Erosion and deposition in interplain channels of the Maury Channel system, Northeast Atlantic

Turbidity currents
Thermohaline currents
Deep-sea erosion
Deep tow
Bottom photographs
Courants de turbidité
Courants de contour
Érosion profonde
Deep tow
Photographies du fond

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ABSTRACT

Large turbidity currents originating on the insular margin of southern Iceland have flowed down a 2 500 km-long pathway comprising rise valleys, unchanneled plains and segments of erosional and depositional deep-sea channels that are collectively called the Maury Channel system. Two steep interplain reaches of the channel have been cut up to 100 m through volcanogenic turbidites of probable Late Pleistocene age. Near-bottom observations with side-scan sonars and profilers across the upper channels (at 59°24'N, 18°50'W, 2 750 m depth) and at the lower interplain channel (around 56°23'N, 24°25'W, 3 340 m depth) defined their structure and morphology. The upper channels, and a tributary to the lower channel, start as broad, shallow depressions that deepen and narrow downstream. The lower channel has a pattern of anastomosing branches that probably evolved by headward extension of low-angle tributaries to the original sinuous channel, and its branches are at different stages of development. Several hundred bottom photographs show well-indurated rocks on channel walls and floors, with such flysch-like characteristics as cyclic graded bedding, clastic dikes, and syndepositional deformation. The lower-channel branches have been cut by turbidity currents with speeds of 5-12 m/sec., and combined discharges exceeding 1×10^6 m^3/sec. Bedrock erosion in and around the channels has proceeded by intense corrasion and fluid stressing, and is marked by such small-scale effects as rock polishing, fluting, pot-holing and ledge recession. Rockfalls have caused retreat of steep channel walls, and conglomerate or pebbly mudstone deposits suggest that debris flows have been locally active. Some coarse debris delivered by these processes and "clay balls" torn from semi-lithified outcrops remain in the channels, but the channel fill is generally thin, with a patchy veneer of pelagic mud that has accumulated since the last major turbidity current event. The surfaces of the unconsolidated sediment have been smoothed and lineated, or moulded into scour moats and occasional fields of ripples, by thermohaline currents.

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RÉSUMÉ

Érosion et sédimentation entre la Ride de Reykjanes et le Banc de Rockall (Maury Channel system)

D'importants courants de turbidité s'écoulant depuis la marge sud de l'Islande sur plus de 2 500 km forment des vallées, des plaines non chenalisées, et des chenaux d'érosion ou d'accrétion. L'ensemble de ces morphologies générées par des courants de turbidité a été appelé le « Maury Channel System ». Deux zones d'interfluve de forte pente que le chenal recoupe ont été érodées sur plus de 100 m par des turbidites volcanogéniques dont l'âge est probablement Pléistocène tardif. Des observations ont été réalisées près du fond à l'aide d'un sonar latéral et d'un écho-sondeur, dans la partie supérieure du système

(59°24'N, 18°50' W, 2 750 m de profondeur) et dans la plaine interchenaux inférieure (environ 56°23'N, 24°25' W, 3 340 m de profondeur). Ces observations permettent de définir la structure et la morphologie de ces zones.
Les chenaux de la zone supérieure, ainsi qu'un chenal distributaire de la zone inférieure débutent par de larges dépressions peu profondes qui se creusent et se rétrécissent vers l'aval.
Le « chenal » inférieur présente un réseau anastomosé de branches qui évoluent par l'extension frontale des chenaux distributifs de faible pente prolongeant le chenal sineux initial. Ces diverses branches sont plus ou moins évoluées.
Plusieurs centaines de photographies du fond illustrent très bien les sédiments étonnamment indurés qui affleurent sur les parois des chenaux et sur leur fond. Ces sédiments présentent des caractéristiques semblables au flysch (bancs granoclassés cycliques, dikes de matériaux clastiques et déformations synsédimentaires).
Les branches du « chenal » inférieur ont été incisées par des courants de turbidité dont la vitesse varie entre 5-12 m/s et dont la capacité de transport est de l'ordre de 10^6 m^3/s.
L'érosion du fond induré des chenaux et de leurs marges est due à une intense abrasion par les courants et à une forte pression des fluides. Elle se traduit à l'échelle macroscopique par un polissage des sédiments, par des figures de courant (flûtes), par des affouillements et par un recul des rebords du chenal.
Des chutes de sédiments affectent les parois des chenaux et les font reculer. La présence de conglomérats suggère que des écoulements viscoplastiques sont localement très actifs. Quelques débris grossiers formés par ce processus ainsi que des galets d'argiles provenant des affleurements semi-lithifiés sont observables dans le chenal. Mais, généralement le remplissage de ce dernier est constitué par des matériaux fins (lutites) que recouvrent par place, des boues pélagiques qui se sont déposées depuis le dernier « événement » turbiditique important. La surface des sédiments non consolidés a été lissée et striée de fines linéations, modelée par érosion ou encore structurée en rides par des courants de contour (thermohalins).

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INTRODUCTION

Maury Channel extends 2 500 km from the insular margin of Iceland to the vicinity of abyssal plains off France and Spain (Ruddiman *et al.*, 1972; Cherkis *et al.*, 1973). It has been considered the northeastern equivalent of the Northwest Atlantic Mid-Ocean Canyon, which meanders down narrow turbidity-current plains within provinces dominated by other depositional agents (Heezen *et al.*, 1969; Egloff, Johnson, 1975; Chough, Hesse, 1976). Our reevaluation of the channel's geography shows that it is a composite structure, with only 500 km of meandering, erosional-depositional channel. Other reaches are identified as interplain channels, defined by Laughton (1960, 1968) as the erosional and headward-extending gullies cut when initially unconfined turbidity currents accelerate down a steep passage between two abyssal plains.

We made near-bottom observations of two sections of interplain channels to examine the erosive effects of distal turbidity currents. The results of episodic cutting of channels and flute marks are well displayed, in contrast to many other deep-sea environments where bed scour is generally obscured immediately by deposition from the turbidity current's body and tail. As in many submarine canyons on continental margins, sediment transport through our two study areas in Maury Channel also proved to be influenced by less episodic bottom currents, of tidal and thermohaline origin, that we were able to measure directly.

METHODS OF STUDY

There have been many previous attempts to map the pattern, morphology and shallow structure of deep-sea channels. However, defining channel pattern is hampered by the need to interpolate between sounding lines, a problem that is solved only by using swath-mapping techniques (e. g., Kenyon *et al.*, 1978) or by having soundings dense enough "to allow contour representation without recourse to geologically or physiographically biased interpretation" (Johnson, Vogt, 1973, p. 3443), as have been collected by the US Naval Oceanographic Office over much of the North Atlantic (Johnson *et al.*, 1971). Complete definition of channel morphology is impossible with any surface-ship system because of the poor resolution of the steep slopes characteristic of channel walls; this problem can be mitigated by using narrow-beam sounders or by graphically correcting broad-beam records (Laughton, 1968), but it can be solved only be making near-bottom observations. Only a few bottom photographs and samples have been obtained from the walls and floors of deep-sea channels (e. g., Laughton, 1968; Chough, Hesse, 1976), because of the difficulty of dangling instruments into such narrow targets.

Our principal survey tool was the Marine Physical Laboratory's deep-tow instrument package (Spiess, Mudie, 1970; Spiess, Tyce, 1973), whose effectiveness for mapping details of turbidite channel morphology has been demonstrated by several previous surveys of deep-sea fans (e. g., Normark, Piper, 1969; Normark, 1978; Normark *et al.*, 1979). This vehicle was towed 10-200 m above the seafloor on a single crossing near the head of Maury Channel, and during a 100 km² survey that straddled the channel 500 km further downstream. The most useful systems were a 40 kHz narrow-beam (4°) echosounder, a 4 kHz subbottom profiler, a pair of 120 kHz side-scan sonars (mapping a 1 km-wide swath), and a stereo-pair of vertical-incidence cameras. On one lowering in the main survey area we added an experimental nephelometer that measured light from a strobe source, scattered back at 160-170° from a water volume 0.5-1.0 m in front of the vehicle; a filtering sediment pump (Mayer, Tyce, 1977) collected some of the suspended sediment responsible for the backscattering. At the main survey area we also made two hydrocasts into branches of the channel, using a Neil Brown CTD, calibrating Niskin bottles, and (on one cast) a Geosecs laser nephelometer. Savonius-rotor current meters were moored 20 m above the channel floor close to the site of each cast for 30 hours, and a nearby meter simultaneously monitored the current outside the channel.

The topographic setting of the tiny patches mapped by deep tow is provided by high quality bathymetry from the US Naval Oceanographic Office, presented in part by Johnson and Schneider (1969). We have supplemented 3.5 kHz and airgun data collected by our towing ship with published seismic profiles for a revised overview of the Maury Channel system.

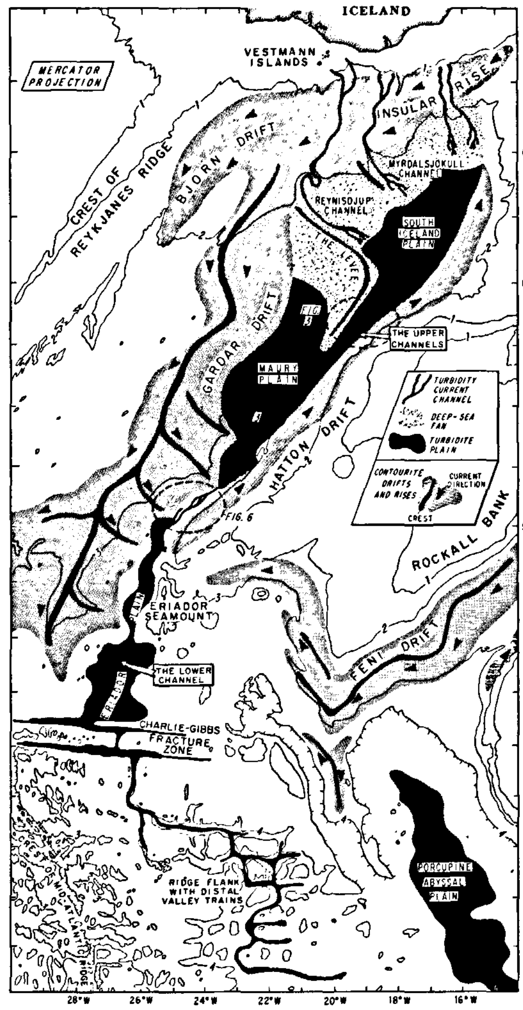


Figure 1

Physiography of entire Maury Channel system. Distributary pattern of fan valleys south of Iceland is schematic.

