickness of the Precambrian crust is close to 30 km (Chian and Loudon, 1992).

The deep reflection at 10 s below the seaward end of the profile we interpret as the base of the new igneous SDRS crust. Larsen and

Table 1. Model velocities and densities.

Crustal unit number	Seismic velocity (km/s)	Model density (g/cm ³) and interpretation Figure 7A	Model density (g/cm ³) and interpretation Figure 7B
1	1.5	1.0 (water)	1.0 (water)
2	2.0	2.1 (sediments)	2.1 (sediments)
3	5.6	2.8 (SDRS, dike zone)	2.8 (SRS, dike zone)
4	5.9	2.67 (upper CC)	2.67 (upper CC)
5	6.2	2.9 (lower OC)	2.9 (lower OC)
6	7.0	3.1 (underplating)	2.9 (lower OC)
7	7.0	3.0 (underplating)	2.9 (lower OC)
8	6.1	2.85 (middle CC)	2.67 (middle CC)
9	7.0	2.85 (lower CC)	2.85 (lower CC)
10	6.3	2.85 (middle CC)	2.9 (lower OC)
11	8.0	3.35 (mantle)	3.35 (mantle)
12	8.0	—	3.12 (underplating)

Notes: The seismic velocities shown were used for depth-converting the line drawing of the deep seismic section (Fig. 5C) and for calculating thicknesses and depths reported in the text. Velocities used for continental crust are close to those from Chian and Loudon (1992). The average seismic velocity of 5.6 km/s for the SDRS complex (including underlying dike zone; see main text for discussion) is significantly higher than what is observed at a shallow (0–1 km) crustal level within the boreholes where alteration of the lavas can reduce the velocity to lower than 4 km/s. However, sonobuoy data (Larsen and Jakobsdóttir, 1988) indicate significant increase in velocity with depth. The velocities for the lower oceanic crust (including underplated material) may be slightly underestimated. The densities are those used in the gravity models in Figure 7. OC = oceanic crust; CC = continental crust.

Jakobsdóttir (1988) showed that the outer part of the profile (Fig. 5) is underlain by crust that is at least 15 km thick. If the margin is symmetric with the conjugate Rockall-Hatton margin, then it may be as much as 20–22 km thick (Spence et al., 1989; Fowler et al., 1989). The interpreted reflection Moho at 10 s is in agreement with the 20–22 km range, and we consider this a fairly accurate estimate. If anything, it is perhaps on the low side because of the relatively low crustal velocities applied (Table 1).

There are two distinctly different possibilities for interpreting the reflection seismic Moho between the landward and seaward ends of the profile. Both interpretations include the same rapid thinning of the continental crust. The difference between the two interpretations is related to the deep reflections (10–12 s) below the central part of the profile. These reflections, which are indicated by white dots on Figure 5A (back pocket), define an elongate, pillow-shaped body approximately 40 km wide and 8–10 km thick, located directly beneath the upper crustal COT. One interpretation assumes that these events are reflections from the residual mantle. In this case a gradual seaward shallowing of the Moho with more moderate, though quite considerable, underplating is present (Fig. 5). In the other interpretation, this pillow-shaped body is part of the underplating, in which case the crust below the COT is as thick as 35 km because of excessive underplating. The one interpretation cannot be favored over the other without refined velocity control (acquisition of data planned for 1996), but in any case, a strong correlation between lower crustal/upper mantle features and the upper crustal COT exists (see also discussion in geological interpretation and gravity modeling).

A point of general concern is to what extent the reflection pattern might relate to old Precambrian structures rather than structures related to breakup. The general structural grain of the Precambrian terrane in this region is more or less parallel with the margin transect and the seismic line, and this geometry could result in complicated reflection patterns. However, within the continental crust, we only interpret the upper crustal, landward-dipping, normal faulting and the mid-crustal reflectivity. The orientation of the faults (upper part) is shown by shallow seismic data to be roughly margin-parallel, and coring data show that the faults were active during breakup (Larsen, Saunders, Clift, et al., 1994). The mid-crustal reflectivity may well be a Precambrian structure, and we do not infer anything different from this; the only assumption made here is that the structure is close to horizontal, which seems consistent with seismic data from a coast-parallel profile south of the present profile (DLC, unpubl. data).

