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Initiation of turbidity currents: outcrop evidence for Eocene hyperpycnal flow turbidites

Piret Plink-Björklund[a,*], Ron J. Steel[b]

[a] *Department of Geology, Earth Sciences Centre, Göteborg University, Box 460 Guldhedsgatan 5A, SE-405 30 Goteborg, Sweden*
[b] *Department of Geology and Geophysics, University of Wyoming, Laramie, WYO 82072-3006, USA*

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Abstract

The most widely used, although not as widely documented, explanation for initiation of turbidites is the conversion of catastrophic sediment failures into turbidites. Such instantaneous sediment failures generate short-lived surge-type flows that deposit turbidite beds described as Bouma sequences. In this communication, we present data from the Eocene Central Basin of Spitsbergen that shows turbidite systems that are volumetrically dominated by thick ungraded and laminated sandy turbidite beds, deposited by gradual aggradation from sustained flows. Sustained flows are long-lived and more or less continuous. They can be generated by a variety of mechanisms, e.g.: (1) instability during volcanic eruptions and the consequent remobilisation of unconsolidated material; (2) seismically triggered subaerial sliding within the drainage; (3) storm surges; (4) retrogressive slope failure; (5) breaching; and (6) hyperpycnal flows. Our database shows that in the Eocene Central Basin, the sustained flows were generated by hyperpycnal flows, i.e., by direct river effluent.

The hyperpycnal flow turbidites have been recognized on the basis of (1) physical connection between fluvial and turbidite channels at the shelf edge, (2) abundance of thick turbidite sandstone beds, (3) sand-prone nature of the turbidite systems, (4) downslope changes of individual thick, sandy turbidite beds, and their collapsed pinch-out segments, (5) great abundance of continental material (leaves, coal fragments) in turbidite beds, (6) low abundance of associated slumped or debris-flow beds, and (7) the occurrence of turbidites in systematically accreted shelf-margins. These features strongly suggest that hyperpycnal flow, generated by direct river effluent, deposited significant number of turbidite beds in the documented slope and basin-floor turbidite successions. We also argue that the hyperpycnal flows are preferentially fed into the deepwater slopes beyond the shelf edge during the falling stage and lowstand of relative sea level.

The turbidite beds were studied in the context of seismic-scale (1 × 15 km), dip-oriented outcrops by measuring detailed vertical sections, following individual beds and packages of beds downslope for distances over 5 km, lateral mapping, and helicopter-taken photomosaics.

Keywords: Hyperpycnal flow; Sustained turbidity currents; Surge-type turbidity currents; Shelf-margin accretion

1. Introduction

The most widely accepted explanation of turbidite initiation relies on conversion of catastrophic sediment failures into turbidites (Heezen and Ewing,

* Corresponding author.
E-mail address: piret@geo.gu.se (P. Plink-Björklund).

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1952; Menard, 1964; Morgenstern, 1967; see review by Normark and Piper, 1991). However, Maar (1999) has shown that the conversion of failure-generated debris flows and landslides to dilute turbidites is an inefficient process. Very large debris flows are required to produce relatively small turbidites (see Parsons et al., 2001). Such instantaneous sediment failures, like seismogenic slumping (e.g., Heezen and Ewing, 1952; Chamberlain, 1964; Morgenstern, 1967; Normark and Piper, 1991; Weaver et al., 1992; García and Hull, 1994), or "oversteepening" and remobilization of sediment deposited at the shelf break or in the canyon head (e.g., Kolla and Perlmutter, 1993; Helland-Hansen and Gjelberg, 1994; Mulder and Syvitski, 1996), generate short-lived (minutes to hours) *surge-type turbidity currents*. Surge-type currents have their experimental analogue in finite-volume currents generated by the release of a lock gate, and may consist of more-or-less clearly differentiated head, body, and tail, with a general decrease in both mean velocity and sediment concentration backwards from the head (e.g., Kneller and Buckee, 2000; Felix, 2002). Typically, the velocity of the current passing a fixed point will very rapidly reach its maximum value as the head arrives, after which flow will wane during passage of the body and tail (Kneller and Buckee, 2000).