MAURY CHANNEL GEOGRAPHY

Some submarine density flows initiated by explosive volcanic eruptions and jokulhlaups on Iceland and the Vestmann Islands transport sediment up to 2 500 km to fault troughs on the east flank of the Mid-Atlantic Ridge. However, the name "Maury Channel" is applicable to only two stretches of this long pathway. The tributary channels that are incised into the Icelandic slope and rise (Egloff, Johnson, 1979) have other names (Fig. 1), and for much of their remaining course the turbidity currents cross unchannelled plains.

The upper channels of the Maury system occupy a 25 km wide interplain gap between the South Iceland Plain (a composite lower fan of the insular rise channels) and the similar Maury Plain. These plains have been divided by the growth of a large curving levee (Fig. 1), and the 200 km-long gap passes between the end of this levee and Hatton Drift on the southeast margin of the basin. Turbidity currents flowing through this narrow corridor increase speed down the steeper gradient there (Fig. 2), and have cut shallow channels. The channels cannot be traced far beyond the interplain gap, south of 59°N, but the downstream continuation of recent turbidity currents is marked by a 20-30 km-wide swath of unburied volcaniclastic sand (inappropriately called "Maury Mid-Ocean Canyon" by Laughton *et al.*, 1972) along the southeast margin of the plain, which elsewhere is veneered with acoustically transparent hemipelagic mud (Ruddiman *et al.*, 1972). Although the upper channels form an interplain channel network similar in origin to those described between abyssal plains (e. g., Laughton, 1960) the obstruction that led to flow acceleration and channel cutting is not the deeply buried bedrock relief but the huge depositional levee.

The lower Maury Channel, extending between 57°40′N and 52°20′N (Fig. 1), is also in part an interplain feature. The downstream end of Maury Plain has been obstructed by lateral growth of spurs from Gardar Drift, a contourite ridge that has been built since the middle Miocene by deposition from the nepheloid layer of the Iceland-Scotland Overflow current (McCave *et al.*, 1980). From 56°30′N to 56°N this drift approaches within 25 km of Hatton Drift, leaving a narrow turbidite corridor in which the lower channel has been incised. Unlike the upper channels this deeper interplain channel has extended upstream for 150 km across Maury Plain,

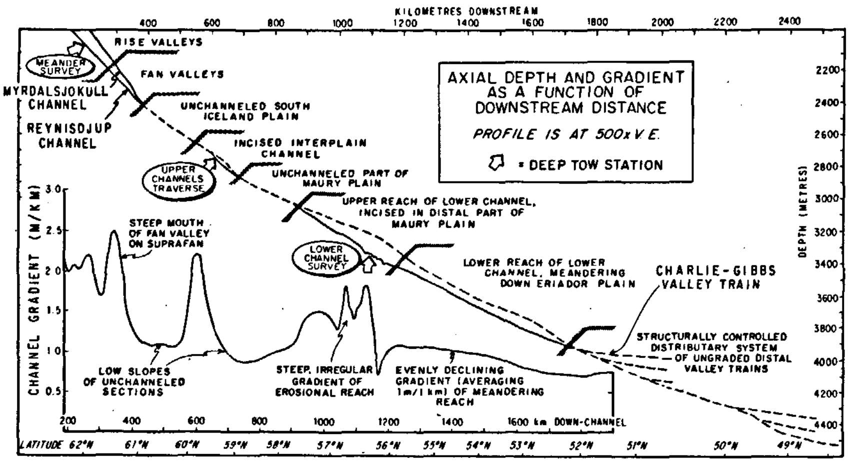


Figure 2

Long profile of the turbidity currents' route south from the Icelandic insular rise. Dashed lines indicate unchanneled sections, or surface of turbidite plains adjacent to channels. Channel gradient in inset plot was calculated for each 20 m of fall; local deviations like the slope reversals in the lower channel survey are smoothed out.

presumably by headward erosion, and at its downstream end it changes into a partly depositional leveed feature (Cherkis *et al.*, 1973). In the constricted reach, where we examined it with a deep tow survey, bifurcation and rejoining of channel branches forms an anastomosing pattern. The lower reach, winding across Eriador Plain (Fig. 1), was not studied closely but we infer that it is a classic erosional-depositional deep-sea channel like the Northwest Atlantic Mid-Ocean Channel. The best indication of concurrent erosion and deposition is the regularly sinuous pattern of the lower reach. For 300 km downstream (to 54°N) it has a sinuosity index of about 1.35, and thus would be considered meandering in some classification schemes (Rust, 1978). The meanders have a mean wavelength of 75 km and an amplitude of 30 km. From 54°N to the end of the channel near 52°30′N the sinuosity decreases and is less regular, apparently because the thalweg impinges on some seamounts (Johnson, Schneider, 1969).

At 52°20′N and 51°50′N, Maury Channel turbidity currents that have been meandering over the thick sediment of Eriador Plain cross the double trace of the Charlie-Gibbs Fracture Zone onto crust only half as old, where sedimentation has not yet buried the ridge flank's bedrock relief. Turbidites are dispersed along the 20-30 km wide troughs formed by intersecting abyssal-hill and fracture-zone fault scarps (Johnson, Vogt, 1973), forming ribbons of highly reflective sediment that we call valley trains. They were misleadingly called "channels" by Ruddiman *et al.* (1972) and Cherkis *et al.* (1973), but are merely narrow abyssal-plain fingers confined by basement ridges, similar to the distal extensions of many other abyssal plains into the Mid-Atlantic Ridge (e. g., Heezen *et al.*, 1964). They appear unchanneled on narrow-beam echograms.

TRAVERSE OF THE UPPER CHANNELS

Morphology and shallow structure

The several interplain channels notching the gap that connects the South Iceland Plain with Maury Plain start near a depth of 2 740 m, where the axial gradient steepens abruptly from 1 to 2 m/km. Regional bathymetry (Fig. 3) indicates that the channels narrow downstream from shallow bowl-shaped heads. This interpretation is supported by observations along the deep tow track, which crossed a single 5 km-wide channel (Fig. 4) just downstream from its head.

The deep-tow traverse was on the line of a Vema 2804 seismic profile which shows the reflective turbidites of the interplain gap contrasting with the finer sediments of the levee and of Hatton Drift. Even our near-bottom 4-kHz profiler achieved no acoustic penetration in the gap. As at Deep Sea Drilling Site 115 in Maury Plain (Laughton *et al.*, 1972), high reflectivity is probably caused by lithification of the Pleistocene turbidites, as well as by their coarseness. Bottom photographs and side-scan sonographs show a blanket of mud on most of the seabed outside the channel, but a few lithified outcrops of fluted sedimentary rock project through the sediment veneer (Fig. 5 A) in narrow strips parallel to the inferred direction of turbidity current flow.

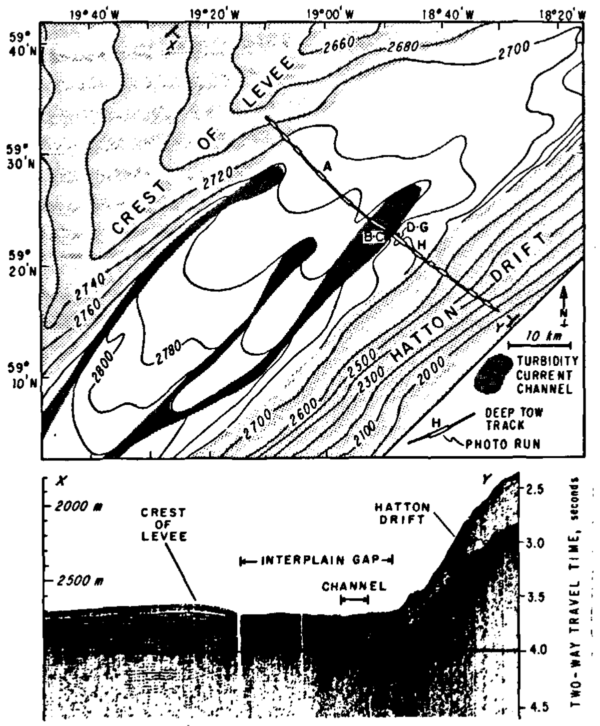


Figure 3

Location of the deep tow traverse across the upper channels' interplain gap. Letters A-H locate bottom photographs of Figure 5. Airgun profile X-Y is a Vema 2804 record (Ewing et al., 1974).

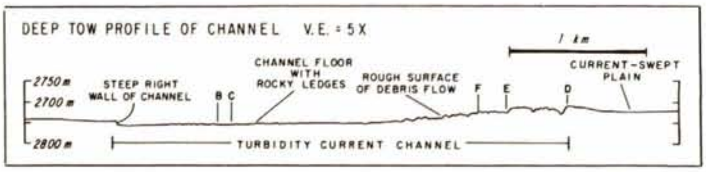


Figure 4

A high-resolution (40 *kHz*) *deep tow profile across one of the channels, located in Figure* 3. *Same vertical exaggeration as in Figure* 8. *Letters locate bottom photos of Figure* 5.