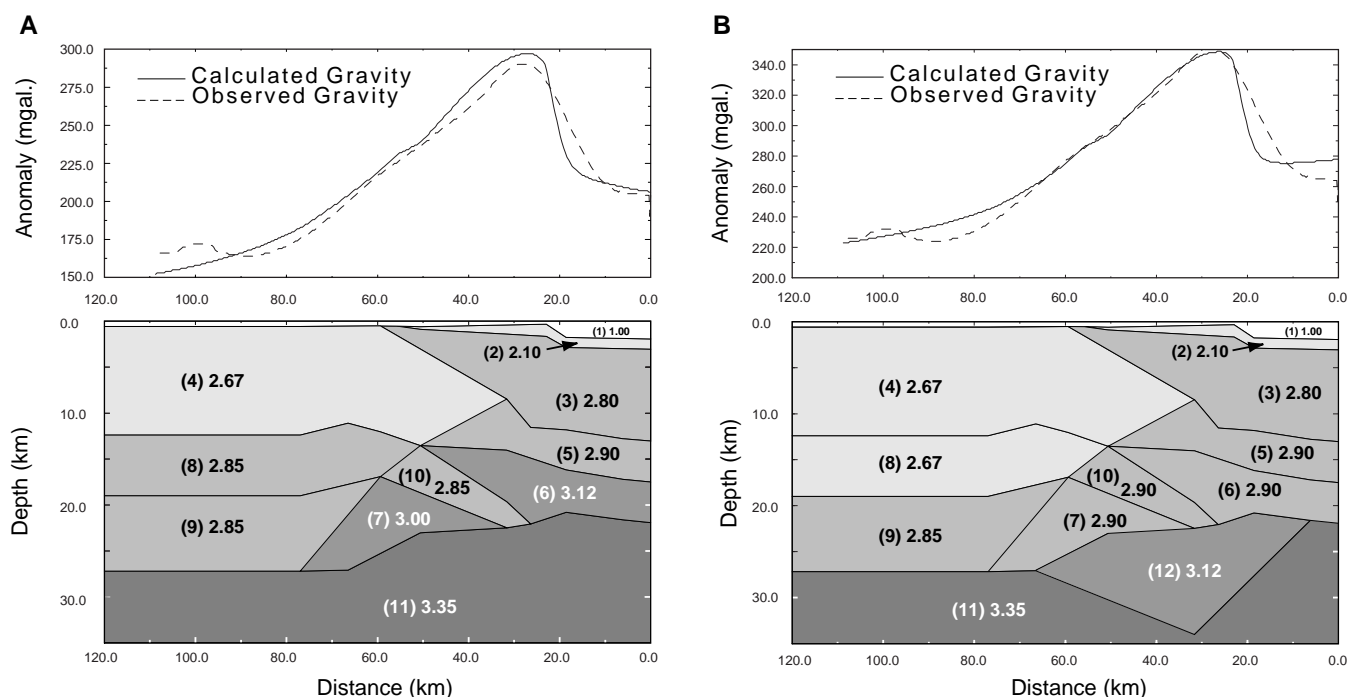


Figure 7. Two models of the crust along the inner part of the Leg 152 transect (area covered with deep crustal seismic data). The crust has been subdivided into crustal units numbered 1 through 12. The different crustal units are discussed in the text. Solid line = calculated free air gravity; dashed line = observed free air gravity. The seismic velocities used for depth conversion are given in Table 1. The difference in absolute gravity is due to the fact that the model in (A) is only calculated to 30 km depth, whereas the model in (B) extends to 35 km depth. Vertical exaggeration is 2:1 in the cross sections. Added in proof: New wide-angle seismic data (DLC, unpubl. data, 1997) favor model A and confirm its general outline.