Documentation of (1) regular bedforms (e.g., Maliverno et al., 1988; Normark et al., 1980; Normark and Piper, 1991), (2) meandering turbidity current channels (e.g., Kastens and Shor, 1985; Manley and Flood, 1988), (3) delta-front chutes (Kostaschuk and McCann, 1987; Bornhold and Prior, 1990; Phillips and Smith, 1992; Carlson et al., 1992), and (4) recognition of a large variety of turbidite beds that do not fit the surge-type (waning) flow model (see Bouma, 1962; also Lowe, 1982; Pickering et al., 1995; Branney et al., 1990; Kneller et al., 1991; Middleton, 1993; Edwards et al., 1994; Kneller and Branney, 1995; Kneller, 1995; Hiscott et al., 1997; Kneller and Buckee, 2000; Plink-Björklund et al., 2001) provide evidence for occurrence of longer-lived, i.e., *sustained turbidity currents* (Branney and Kokelaar, 1992; Kneller and Branney, 1995; Kneller, 1995). Such turbidity currents are steady or quasi-steady flows that last for days or weeks (e.g., Kneller and Branney, 1995). Evidence for sustained flows has been collected from some few ancient turbidite successions (e.g., Kneller, 1995; Kneller and Branney, 1995; Plink-Björklund et al., 2001), and numerous analytical and numerical models of turbidity currents have produced solutions for steady flow (see Kneller and Buckee, 2000 and references therein).

Several studies have suggested *hyperpycnal flows* (underflows generated by direct river effluent) as an explanation for such observed deposits (Piper and Savoye, 1993; Mulder and Syvitski, 1995; Nemec, 1995; Piper et al., 1999; Mulder and Alexander, 2001; Plink-Björklund et al., 2001; Mellere et al., 2002). Mulder and Alexander (2001) even argue that all the sustained flows are hyperpycnal flows, whereas most workers suggest that there is a variety of other mechanisms that can produce sustained flows, e.g.: (1) instability during volcanic eruptions and the consequent remobilisation of unconsolidated material (Lipman and Mullineaux, 1981; Kokelaar, 1992); (2) seismically triggered subaerial sliding within the drainage basin (Syvitski and Schafer, 1996); (3) storm surges (e.g., Inman et al., 1976; Shepard et al., 1977; Dengler et al., 1984); (4) retrogressive slope failure (e.g., Andresen and Bjerrum, 1967); and (5) breaching (gradual retrogression of very steep slopes) in fine-grained sands (Van den Berg et al., 2002).

Hyperpycnal flows generated by direct fluvial discharge clearly do exist, as documented from lakes (Lambert et al., 1976; Strum and Matter, 1978; Weirich, 1986; Lambert and Giovanoli, 1988), fjord delta fronts (Prior et al., 1986; Prior and Bornhold, 1990; Nemec, 1990; Zeng et al., 1991; Carlson et al., 1992; Phillips and Smith, 1992), and open marine shelf–margin systems (Heezen et al., 1964; Shepard and Emery, 1973; Reynolds, 1987; Wright et al., 1988, 1990; Piper et al., 1999). The common problem with acceptance of hyperpycnal plume generation from riverine outflow is, perhaps, the traditional criterion for hyperpycnal flow initiation, i.e., that the *bulk density* of a river effluent, has to exceed that of the basinal ambient in order to generate an underflow (Mulder and Syvitski, 1995). As river outflows are typically fresh and warm relative to the ocean, such a criterion requires bulk sediment concentrations approaching 40 kg m^{-3}. According to this criterion, only a few small rivers in mountainous terrain are capable of producing hyperpycnal plumes once every 100 years or so (Mulder and Syvitski, 1995). This constant has been shown now to be a rather conservative estimate, as the actual density required for a river effluent to generate a hyperpycnal flow can be considerably lower, if the hyperpycnal plume forms a few days after the beginning of the flood (when the salinity at the river mouth has been lowered) or because of general high freshwater discharge (Mulder et al., 1998). Moreover, sediment-charged rivers show considerable density stratification, and many flood events produce both hypopycnal and hyperpycnal flows (McLeod et al., 1999; Kneller and Buckee, 2000).