The channel we crossed has a highly asymmetric cross-section (Fig. 4). The northwest wall is a rocky step only 5 m high, with a succession of smaller parallel ledges above it. The valley floor at its base has many exposed strips of scoured, thin-bedded rock (Fig. 5 B, C), all trending southwest. In contrast to the down-current lineation of the right side of the channel, the southeastern side has a rough, uneven topography, and a channel wall whose plan shape is irregular and arcuate. At the top of the wall, a thin lithified layer resembling a hard-pan is being undercut and has collapsed to form angular, platy debris (Fig. 5 D). Deeper stratigraphic horizons out-cropping lower in the wall are fairly massive sandstones with visible evidence of coarse graded bedding (Fig. 5 E). Their pitted surfaces have been scoured into rounded and streamlined shapes and are free of sessile organisms. At the foot of the wall are rough piles of rounded sandstone (Fig. 5 F), with both residual masses and detached boulders. The adjacent channel floor, in a strip almost 1 km wide, is also covered with ill-sorted rounded boulders, up to 1 m in diameter, partly buried with fine mud (Fig. 5 G). The seabed appears identical to the surfaces of deep-sea debris flows that have been photographed in other, base-of-slope, environments (e. g., Lonsdale, 1978, Fig. 8), and the local channel conglomerate superficially resembles the pebbly and bouldery mudstones, also interpreted as debris flow deposits, in the axes of some ancient and modern submarine canyons (Stanley *et al.*, 1978; Stanley, 1974; Ryan *et al.*, 1978). However, our lack of samples prevents an estimate of the proportion of muddy matrix, and the Maury Channel deposit could be a disorganized conglomerate of Walker and Mutti's (1973) Turbidite Facies A.

Evidence for bottom currents

No direct current measurements were made on this traverse. Along the southeast margin of the interplain gap, Hatton Drift (Fig. 3) is swept by a northeasterly thermohaline flow fast enough (>18 cm/sec.) to maintain winnowed beds of rippled sand and lag gravel at depths below 2 150 m (McCave *et al.*, 1980), and a field of transverse sand waves encroaches on the edge of the turbidite plain. Side-scan sonographs (McCave *et al.*, 1980, Fig. 6) and photographs (Fig. 5 H) show that the steep down-stream slip-faces of the waves and their superimposed ripples have migrated northeast. The substrate of the waves, and most of the strip of plain traversed between the drift and the channel, is a non-depositional surface (see Fig. 5 D) with scour around glacial dropstones and other rock outcrops also indicating a current from the southwest, i. e., opposite to the presumed direction of downchannel turbidity current flow.

Within the channel there is little photographic evidence that steady thermohaline currents have much geological effect, though at the time of our traverse some animals attached to rock outcrops were being deflected northeast,

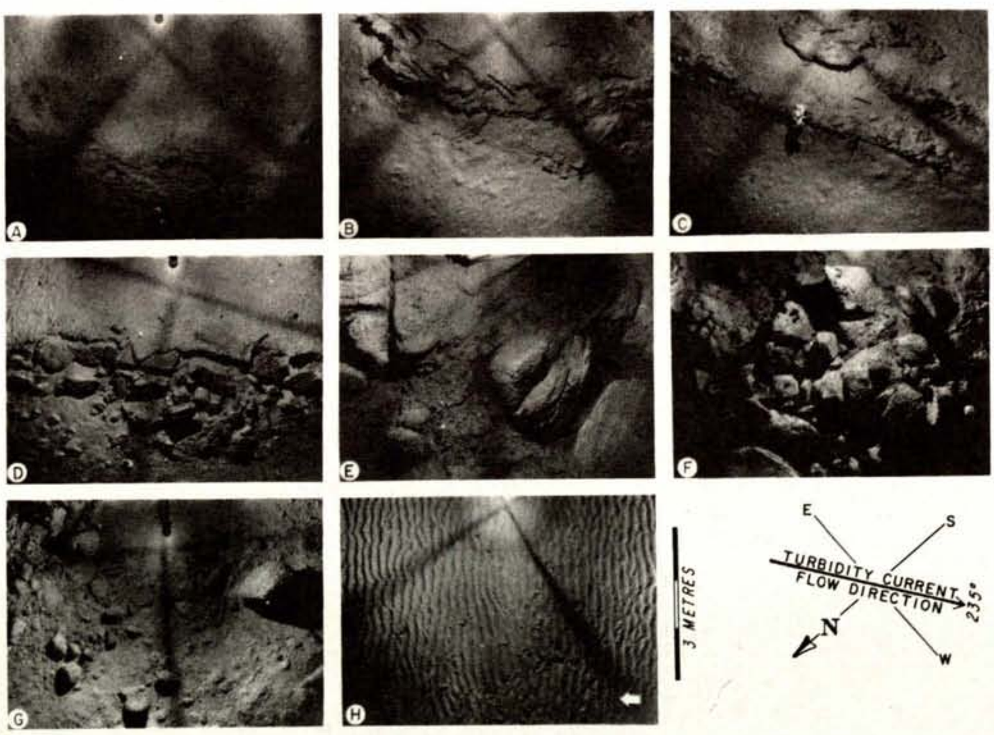


Figure 5

Bottom photos taken on the deep tow traverse of the upper interplain gap, located in Figures 3 and 4 and described in the text. Frame A is from an unchannelled part of the turbidite plain, B-G are from the floor and left wall of a channel, and H shows part of a rippled sand wave on the plain beside Hatton Drift.

up-channel (Fig. 5 B). Further to the northwest the bottom water was murkier (e. g., Fig. 5 A), presumably because of increased load of suspended sediment, and asymmetric scour crescents around rare rock fragments on the levee indicated a prevailing southerly current.

DEEP TOW SURVEY OF THE LOWER CHANNEL

Morphology and shallow structure

The turbidite corridor that links Maury and Eriador Plains is as wide as the interplain gap used by the upper channels, but the effects of channelized turbidity currents are confined to a 6 km-wide belt along its northwest margin beside Gardar Drift. Our near-bottom survey (Fig. 6) covered a 13 km length of this belt near the upstream end of the anastomosing reach. Within the survey area the principal (southeast) branch of Maury Channel is a rectangular trough 80-100 m deep and 1 500 m wide. Its slightly sinuous course is continued upstream by the similar northeast branch, but the channel thalweg passes into the southeast branch from a northwest branch by a deep, narrow slot through the intervening "island". There are two of these north-south links: the eastern one is a deep through-channel, while the western link is partly obstructed near its northern end, and could be described as a short, steep (Fig. 7) tributary gully to the southeast branch. The southwest branch is a major channel that starts as a shallow cut just downstream from the end of the much deeper northwest branch. For a 5 km reach between profiles 3 and 6 (Fig. 8) it has a steep average gradient (7.5 m/km) and a very rough floor, but in the southwest corner of our survey (profile 7) it resembles the southeast branch in depth and cross-section. These two subparallel branches eventually join some 20 km downstream (Fig. 6, inset), having isolated a 3-4 km wide "middle ground" or "island". In the deep tow survey area both branches have gentle bends (minimum radius of curvature 7.5 km) that are convex to the southeast. Near the other limit of its bend the southeast branch received another tributary, a broad but shallow channel that has its source on the

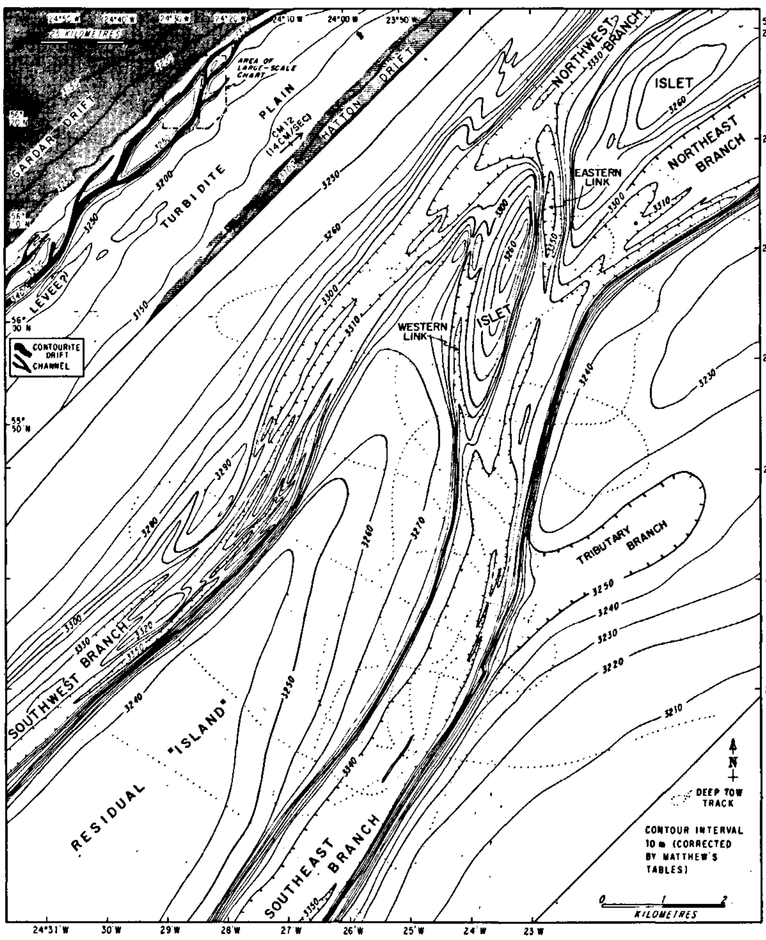


Figure 6

Bathymetry of the lower channel survey area. The large-scale chart relies mainly on deep tow data, supplemented by some transponder-navigated surface-ship profiles at the margins of the survey. The regional inset (top left) is based on narrow-beam surface-ship bathymetry.

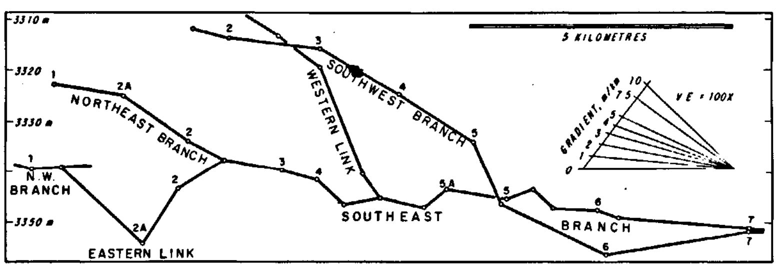


Figure 7

Thalwegs of channel branches in the deep tow survey area. Data points (dots) show the maximum depth at each deep tow crossing; numbered points locate the crossings of Figure 8.