GEOLOGICAL INTERPRETATION AND GRAVITY MODELING OF THE COT

The interpretation of the deep seismic data is summarized by two crustal models for the COT as shown in Figure 7. The models include a number of layers and subunits that can be interpreted various ways. Our preferred model is the one shown in Figure 7A, and we first discuss that one in detail. The primary difference between the two models is whether or not the pillow-shaped body described above is included as mantle material or as underplated material. The differences are discussed further below. The crustal models are partly constrained by the free-air gravity data in the area. Both of the models show a reasonable fit to the gravity anomaly.

In Figure 7A, the crustal units are numbered 1 (water) through 11 (mantle). We first discuss the upper crustal units (2, 3, 4) for which we have drilling control. Next we discuss the deeper crustal layers within the oceanic crust (5, 6), and then the lower crust below the continental margin (4, 7, 8, 9, 10). Crustal unit 2 comprises the outer shelf and deep basin sequence of post-rift sediments (see Clausen; Lykke-Andersen; Holmes; and Vallier et al., all this volume). Below these sediments, the SDRS and its source region are present (crustal unit 3). As pointed out by Larsen and Jakobsdóttir (1988), it is inherent in the Pálmason (1986) model for SDRS type crust that this volcanic cover is replaced downward by a sheeted dike complex. However, neither the model for crustal accretion in Iceland nor our observations constrain the thickness of this dike complex. Within the rift zone of northern Iceland, a magma chamber that is known to feed the fissure swarm within the active rift is located at a subsurface depth of 2–3 km. This is most likely the maximum thickness of the dike zone in northern Iceland. Thicknesses of dike complexes in ophiolites are typically less than this (1–1.5 km; e.g., Lippard et al., 1986). Thus, a conservative estimate of the thickness of the upper crust, which includes 5–6 km of lavas and 1–2 km of dike complexes, is 7 km.

The SDRS crust (upper and lower crust) represents the total melt layer extracted from the mantle. The melt layer is partitioned into an upper part (feeder dikes and lavas) and a lower, plutonic residue (lower gabbroic crust; units 5 and 6). Large variations in the ratio between these upper and lower crustal components (i.e., the ratio of layer 3 thickness over layer 2 thickness) can be expected (Mutter and Mutter, 1993). At Hole 504B, Detrick et al. (1994) report that the layer3/layer2 ratio is about 2.5. In magmatically starved, slow-spreading environments with little or no volcanic activity, the ratio is likely to be much higher. In our model, the crust is about 20–22 km thick and the ratio is between 1 and 2, which seems anomalously low. We have arbitrarily separated the lower oceanic crust into units 5 and 6 with increasing density at depth. We expect unit 5 to have physical properties (density and seismic velocity) close to normal oceanic crust, and layer 6 to be denser and have higher velocities similar to the conjugate Hatton margin (Spence et al., 1989; Fowler et al., 1989). Larsen and Jakobsdóttir (1988) report high seismic velocities (7.3 km/s) from a mid-crustal level that is shallower than that adopted here. However, their identification of a possible Moho at around 15-km depth may in fact correspond to the unit 5/6 boundary within the crust (i.e., the top of the high velocity layer).

Crustal unit 4 is the upper continental crust deformed by landward-dipping normal faults. Recovery at Site 917 of pre-rift sedimentary rocks within one of the tilted fault blocks indicates the former presence of a sediment basin, which was largely removed by uplift and erosion prior to volcanism. The feather edge of the SDRS originally extended farther inland across unit 4, but was removed by margin uplift and erosion (particularly glacial erosion; Larsen et al., 1994a; see also Larsen and Saunders, this volume). It is possible that Tertiary intrusives (dikes, sills, and plutons) are present within this crustal unit.

A continental origin of the seismically reflective mid- to lower-crustal unit 8 (the mid-crustal reflector band) is assumed. The origin

of this type of seismic layering and reflectivity is not known, though it could be speculated that it relates to horizontal flow accommodated extensional fault movements in the upper crust. It is also possible in this case that the layering could represent sill-type magmatic bodies emplaced during breakup. This band appears to continue seaward (unit 10) and caps a region of igneous underplating (unit 7). Landward, unit 7 is separated from the lower continental crust (unit 9) by a fairly steep zone of reflections; these may represent intrusive contacts between underplated material and the continental crust.