Multiple river-effluent-generated turbid underflows with calculated concentrations of only ca. 1 kg m^{-3} have been reported from the Salinas River into and down the Monterey Canyon, based on observations over a period of 12 years (Johnson et al., 2001). The underflows were detected by lower salinity, warmer water and high particle content in the plume, in water depths up to 1020 m. The authors suggest that turbid underflows, capable of transporting much of the suspended river load down the continental slope, occur with a recurrence rate of 2.5–3 years. It is possible that small underflows are even more frequent.

Parsons et al. (2001) have provided a mechanism for generation of very dilute hyperpycnal flows. With this suggested mechanism, river outflow concentrations as low as 1 kg m^{-3} are sufficient, even if the salinity is normal in the receiving basin. They show experimentally, that the warm, fresh, particle-ladden hypopycnal (surface) plume, when it flows over a relatively dense, cool brine, is subject to a convective instability and the consequent generation of a hyperpycnal (bottom riding) plume. The convective instability is produced by the downward transport of heat and sediment from the upper layer within the hypopycnal plume. Lowering the initial required density of ca. 40 kg m^{-3} (Mulder and Syvitski, 1995) to only ca. 1–5 kg m^{-3} (Johnson et al., 2001; Parsons et al., 2001) suggests that many modern rivers can produce hyperpycnal flows annually (Parsons et al., 2001; see also Mulder and Syvitski, 1995).

Our work on clinoforms of the Eocene shelf margin of Spitsbergen (Fig. 1) suggests that significant volumes of deepwater sands were deposited from sustained (steady or quasi-steady) turbidity currents initiated by direct river effluents and accumulated on both slope and basin-floor segments of the shelf margin (Steel et al., 2000; Plink-Björklund et al., 2001; Mellere et al., 2002). The Spitsbergen Central Basin presents a unique potential for collecting a data set to evaluate the initiation mechanisms of turbidity currents. The dip-parallel clinoform exposures (1 × 30 km scale, Fig. 1) allow a "walk-out" of the transition from the shelf edge to the deepwater slope and basin floor, and therefore to monitor sediment transport, and sediment volume partitioning from river mouth into slope channels and canyons, and onto the basin floor. Populations of turbidites accumulated in different sub-environments on the clinoforms, i.e., upper slope, lower slope, basin floor, and at different times within the fall-to-rise cycle of relative sea level. We also know exactly when the rivers were located at the shelf edge (and therefore when sediment-laden river floods had direct access to the deepwater slope) versus times when the deltas were still far back on the shelf. We also have a good knowledge about the type, size, and salinity of the basin, as well as the regime of the deltas and feeding river systems, and the source area. We can visually observe the character of the shelf edge (e.g., presence or absence of canyons, amount of wave action, etc.), and "walk out" the connection from the river/distributary channels down onto the slope turbidite channels, and further to the basin floor (for distances of 5–20 km). This knowledge, combined with the high quality and detail of the outcrops, allows us to eliminate some of the mechanisms that could potentially generate sustained turbidity currents, and focus on the most probable ones.

To the best of our knowledge, this paper is the first systematic outcrop documentation of hyperpycnal flow beds from ancient turbidite systems. We argue that direct river input by hyperpycnal flows is responsible for generation of most of the sustained flow turbidites in the Eocene Central Basin of Spitsbergen.