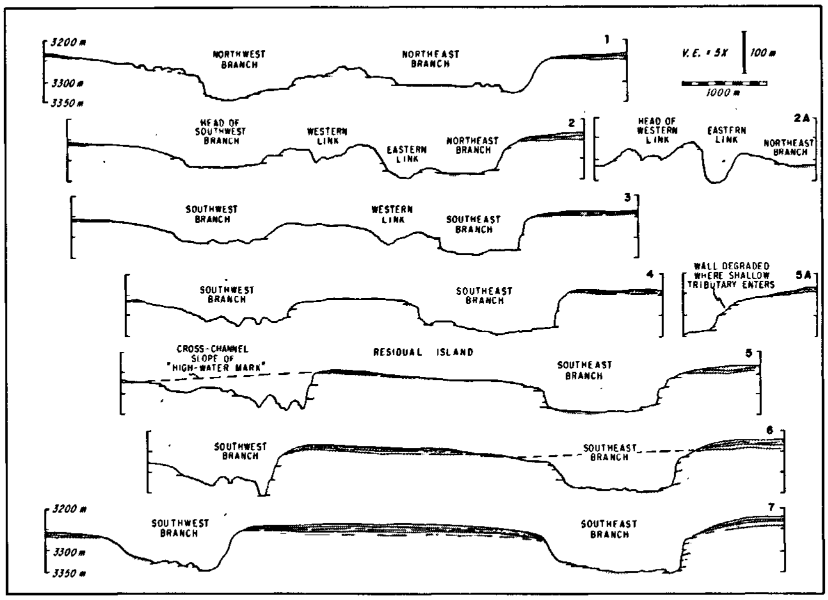


Figure 8

Transverse profiles of Maury Channel in its anastomosing reach, projected normal to the flow path from the tracks shown in Figure 9. The profiles use data from deep tow's narrow-beam 40 kHz echo-sounder for the bathymetry, and from the 4 kHz profiler for the limited information on subbottom structure. Dashed lines above profiles 5 and 6 show the slope of the upper interface of erosional turbidity currents, as inferred from the extent of scour.

adjacent plain and ends discordantly, hanging 60 m above the floor of the main branch.

Surface-ship seismic and acoustic profilers achieve poor results in the channeled region, but their records suggest a substrate of flat-lying strata. Near-bottom 4 kHz profiles also show limited acoustic penetration in most of the area, though on some crossings (Fig. 8) the steep channel walls are seen to truncate horizontal reflectors. Deep tow photographs and sonographs reveal a major reason for the poor penetration: much of the seabed within the channels is indurated sedimentary rock. Lithified outcrops also extend beyond the channels (Fig. 9), most notably over part or all of the surfaces of the interchannel islands and islets, where rocky slopes show that these features are erosional residuals rather than depositional mid-channel bars. Outside of the channel branches only semi-lithified material is exposed, forming low-relief outcrops identifiable on side-scan sonar records (e. g., Fig. 10) as patches of high reflectivity, and on bottom photographs (e. g.,

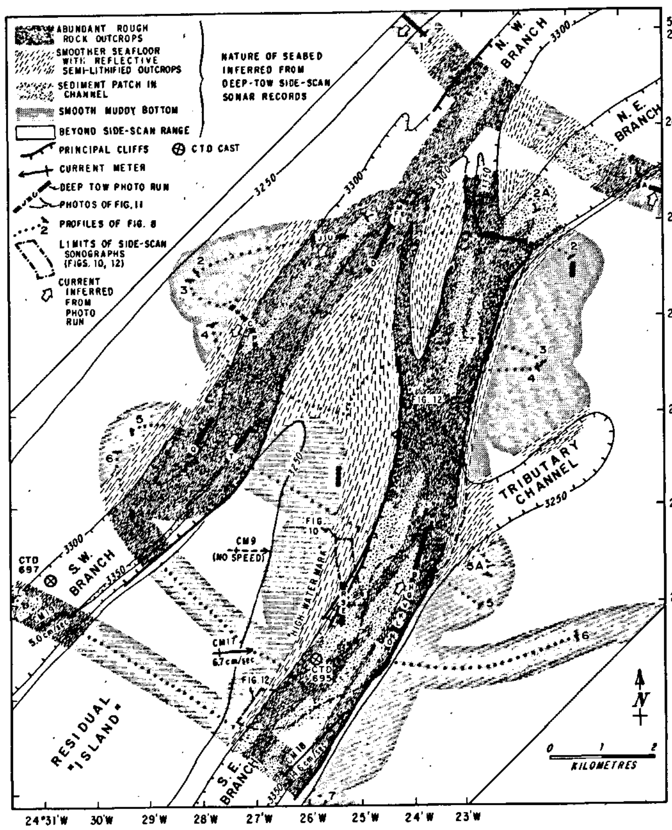


Figure 9

Station locations and character of sea-bed in the anastomosing reach of Maury Channel. Length of current meter arrows is proportional to measured velocity; Meter CM 9 measured direction only, for 3 weeks, and its dashed arrow indicates modal flow direction.

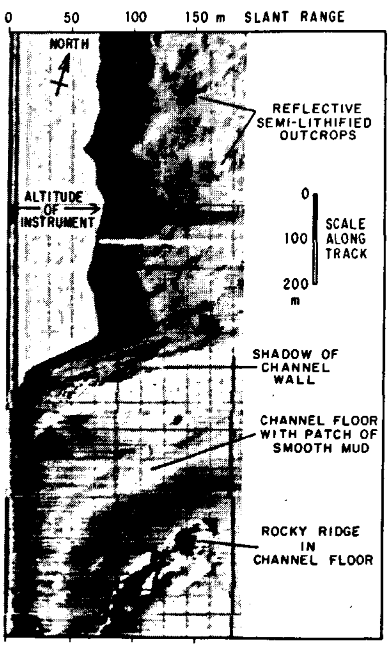
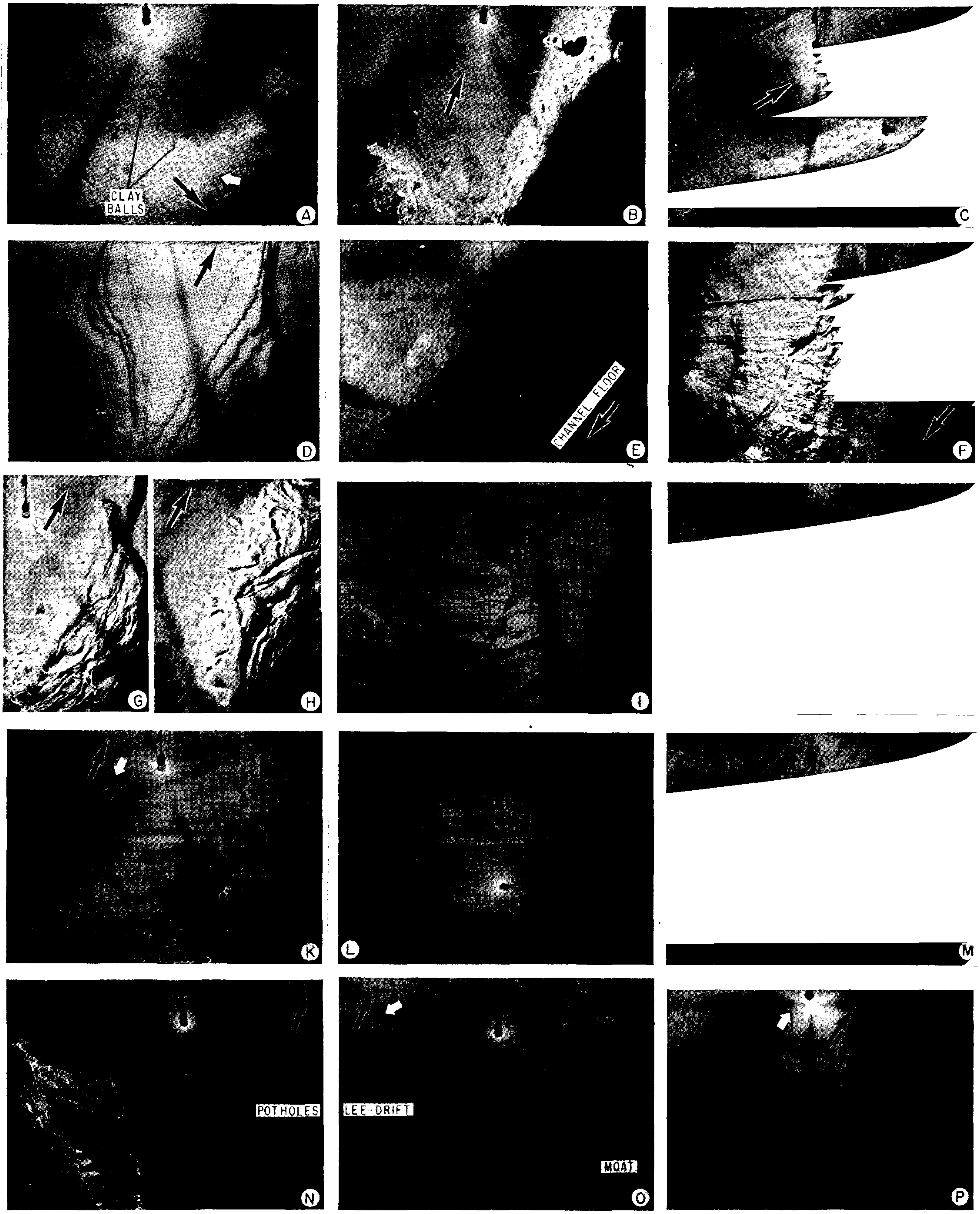


Figure 10

A sample of side-scan sonar record, located in Figure 9, that crosses the right wall of the southeast branch, and some of the superficially scoured zone above it. Near the bottom of this image the altitude of the instrument was drastically reduced (to 10 m) while photographs (e. g., Fig. 11 R, S) were taken.