We also interpret crustal unit 9 as being of continental origin. Attempts to model this unit with high densities result in an excessive mass landward of the outer shelf maximum. However, the possibility that minor underplating is present below unit 9 cannot be rejected and is to some extent indicated by the seismic reflection data.

Figure 7B shows an alternative model that is also consistent with the observed gravity data. As already stated, the main difference here is that the pillow-shaped body (unit 12) is modeled with a density appropriate for igneous underplating instead of with mantle densities. To maintain the size and shape of the gravity anomaly, several changes to the crustal density structure were necessary. In particular, the middle crust (unit 8) now has a density like that of the upper crust rather than the lower crust, and units 6, 7, and 10 all have a density equal to that used for normal lower oceanic crust (unit 5). The reflectivity in unit 10 might therefore be related to laterally spreading intrusions in an igneous lower crust. However, elsewhere the igneous lower crust is little reflective, and such an interpretation is difficult to defend. Given the inherent uncertainties in estimating reasonable lower crustal densities in both continental crust and oceanic crust, Figure 7B shows that the gravity is insensitive to the different models. In all cases, including models not shown here, some excess density is required to explain the large amplitude gravity anomaly situated over the continent-ocean transition. What is not resolved, however, is the detailed distribution of density.

DISCUSSION AND CONCLUSIONS

The interpretation of the landward and central part of the deep seismic profile provides new information on the deep structure of the southeast Greenland COT. The interpretation of the seaward part of the profile over Icelandic-type oceanic crust is tenuous and in part depends on correlation with other areas. Also, it is clear from the modeling that the gravity field is insensitive to a number of important features we interpret from the reflection seismic data. These features are not resolvable without proper wide-angle seismic data to provide detailed crustal velocity information. Thus, the interpretations presented here are likely to undergo modifications following wide-angle seismic studies planned for 1996. Nonetheless, our interpretation of the new seismic data together with older seismic data, drilling data, and shore-based observations show the following five important features to be present within the southeast Greenland COT.

1. Landward-dipping normal faults soling out at or above the mid-crustal level; the array of these faults has produced a seaward thinning wedge of upper continental crust below the feather edge of the SDRS.
2. The mid-shelf crustal flexure or hinge line is closely matched by a rapid seaward thinning of the continental crust from a thickness of about 27–28 km below the landward part of the flexure. The continental crustal thinning is partly, perhaps even fully, compensated by emplacement of dense underplated igneous material at lower crustal levels. Prior to final breakup, the most landward part of underplated material may have formed an early, "intrusive" wedge immediately below the later hinge line.
3. A pillow-like zone, 8 to 10 km thick, is present at approximately 25–35 km depth below the central part of the COT. The na-

ture of this zone is not resolvable on the basis of existing data. It might represent dense underplated igneous material, or residual pre-rift mantle. In the first case, the central COT has a thicker crust (30–35 km) than elsewhere along the profile.

4. Seaward of the upper crustal hinge line, mid- to lower-crustal reflectors dip seaward and may indicate a former detachment between upper and lower crustal material. Igneous material eventually spread out along this region and completed breakup of the continental crust.
5. The COT is about 40 km wide measured from a point of negligible rifting and a continental crustal thickness of ~30 km to a point where the crust is entirely of igneous origin. Site 917 is situated very near the midpoint of the entire transition zone.

With regard to point 1, investigations along the coastal flexure zone farther north have revealed the existence of landward-dipping fault- and shear-zones (including pseudo-tachylites) within the Precambrian basement and its volcanic cover (Nielsen and Brooks, 1981; Karson et al., 1994). Massive, glassy fault gouges described as pseudo-tachylites were also observed in the faults offsetting the lavas at Site 917 (p. 130 in Larsen, Saunders, Clift, et al., 1994). Thus, there is strong evidence for normal faulting and associated extension within the upper crust along landward-dipping faults during breakup.