CLAY BALLS
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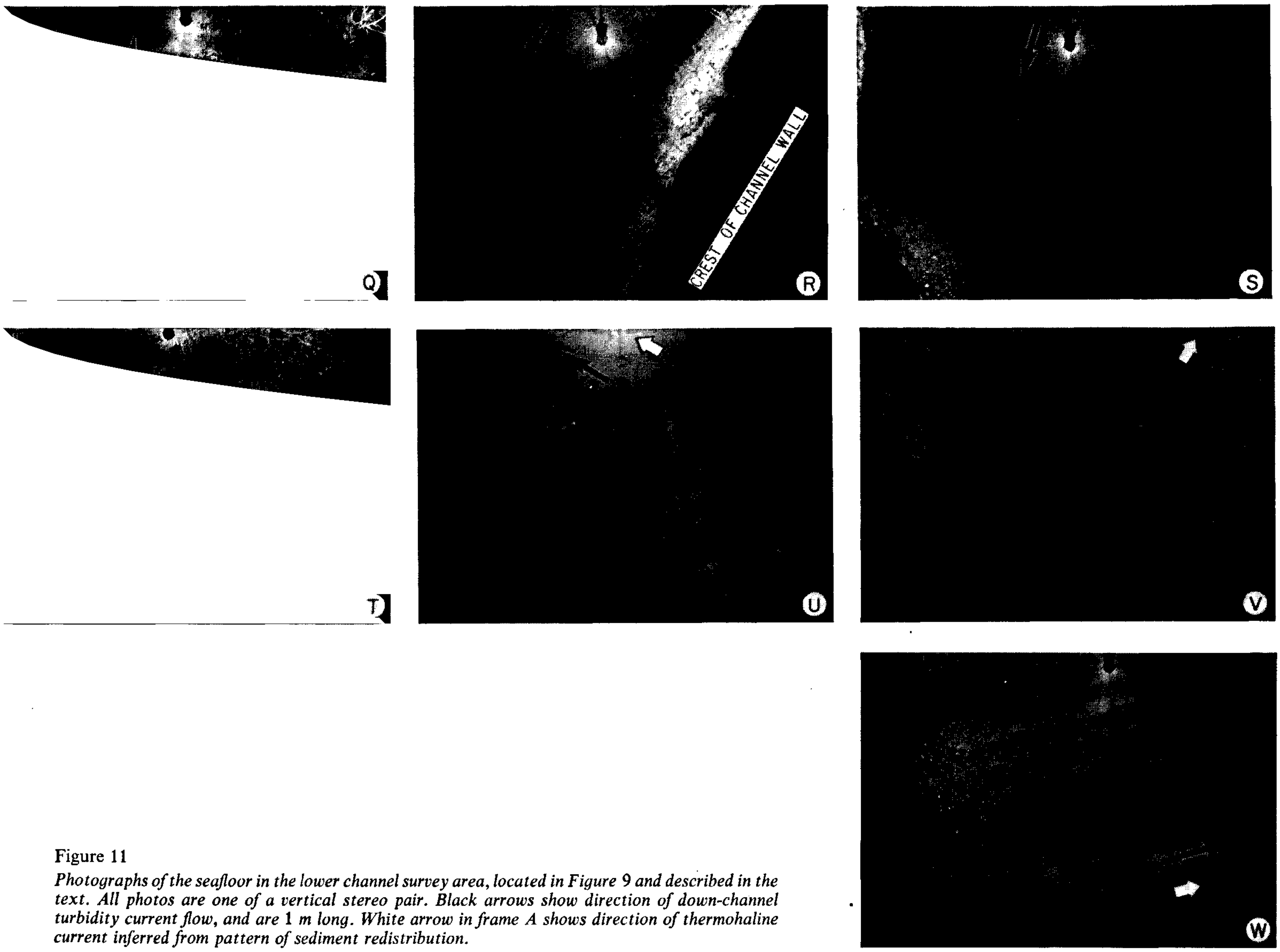

Figure 11
Photographs of the seafloor in the lower channel survey area, located in Figure 9 and described in the text. All photos are one of a vertical stereo pair. Black arrows show direction of down-channel turbidity current flow, and are 1 m long. White arrow in frame A shows direction of thermohaline current inferred from pattern of sediment redistribution.

Fig. 11 A) as areas free of unconsolidated sediment. Where deeper stratigraphic levels are truncated by the channel floor and walls (Fig. 11 B, C), more indurated sedimentary rocks form rough outcrops with steep irregular slopes. An exceptional outcrop of similar rock on the interchannel island occurs at its rough, deeply gullied upstream end (e. g., Fig. 11 D).

The relief of the channel floors, except in the shallow tributary branch, is accented by resistant rock ridges and ledges that are aligned, with little apparent structural control, parallel to the channel walls. Streamlined parallel ridges are evident on the bathymetric chart of the rough, steep reach of the southwest branch (Fig. 6). Similar landforms, generally less than 8 m in height and 100 m in length, are well displayed on side-scan sonar records (Figs. 10, 12) of the southeast branch. In both branches these roughness elements are superimposed on a cross-channel gradient, down towards the outside of the channel bends. Asymmetric excavation of the northwest and northeast branches (Fig. 8, profile 1) probably has the same relationship to channel curvature, but the plan of these branches was inadequately mapped with the limited deep tow coverage.

Some of the walls of the branches, especially at the outside of channel bends, have precipitous cliffs developed on nearly flat-lying sedimentary strata. Our near-bottom sonar records and a few photographs (e. g., Fig. 11 E) indicate that some sections tens of metres high are sheer, or perhaps even significantly undercut like the wall of some submarine canyons (Palmer, 1977). Most of the cliffed walls have a stepped profile, and an overall slope of 60-80° that diminishes to less than 20° on weaker, younger strata near the top of the truncated section. Their feet make an abrupt angle with the channel floors, with no ramps indicative of a substantial accumulation of talus. Cliffs are poorly developed on the left side of the southeast branch where flows down the shallow tributary have entered the main channel. At this junction a kilometre-long section of the steep wall has been degraded to an irregular, gentler scarp (profile 5 A, Fig. 8) with an uneven, scalloped plan (Fig. 12). Less steep channel walls elsewhere include the sides of the northeast branch, where even the high left wall has a smooth convexo-concave profile and a mean slope of only 15° (profile 1), and the right side of the highly asymmetric southwest branch.

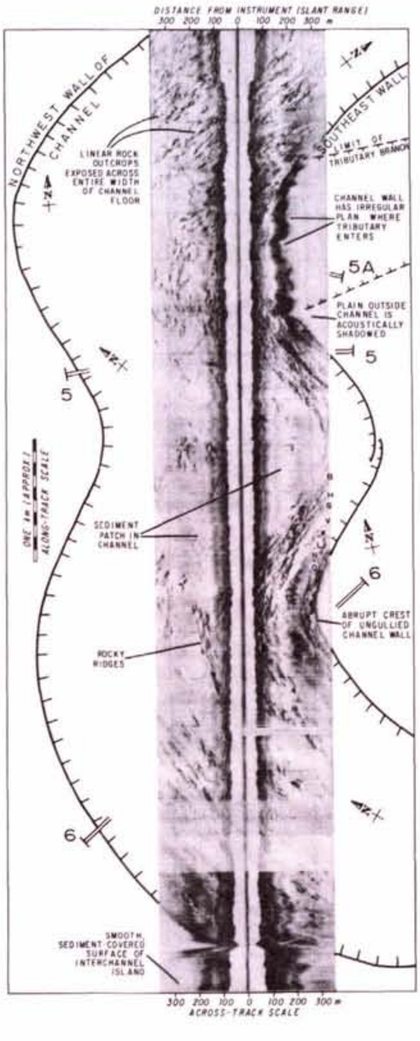

Figure 12

Side-scan sonar image of part of the southeast branch, located on Figure 9. The sonograph is unrectified: across-track and along-track scales are different, causing some shape distorsion. Note also that most of the apparent sinuosity of the channel walls is caused by the winding track of the deep tow instrument. Letters locate bottom photographs of Figure 11, taken on another track across the channel. Numbers 5 A, 5 and 6 locate transverse profiles of Figure 8.

Nature of rock outcrops

We lack samples of the exposed sedimentary rock, but more than 500 stereo pairs of bottom photographs (selected frames in Figure 11) show features characteristic of turbidite flysch sequences (Dzulynski, Walton, 1965). The channel branches have probably been incised through 100 m of muddy and sandy volcaniclastic turbidites similar to the Late Pleistocene sequence cored at DSDP Site 115 in Maury Plain (Laughton *et al.*, 1972). Our vertical-incidence photos usually show the stripped or obliquely truncated surfaces of beds, but the equivalent of a vertical section is seen in a few photos of large displaced blocks, as in Figure 11 F, where a thin-bedded clast displays channel-fill cross-bedding. Most of the observed bedding is planar, but some photos (e. g., Fig. 11 G, H) show contortions that were probably created by synsedimentary folding as a rapidly accumulated sequence suffered partial liquefaction. Other evidence for post-depositional mobilisation of sediment is the pattern of thin (1-5 cm) clastic dikes that penetrates some thin-bedded sequences at a high angle (45-70°) to the stratification (Fig. 11 I, 11 F). Their greater resistance causes these joint-fillings to stand out several centimetres above the scoured rock surface as narrow ridges, which intersect similar ridges, parallel to the bedding, that are either *in situ* source beds or clastic sills. Similar injection structures in flysch sequences are generally made of liquified sand (e. g., Dzulynski, Radomski, 1956); we infer that the projecting ridges and, by extrapolation, the cyclic micro-ledges of other photographs (e. g., Fig. 11 D) are sandy layers. Their enhanced resistance to erosion, compared to the intervening muddy beds, is attributable to the early cementation characteristic of immature volcaniclastic sands, as at DSDP Site 115 where hard sandstone layers within softer clays are common in the top 100 m of the section (Laughton *et al.*, 1972). Rapid indurating diagenesis of sandstone beds is a prerequisite for such surprising features of the surveyed erosional terrain as the formation and preservation of cliffed channel walls, bedrock ridges, and large joint blocks (Fig. 11 E) in young deep-sea strata that have never been deeply buried.

One common feature of flysch, the presence of sole markings, cannot be recognised, presumably because modern scour has bared the tops of resistant sandy beds and not the bottoms where flute moulds would be expected. However the formation of transverse scour marks similar to those described from the soles of turbidites has accompanied the recent incision of Maury Channel by erosive turbidity currents, as in the downcutting of bedrock channels of some rivers with a large load of abrasive sand (Maxson, Campbell, 1935). The most regular patterns of flute marks, only partly infilled by unconsolidated sediment, are found in homogenous mudstone, where we photographed (Fig. 11 J, K) fields of elongate-symmetric flutes in diagonal assemblages, and bulbous flutes. Some fields of flutes extend without interruption on rock surfaces that obliquely truncate the bedding, proving that they are not exhumed fossil features that were cut in unconsolidated mud during the period of turbidite accumulation. Fluted surfaces are not necessarily those most exposed to erosion: on some longitudinal channel-floor ridges that are small enough to fit within a single photograph (e. g., Fig. 11 L) it can be seen that flutes preferentially occur towards the tapering downstream ends, while the steeper upstream noses of the ridges have undercut ledges that may recede too fast for flute preservation. The orientation of all photographed flutes is parallel to the channel walls except near obstacles in the channel, where there may be rapid spatial changes in the orientation of flutes within a field (Fig. 11 M), marking local deformation of the scouring flow. Similar changes are found in the fluted surfaces of large river-bed boulders (Maxson, Campbell, 1935).

Potholes are another characteristic feature of river beds with resistant bedrock that is corraded by a sand and gravel load, and we photographed several potholes in the level, resistant stratum that floors the southeast branch of Maury Channel. Deep enclosed depressions have been discovered in the floors of some submarine canyons

(Ryan *et al.*, 1978), but are of the plunge-pool type, located below very steep sections of the thalweg. The larger, elliptical pothole shown in Figure 11 N is 4 m in maximum diameter and 3-4 m deep, with a vertical wall and only a partial cover of (light-coloured) sediment on its floor. We cannot determine whether the potholes have been drilled into uniform rock, or result from differential erosion of weaker parts of a heterogeneous unit. Near some of the potholes are equally large upstanding bedrock masses, apparently exposed by removal of weaker strata that once surrounded them (Fig. 11 O); they may be large sandstone pillows, common deformation structures in flysch sequences.

Channel fills of unconsolidated sediment

There is little sediment fill in the bedrock channels. The side-scan sonar records (e. g., Fig. 12) and photographs (e. g., Fig. 11 P) show extensive patches of light-coloured mud, burrowed and tracked by organisms, but only in the northeast branch do near-bottom profiler records show sediment more than 2 m thick. Sediment patches elsewhere are not ponded, but occur on sloping surfaces and as veneers on the crests of rocky ridges (e. g., Fig. 11 Q), indicating that they were not emplaced by turbidity currents.

From the lower Maury Channel about 200 km upstream of our survey area, Ruddiman (1972) cored "scoured balls of clay several centimeters in diameter" inferred to have been torn from the channel wall and shaped as they were rolled down the channel by a turbidity current. We photographed similar rounded clasts, up to 15 cm in diameter, near the semi-lithified outcrops high in the section beside the channel (e. g., Fig. 11 A). The most obvious coarse debris within the channel branches is local ill-sorted and often angular talus at the feet of some cliffed channel walls (Fig. 11 R, S) and some steep mid-channel ridges (Fig. 11 T). Though some of these boulders have been smoothed by scour, there is little evidence that they have been entrained by any flow. A few large exotic and obviously overturned clasts (e. g., Fig. 11 U) are found at sites in the middle of a channel that are remote from cliffs, though quite near similarly displaced fragments of the same lithology (Fig. 11 F).

50 km downstream, near the end of the steep, anastomosing erosional reach, the lower channel is partly filled with coarse, black volcanic sand that has been brought from Iceland by turbidity currents (Ruddiman, 1972, Cores K714-6, 7, 9). The only clear evidence of similar material in the reach we photographed is a small field of sharp-creasted black ripples in the lee of the large displaced clasts at the head of the southwest branch (Fig. 11 U). These ripples have avalanche slopes that face down-channel, so they could have been moulded by the most recent turbidity current to pass down the southwest branch. However, many of the patches of finer pelagic muds in this branch have also been affected by a southwesterly current, giving their surfaces (e. g., Fig. 11 P) the subtle lineation characteristic of muddy bottoms swept by more gentle thermohaline currents. In the southeast branch it is easier to differentiate the effects of thermohaline currents, for lineated mud surfaces, scour of unconsolidated sediment around rocky outcrops (e. g., Fig. 11 O), and strongly deflected sessile organisms (e. g., Fig. 11 K) indicate an up-channel current that could not have been a density flow. The only other photographed field of ripples (Fig. 11 V) is close to the southeast wall of this branch, where acceleration of a northeasterly current would be expected against a steep boundary.

Outside the channels our near-bottom 4 kHz profiler recorded unlithified sediment thickening gradually away from non-depositional feather edges (Fig. 8). We infer that accumulation of this unconsolidated cover has coincided with cutting of the adjacent channels. Photographs of the muddy surface of the residual island in the middle of the channelled belt, where the cover thickens to more than 30 m, show no bottom-current evidence. However, at both outer margins of the belt, where semi-lithified turbidites crop out through thin outliers of muddy sediment, prominent scour crescents indicate significant and opposite sediment transport by currents: toward the northeast on the southeast margin (e. g., Fig. 11 A), to the southwest on the northwest (e. g., Fig. 11 W).

AGENTS OF SEDIMENT TRANSPORT IN THE INTERPLAIN CHANNELS

Turbidity currents

Turbulent density flows with a heavy suspended load have been the most effective agents shaping the channels. Their capacity for doing geologic work in each reach depends on the composition and concentration of the sediment load that enters from upstream, and their velocity. The composition is known to be dominated by sand-size basaltic ash and lithic fragments, from the lithology of turbidites deposited downstream (Ruddiman, 1972). The volume concentration of sandy sediment is likely to be less than 9 %, giving flow densities of 1.1 to 1.2 g/cm^3 (Komar, 1977). The currents' velocities depend on the channels' gradients and roughness, and flow thicknesses. The extent and thickness of the largest flows down the lower channel can be estimated from the distribution of outcrops exposed by turbidity current scour (Fig. 9). From their observed effects in the survey area it is clear that many erosive flows have been more than bankfull. The "high water mark" of noticeable scour extends across the interchannel islands and part of the adjacent plain, and defines a surface that was neither level nor uniformly sloping, but in the centre of the survey was about 75 m above the adjacent channel floor. This underestimates the total thickness of the turbidity current, because the "high-water mark" defines a threshold where the current was thick enough to erode cohesive sediment. As in the flooding of an alluvial valley, there must have been thinner, slower flow over much of the adjacent plain, which drained back into the channel by shallow tributaries like the one at 56°22′N (Fig. 9). The upper limit of scour in this tributary (100 m above the adjacent channel floor) gives a better estimate of flow thickness.

If the flow thickness and channel geometry are known, the mean velocity of steady flow can be estimated with an appropriately modified Chezy equation (Hurley, 1964; Komar, 1977). This calculation requires an estimate of the drag coefficient; for the relatively smooth-floored southeast branch of the anastomosing channel we took the value of 0.0035 that has been used for smooth depositional deep-sea channels and for erosional subaerial channels in bedrock (see discussion in Komar, 1979, Appendix A). This method yields mean velocities of 5-7 m/sec. for 100 m-thick flows of density 1.1-1.2 g/cm^3 down the southeast branch.

Komar (1969) estimated bankfull velocity and discharge in leveed depositional channels by an alternative method that uses indirect measurements of the cross-flow slope of a turbidity current's surface. The same procedure was followed in the erosional channels, taking cross-channel differences in the depth of the "high water mark" of scour to define the tilt of the currents' surfaces. The advantage of this method is that no assumption of drag coefficient or measurement of hydraulic radius is necessary, so it is more easily applicable to flow down the rough, irregular bed of the southwest branch. For flow in the right bends of the southwest and southeast branches the mean velocity (u) is given by a rearrangement of Komar's (1969) equation 2:

$$\frac{u^2}{R} = 2\Omega \sin\Phi\, u + g\left(\frac{\rho_t - \rho}{\rho_t}\right)\tan S,$$

in which R, radius of curvature of channel (7.5 km); $2\Omega \sin\Phi$, Coriolis parameter; g, gravitational acceleration; ρ_t, flow density (estimated as 1.10-1.20 g/cm^3); ρ, density of seawater (1.03 g/cm^3).

The cross-flow slope (S) of the interface was 1 : 100 at profile 5 in the southwest branch (Fig. 8) where the down-channel axial gradient is about 10 m/km (Fig. 7), and 1 : 200 at profile 6 in the less steep southeast branch. With these values, calculated flow velocities and discharges through the section of the southwest branch are 7.3-10.7 m/sec. and 0.51-0.75 $\times 10^6$ m^3/sec. Comparable estimates for the southeast branch (5.3-7,7 m/sec., 0.56-0.81 $\times 10^6$ m^3/sec.) are similar to those derived from the Chezy formula. According to these numerical results the southwest branch, by virtue of its steeper gradient, is able to discharge 48 % of the total flow through a channel that (at profile 5) has only two-thirds the cross-section of the southeast branch. Downchannel at profile 7 (Fig. 8), where similar gradients and shapes of the two branches should cause similar current velocities, their cross-sectional areas are more nearly equal. Upchannel, continuity of flow dictates high velocities (8-12 m/sec.) down the shallow slot at the head of the southwest branch (Fig. 8, profiles 2 and 3), even though the gradient there is slight (Fig. 7); this suggests that the 30 m-high sill between the northwest and southwest branches may have acted as a weir, where channelized flow was thinned and accelerated (cf. Lonsdale, 1977).