The landward dip of the faults and associated seaward rotation of fault blocks is opposite to what is generally associated with continental rifts and rifted margins (see Larsen and Saunders, this volume, for further discussion). The more landward faults root into the lower part of the upper continental crust at about the mid-crustal reflector zone, which may be a detachment layer and a transition into ductile deformation. The most seaward faults, in particular those close to the base of the outer continental wedge, may be partly rooted in gabbroic intrusions. Such intrusions are often seen in close association with the coastal flexure and fault zones (Myers, 1980; Myers et al., 1993; Karson et al., 1994). In line with these observations from coastal exposures, gabbro clasts of presumed Tertiary origin were found in the basal conglomerate overlying the lavas at Site 915 (Larsen, Saunders, Clift, et al., 1994). This suggests that shortly after volcanism, gabbroic material was exposed and eroded close to this site and the faulted zone.

The zone of upper crustal thinning across the COT is only about 25 km. This seems to be comparable to the Vøring margin offshore Norway, the Cuvier margin offshore northwest Australia, and the Carolina Trough of the U.S. East Coast (S. Planke, pers. comm., 1996; Hopper et al., 1992; and Holbrook et al., 1994; respectively). It is remarkable that seaward thinning of the lower continental crust has taken place directly beneath the thinning of the upper crust because crustal extension seems to be of a different nature in the upper and lower crust and is perhaps decoupled across the mid-crustal reflective zone. Nevertheless, the main area of extension seems confined to a roughly vertical column within the crust.

The interpreted distribution of extension is close to the geometry of crustal scale pure shear. However, the nature of the lower continental crustal thinning is not known. We suggest that, while the upper crust extended and thinned by faulting, the lower crust merely parted and extended around a subvertical zone, into which underplated material was emplaced at rates controlled by the extension. The continentally contaminated Lower and Middle Series (Site 917) may have a source region from such an early accumulation and emplacement of hot igneous material into the lower crust (see also Fitton et al., Chaps. 28 and 29, this volume). Also, such intrusive underplating could help the crust avoid complete collapse and would counteract tectonic subsidence within the rift zone. The latter is consistent with the subaerial nature of the volcanism.

The seaward boundary of the COT as defined by the deep crustal data is close to that defined by application of the kinematic model for spreading in Iceland to the stratigraphy of the SDRS (Figs. 4, 7; Larsen and Jakobsdóttir, 1988). In addition, this location of the COT is consistent with the volcanic development from the lower continental

lava succession to the oceanic lava succession within the landward drill Sites 917/915 (Fig. 4).

The possible presence of a partly detached mid-crustal lid of continental origin (crustal unit 10 in Fig. 7) is difficult to understand. Although it may be compatible with the volcanic history (Fig. 8), restoration of a balanced cross section across the original rift zone, including this peculiar feature, seems difficult. If our model is correct, it indicates a mid-crustal change in rheology and density that might have caused magma to accumulate and spread laterally at this level before final breakup. Such a development would be consistent with the starved nature of the upper part of the continental succession (Middle Series, Site 917) and a subsequent hiatus between the continental and oceanic succession. In support of this, we note that Chian and Loudon (1992) found that the Precambrian crust below the southwest Greenland shelf comprises an upper and lower part with significantly higher velocities (and densities?) within the lower crust.

Alternative interpretations of unit 10 naturally include the possibility of underplating; only detailed seismic velocity imaging of this zone (planned for 1996) can possibly resolve this issue. However, it seems unlikely that alternative possibilities would significantly change the location of the COT, though the details of the depth profiles remain uncertain.

Several questions can be posed: What is the nature of the continental crust within the outer, upper crustal wedge below the feather edge of the SDRs? How much of this upper crustal wedge is new igneous material intruded into a continental host rock? The coastal exposures north of the Leg 152 study area show a close association of seaward-rotated crust and synrift intrusion of dikes and gabbros. Within these coastal outcrops, the intensity of dike injection increases dramatically over a distance of only 10 km from a few scattered dikes inland to more than 50% Tertiary igneous rocks along the outer coast. Locally, the dike swarm develops into a sheeted complex with hardly any host rock left (Nielsen, 1978; Nielsen and Brooks, 1981; Myers, 1980; Karson et al., 1994).