Flows with mean speeds of 5-12 m/sec. are certainly powerful enough to sweep away any fine sediment deposited in the channels from thermohaline currents or lesser turbidity currents. Predicting the largest clasts that could be transported is hampered by the limited experimental or observational data on the competence of turbulent floods. Conventional analysis by estimating steady shear stress and using a Shields threshold criterion (Komar, 1970; Miller *et al.*, 1977) suggests that the maximum size of boulder that would be entrained by a steady 12 m/sec. flow of 1.2 g/cm^3 is just less than 1 m in diameter. However, empirical data that have been applied to subaerial flood paleohydrology (e. g., Malde, 1968) show that 12 m/sec. flows of clean water can move boulders several metres in diameter. Our observations in the lower channels suggest that the thick flows there were unable to remove some of the larger talus clasts (e. g., Fig. 11 T), which await reduction by corrasion and attrition. However, some of the huge and seemingly fragile clasts of lithified strata (e. g., Fig. 11 F and U) in the head of the southwest branches, where enhanced current speeds are anticipated, probably were carried short distances, perhaps by the highly turbulent head of a turbidity current rather than by its body. These isolated clasts, several meters in diameter, closely resemble the angular semilithified blocks abandoned by some subaerial floods, such as the 4-5 m diameter blocks of tuff-breccia scattered by an historic jokulhlaup (Thorarinsson, 1958) on an Icelandic sandur near the source of Maury Channel's turbidity currents.

Turbidity currents discharging at 1×10^6 m^3/sec. through the gap between the South Iceland and Maury Plains would be much slower and less competent than in the lower channel, because of the less confined flow. Although their beds' gradient steepens markedly at the heads of the upper channels, Froude numbers remain low, and there is no likelihood of a hydraulic jump. On the deep-tow traverse (Fig. 3) a flow with a mean depth of 20 m would be 25 km wide, and (by the Chezy equation) would have a mean velocity of only 2-3 m/sec. In the deepest part of the channel on this cross-section currents might be twice as thick and almost twice as fast. Quite unreasonable thicknesses, several hundred metres according to Komar's (1970) method of analysis, would be required to move the 1 m boulders photographed there and the much larger rocks seen on side-scan records.

In both surveyed reaches of Maury Channel the most important role of turbidity currents has probably been corrasive scour by the far-travelled sandy load, rather than local transport of large, locally derived clasts. The effectiveness of this process, which liberates additional sand grains from the Pleistocene turbidites, is attested by the abundance of such corrasion forms as flutes and potholes, and by shaping of channel-floor ridges into asymmetric blunt-nosed forms (Fig. 11 L) closely resembling the streamlined yardangs abraded by sand-laden winds in deserts (Mainguet, 1968). Despite this evidence for recent corrasion, many other rock surfaces, particularly in the southwest branch, have a profuse growth of fragile attached animals (e. g., Fig. 11 T), suggesting that in places scour recurs only at long intervals. Local variations in the rate of corrasion depend on the flux of suspended particles, and no doubt on bedrock lithogy. In the anastomosing reach of the lower channel the deepest downcutting has been in the eastern link, where a 15 m deep scour pit has been excavated (Fig. 7), and along the outside of channel bends. The overdeepening of the eastern link is clearly related to the lateral constriction and therefore acceleration of flow from the northwest to the southeast branch, but the reason for the narrowness of the connection is uncertain. Perhaps this link, and the parallel one to the west, follow narrow zones of lithologic weakness such as a north-south fracture system. The deepening of the channels on the outside of bends, which is most pronounced in the southwest branch (Fig. 8, profiles 5 and 6) is another result of the thickening and acceleration of the flow there by centrifugal force. This effect of the gentle right bends in the branches was more influential than Coriolis force, despite the high latitude, because the flows were so fast. As a result we do not observe the systematic cross-channel asymmetry that is found in most levee-building depositional channels (Menard, 1955), and is predicted for erosional channels by Baer's Law (Kabelac, 1957).

Direct fluid stress of the turbulent flow on poorly indurated or weakened beds has also caused erosion. Small-scale effects have included tearing clay balls from semi-lithified muds and plucking platy fragments from undercut ledges. Largerscale effects such as detachment of large blocks (e. g., Fig. 11 U) and scalloping of part of the channel wall (Fig. 12) are concentrated at sites of accelerated currents in the head of the southwest branch and down the "cataract" where tributary flow has cascaded into the southeast branch.

Mass-movements

Recent rockfalls are responsible for the piles of angular and joint-block talus that mark recession of the walls and reduction of some bedrock ridges in the lower channel (Fig. 11 R-T). Collapse of thin, undercut ledges of brittle rock is also common at both survey areas (e. g., Fig. 5 D).

Debris flows are active in some canyons and able to move large clasts down slopes with gentler gradients than the interplain channels. They probably deposited the bouldery beds of the upper channel (Fig. 5 G), though it is difficult to explain how such a flow could have been initiated after the sedimentary section had become indurated. Perhaps rapid recession of channel walls and heads by catastrophic turbidity-current erosion of the typical alternation of cemented sandstones and unlithified clays could create a mixture of resistant clasts and slurry that continues as a debris flow. Only massive entrainment of sediment by the density flow would allow reversal of the normal (?) transformation, from debris flow to turbidity current (Hampton, 1972).

Thermohaline and tidal currents

Maury Channel lies between the axes of two boundary currents of Iceland-Scotland Overflow Water: a broad southwesterly flow in the western part of the Iceland Basin, and a narrow northeasterly current along the foot of Hatton Drift (McCave *et al.*, 1980). This water type was identified below 3 200 m at our hydrocasts CTD 695 and CTD 697 into branches of the lower channel (Fig. 9), and current meters over the intervening residual island recorded a slow easterly flow which probably feeds the northeasterly boundary current. Within the channel branches at Meters CM 18 and CM 19 (Fig. 9) this flow was diverted northeast (up-channel) by the steep topography, and in the southeast branch it was accelerated to a mean velocity of 14.5 cm/sec. (maximum speed 20 cm/sec.). Measurement of fast, geologically competent thermohaline flow in the southeast branch explains the current scours and ripple marks found in unconsolidated sediment there (e. g., Fig. 11 V). The photographic evidence for recent downchannel currents in parts of the southwest branch may be caused by periodic expansion of the main southwesterly Overflow Water current across the flank of Gardar Drift. Suggestions that the channel formed or has been kept open as a direct "drain" for dense but non-turbid waters from the Iceland-Scotland Overflow (Cherkis *et al.*, 1973) are not supported by our data: the fastest measured currents and some of the clearest signs of sediment transport are directed up-channel (Fig. 9).

In both channel branches bottom-water motion over the rough topography around the CTD casts created a nearly homogenous bottom mixed layer about 100 m thick. Mixed layers are thinner at all nearby stations over Gardar and Hatton Drift (McCave *et al.*, 1980), and over gently sloping seafloor elsewhere they reach this great vertical extent only for the most energetic flows (Armi, Millard, 1976). The bottom mixed layer in the channel at CTD 697 has a fairly intense nepheloid layer, and gravimetric analysis of a filtered water sample revealed a suspended concentration of 43 µg/l. This sediment load is significant but much less than in the Overflow Water on the rise directly south of Iceland (> 100 µg/l; Richardson *et al.*, 1978). To explore the lateral variation in suspended sediment load, a nephelometer was attached to the deep tow instrument on profile 7 (Fig. 8) across the channel. There was a strong tendency for higher values over channel floors, especially within the current-swept southeast branch. The deep tow sediment pump collected an integrated sample of the suspended material along this profile (Mayer, Tyce, 1977): it was a mixture of clay flocks, skeletal planktonic debris, and organic-rich particles that may have been faecal pellets.

Bedload current ripples were photographed at two places in the channels, but the main load of the thermohaline currents must be in suspension, causing the measured nepheloid layer and the observed murkiness of bottom-water photographed near the upper channels (e. g., Fig. 5 A). This transport is not restricted to Maury Channel, but is part of the large-scale lateral flux of suspended sediment that supplies the basin's contourite drifts. Over the rough channel topography, increased turbulence in the thick bottom mixed layers probably enhances sediment entrainment; this may account for the abundance of rock exposures that might otherwise be veiled in mud, and for the local increases in nepheloid layer concentration. However, the role of thermohaline currents within our survey area seems to be local redistribution of sediment, rather than its wholesale removal, and there is no evidence that they erode lithified outcrops.

Several submersible studies of submarine canyons cut into continental slopes have found that the most impressive marks of contemporary sediment transport are fields of ripples and sand waves that may move alternately up and down canyons under the influence of oscillating currents (Keller, Shepard, 1978; Ryan *et al.*, 1978). At depths of more than a few hundred metres these oscillating currents are tidal surges (Shepard *et al.*, 1979), diverted and intensified by the topography just as the thermohaline currents in Maury Channel are. During our measurements at depths of 3 320-3 330 m the semi-diurnal tide (whose ellipse was oriented east-west at Meter CM 17 above the interchannel island) was also constrained to flow up and down the channel. However, its 5 cm/sec. current was so puny that if our records are representative of an entire monthly cycle, then the role of this abyssal tide is limited to episodically reinforcing the thermohaline current.

DEVELOPMENT OF THE CHANNEL SYSTEMS

As an unconfined turbidity current flowing down a depositional surface like the South Iceland or Maury Plains begins to erode its bed it will tend to become channelized because any local downslope depressions will have thicker, faster, and more erosive flow, which will further deepen them. By this positive feedback all the scouring flow through a broad cross-section may be collected in a few dominant channels, as in experimental studies of river incision by bedrock corrasion (Shepherd, Schumm, 1974). Subsequent development will be closely controlled by bedrock lithology, which affects the relative importance of channel deepening and channel widening. If, as in Maury Channel's bedrock, there is a rapid increase in resistance to erosion with depth in the section, then the channels' cross-section will vary with age.

The upper channels are in an early stage of incision of their interplain gap. Two almost straight, shallow channels, one with a tributary (Fig. 3), have cut back tens of kilometers from a "knick point" between the two plains. The young channel heads are broad, with degradation of their weak walls probably enhanced by generation of local debris flows there. The downstream narrowing of the older parts of the upper channels (Fig. 3) probably reflects incision into deeper, more competent beds that can form stronger walls.

Further development of channel cross-sections is seen in the lower channel survey, where only the broad shallow tributary and the head of the southwest branch are still at the nascent stage of stripping of weak superficial strata. The steep middle section of the southwest branch (Fig. 8, profiles 4-6) is in a "youthful" stage, where the feedback between rate of scour and channel depth is expressed by narrow inner channels between rock ridges, and an extreme channel asymmetry engendered by an initially slight cross-flow variation due to centrifugal force. The slightly sinuous plan that is characteristic of the erosional channels is probably produced at this stage of exaggerated asymmetry, where only one side is able to recede by undercutting and rockfall of a cliffed wall. Continued rapid downcutting is eventually impeded by increasingly resistant rock in the sedimentary section, by a decrease in the channel gradient, and by the finite size of flood flows: when the channel becomes deep enough to discharge entire turbidity currents, further deepening will not thicken and accelerate the chanelled flow. Lateral scarp recession, always important for reducing rocky ridges in the channel floor (e. g., Fig. 11 T), becomes the dominant erosion process, developing a broad channel with cliffs on both walls and a relatively smooth floor on a resistant stratum. Widening the channels is self-limiting, because as the cross-section enlarges then the mean velocity must decrease (assuming equal near-bankful discharges), and the flow is increasingly unable to sweep away large clasts that fall into the channel from its walls. The lower part of the southwest branch (profile 7, Fig. 8) is almost at this "mature" stage, which is best displayed by the southeast branch: a 1.5 km-wide channel, with almost constant cross-section and cliffed walls even on the inside of a bend, which is just able to contain all the flood water supplied to it while discharging fast enough to remove all but the largest talus (e. g., Fig. 11 S). This branch is quite comparable to the anomalously large Snake River canyon that was shaped by a just-brimful catastrophic flood (Malde, 1968).

Continuing the analogy to the Davisian fluvial cycle, a branch may enter "old age" if scouring turbidity currents are diverted from it during evolution of an anastomosing channel. This pattern cannot be caused by branching and rejoining around depositional bars as in most braided rivers (e. g., Krigstrom, 1962), for the interchannel islands are residual bedrock. Our genetic model (Fig. 13) involves straightening and shortening of the flow path during peak discharge events. When a flood larger than the original sinuous channel can contain flows down the interplain gap, part of the turbidity current spills out of the channel on the downstream sides of bends, and reenters on the upstream side of the next bend (Fig. 13, stage 1). The shallow, low-angle tributaries cut by this reentering flow, like the one mapped at 56°22'N (Fig. 9), extend headward up the plain as straight branches similar to the upper channels (stage 2). Rapid deepening of the former tributaries occurs after their heads intersect the sinuous channel and allow them to tap more of the scouring flow, even at times of lower discharge. As a result their downstream junctions where the channel is rejoined are converted from hanging to concordant, and their upstream ends experience violent turbulent flows over the sill out of the main channel (stage 3). We propose that the southwest branch is at this stage. As the steeper, youthful branch is further enlarged, its shorter route may intercept an increasing share of the flow, and the older channel may be partially abandoned, eventually becoming a backwater for turbidite deposition, just as in cut-off reaches of meandering channels (Lonsdale, Hollister, 1979). The southeast branch in our

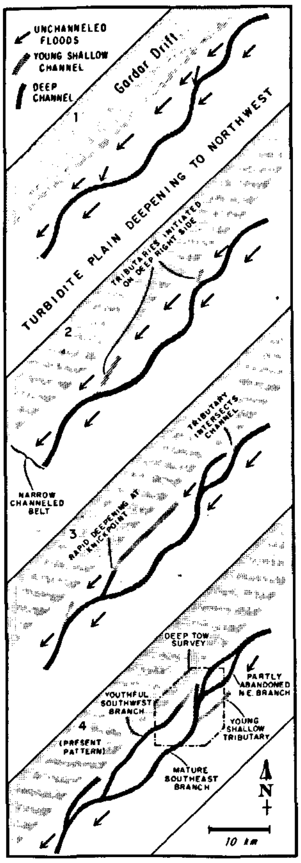

Figure 13
A model for the development of anastomosis in erosional turbidity current channels, applied specifically to the 56°30′N-56°10′N reach of Maury Channel mapped in Figure 6 *(inset).*

survey area has probably experienced diminishing discharge as the southwest branch has enlarged, but the best example of an "old age" channel may be the northeast branch, which has degraded, uncliffed walls (Fig. 9), and an unsampled sediment fill up to 5 m thick.

Development of anastomosis between 56°30′N and 56°00′N (Fig. 6, inset) may be helped by the 1 : 200 cross-flow tilt of the interplain gap between Maury and Eriador Plains. This tilt restricts thick flows that are capable of channel incision and headward recession to the narrow (6 km) belt at the foot of Gardar Drift, increasing the frequency with which extending channels will intersect; it may also account for the better development of right-bank tributaries (Fig. 13).

SPECULATIONS ON THE EROSIONAL HISTORY OF THE INTERPLAIN CHANNELS

The upper and lower channels truncate turbidites of probable Late Pleistocene age whose lithologies and flat-lying attitudes show no evidence of major channels at the time of their deposition, though some cross-bedding preserves a record of minor cut-and-fill (e. g., Fig. 11 F). We deduce that channel incision has been quite recent, in Late Pleistocene and/or post-glacial times. It is possible for excavation of interplain channels at formerly depositional sites to be initiated by processes internal to the deep-sea sediment dispersal system: by rejuvenation of turbidity currents as they overtop a former dam (Laughton, 1960), or by gradually accelerating them as their routes are constricted by the growth of sediment ridges (in the Iceland Basin). However, it seems more plausible that the change from turbidite deposition to scour in Maury Channel was triggered by external, climatic factors, since the ultimate source of its turbidity currents is phreatomagmatic and subglacial eruptions in Iceland, which experienced drastic climatic fluctuations during the likely period of channel incision. Laughton *et al.* (1972) suggested that all of the thick turbidite section on Maury Plain is of Pleistocene age, and that the dominance of hyaloclastite fragments suggests that it was mostly deposited during glacial periods, when Icelandic volcanism produced prolific hyaloclastite beneath its almost complete ice cover. The change in regime at the end of the last glacial period, when Icelandic turbidity currents may have increased in intensity as they decreased in frequency, could have initiated deep-sea erosion. A change in the type of sediment yielded by the island, with an increased proportion of sandy phreatomagmatic debris, may have been responsible, as might a change in the amount of fine debris pirated by the Iceland-Scotland Overflow Current. Whatever the trigger, the interplain reaches of the currents' route were favoured for erosion because they were steeper.

The best direct evidence that turbidity currents continue to flow down the interplain channels and scour them, albeit at long recurrence intervals, is the fresh appearance of many of the rock surfaces, especially in the upper channels (e. g., Fig. 5 E). Indirect evidence that turbidity currents still enter the system, though they may not penetrate to its distal reaches, comes from observation of very recent catastrophic erosion in Reynisdjup Channel, one of the principal rise-valley feeders (Lonsdale, Hollister, 1979). Recent events that are likely to have created such turbidity currents are the phreatomagmatic eruptions of Surtsey, just 8 years before our deep tow surveys.

If the interplain channels formed only in brief post-glacial times, they must have been excavated by a few major floods. The complex evolution of channel morphologies and patterns may have been compressed into a few hours of catastrophic erosion. Compared to river channels there is a very rapid spatial transition from "youth" to "maturity" and "old age" (e. g., downstream in the southwest branch, Fig. 6), and there may have been correspondingly brief temporal transitions. Most of the sediment excavated during incision of the channel systems, amounting to only 20-30 km^3, was probably dispersed over the downstream plains, together with a much greater volume of first-cycle volcanogenic sediment. However in the lower channel the sharp local decrease of channel gradient at the transition from the anastomosing erosional upper reach to the meandering depositional lower reach (Fig. 2, 1 160 km downchannel) may mark a small flood deposit.

SIGNIFICANCE OF OUR OBSERVATIONS

(1) Our high-resolution surveys have described deep-sea erosional terrains with straight, sinuous and anastomosing bedrock channels; our discussion of their probable formation and evolution emphasizes the role of catastrophic erosion by episodic turbidity currents. The

observations invite comparison with large-scale flood channels eroded in other environments, which have aroused widespread interest since the discovery of spectacular "outflow channels" on Mars (e. g., Baker, 1978; Komar, 1979). Many years of controversy followed "catastrophists" claims that the subaerial geomorphology of large areas like the Columbia Plateau had been transformed by a few hours of intense flooding (Baker, 1978); the concept of similar modification of the floor of a deep ocean basin is quite novel.

(2) We discovered and incompletely described excellent deep-sea exposures of distal volcanogenic turbidites, whose structures may be compared to those of flysch and graywacke sequences believed to be of deep-water origin but now exposed on land. For more fruitful comparisons, precision sampling of the channel walls would be desireable. Our photographs also show modern deposits similar to those of submarine canyons, far out on the abyssal plains. Observation of huge bedload clasts, debris flow deposits and rockfall talus emphasize that local high-energy environments occur on the basin floors, and that some ancient distal sequences may contain more than the thin-bedded muddy turbidites assumed by most facies models (e. g., Walker, Mutti, 1973).

(3) Direct and photographic evidence shows that in the long intervals between density flow events, thermohaline currents move sediment along the channels as suspended load and, rarely, as bedload. The currents are topographically intensified and diverted to flow along the channels, but they flow both up and down, and may leave a record of scour marks oriented opposite to the turbidity current scours. Tidal surges, a major force in many submarine canyons, were too weak to have much effect on sedimentation in these deep-sea channels.

(4) Although near-bottom surveys encompass only short reaches of Maury Channel, we have tried to integrate them with the best-available surface-ship data to present a revised overview of the entire system. It is a complex of channeled and unchanneled routes for sediment dispersal, in which deposition and erosion alternate both spatially and temporally. The interplain channels have had a brief geologic history, and are vulnerable to further rapid evolution.

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