At Site 917 only one dike was identified within the lavas and no dikes were found within the pre-rift meta-sediments. This seems to preclude massive dike injection in the area around Site 917. However, based on analogy with the coastal area, this does not exclude a significant seaward increase in igneous material within the thinned wedge of continental crust seaward of Site 917. Hence, it could be suggested that the outer part of the wedge-shaped continental crust includes significant amounts of igneous material similar to what is exposed along the coast farther north. In this regard, it might be important to note that within this outer continental wedge, mainly the fault planes—and not the fault blocks themselves—are seismically imaged. Although this could easily be a data problem (low signal-to-noise ratio), it could also reflect a blurring of the fault blocks (break-up unconformity) by dike injection continuing past, and hence stratigraphically higher, than initial volcanism.

We consider the pillow-shaped body, 8 to 10 km thick, at approximately 25–35 km depth directly beneath the COT a first-order observation, though its precise nature remains undetermined due to lack of deep seismic velocity control. Larsen and Marcussen (1992) reported strong reflectors from a similar to slightly deeper depth below the sill-intruded and previously flood-basalt-covered Jameson Land basin, which is located within the flood basalt area of the northeast Greenland margin. Larsen and Marcussen interpreted these deep reflectors as truly within the mantle, possibly caused by melting processes during breakup (i.e., residual mantle). A similar origin may apply to the deep pillow body identified below the COT in this study.

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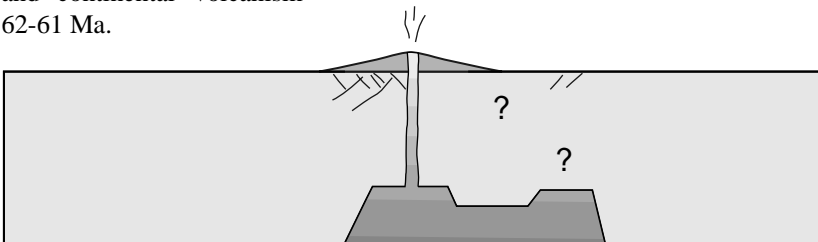
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Ms 152SR-245

1) Deposition of pre-rift sediments



2) First igneous underplating, stretching of upper crust, uplift, erosion and "continental" volcanism 62-61 Ma.



3) Spreading of igneous material at mid crustal level, final breakup and steady state "oceanic" volcanism.

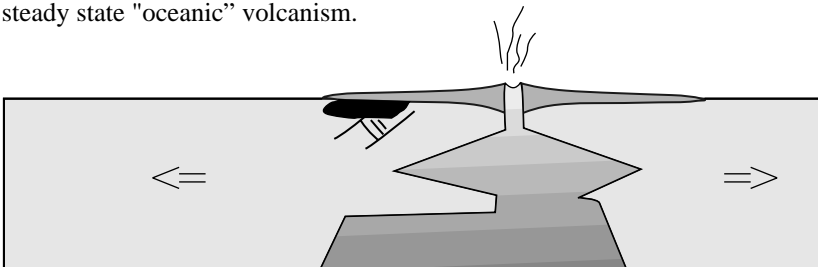


Figure 8. Possible three-stage development of the margin structure. A small amount of initial thinning of the crust is associated with pre-rift sedimentation. The age of this phase is not known, but similar developments on shore are of Late Cretaceous to early Paleocene age (Nielsen et al., 1981). The first magmatism follows a phase of larger scale crustal extension, thinning, and uplift (see also Fig. 6). This suggests that hot material has risen high into the lithosphere and possibly into the lower crust. The continentally contaminated Lower and Middle Series lavas (continental succession) are erupted, possibly through intra-crustal replenishment magma chambers (not shown; see also Fitton et al., Chap. 28, this volume). In the third phase, magmatic melts are penetrating higher and spread at mid-crustal level. The Upper Series (including picritic lavas) and the subsequent Icelandic tholeiites (oceanic succession) are then erupted from a shallow crustal pool of magma, and complete continental breakup is accomplished. The time from stage 2 to 3 may be as much as 3–5 m.y., but this is not constrained by precise radiometric age determinations. The model implies that thinning of the upper and lower continental crust is of different nature. The upper crustal extension and thinning is taken up along landward-dipping faults, while the lower crust may not thin at all, but simply part and allow "intrusive" underplating.