2. Geological setting

The Eocene Central Basin was a small (100 × 200 km) foreland (Steel et al., 1985) or piggy-back (Blythe and Kleinspehn, 1998) basin, bounded to the west by the active West Spitsbergen fold-and-thrust belt (Fig. 1). The clinoforms, and their equivalent 'rock units' referred to as clinothems (Rich, 1951), reflect the outgrowth of the basin margin from the fold-and-

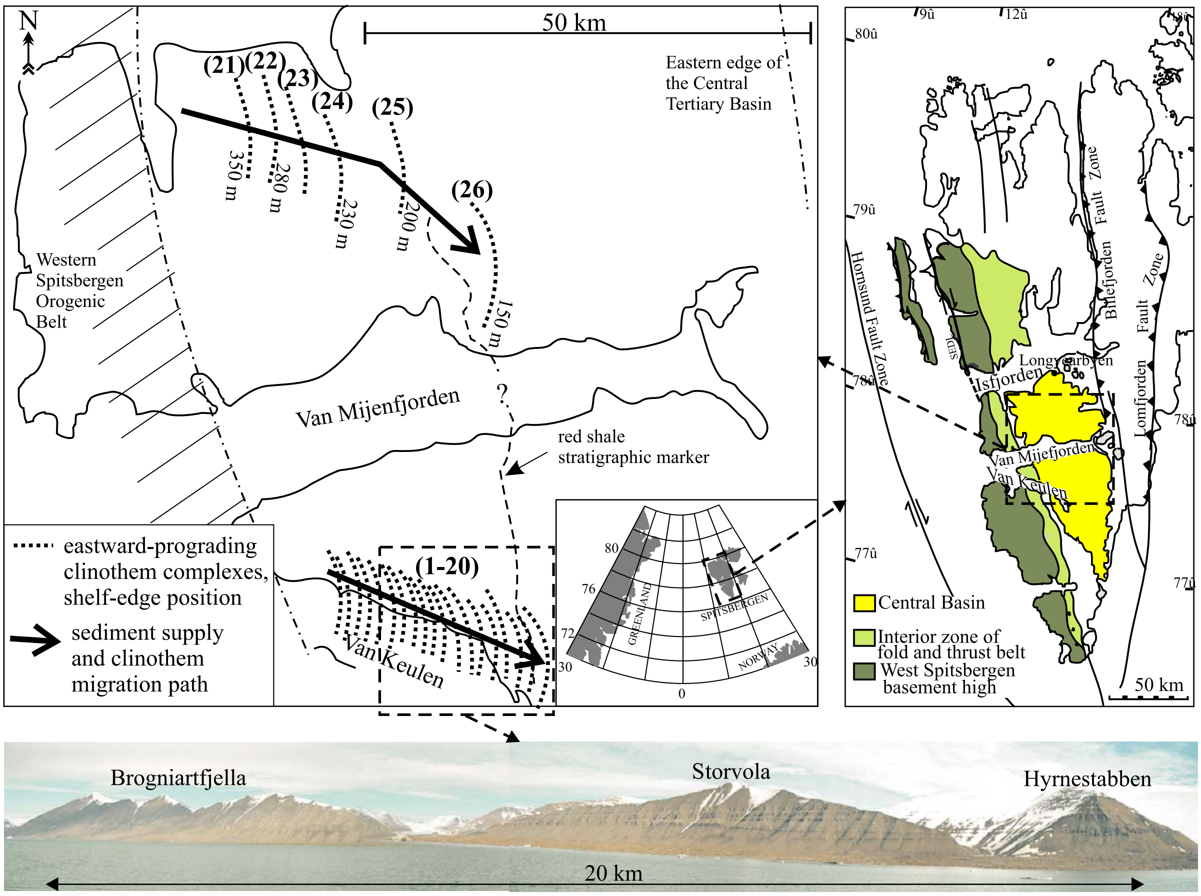

Fig. 1. Eocene Central Basin was bounded to the west by an active West Spitsbergen fold-and-thrust belt. The clinoforms reflect the outgrowth of the basin margin from the fold-and-thrust belt towards the east/southeast. The database consists of two broadly time equivalent transects: the Van Keulenfjorden transect (clinothem complexes 1–20) and the Van Mijenfjorden transect (clinothem complexes 21–26). Each of the transects contains eastward-younging coastal plain–shelf–slope–basin-floor clinothem complexes.

thrust belt. Clinoforms reflect successive time lines as the basin infilled by asymmetric accretion (Fig. 2). The surface of any clinoform represents the morphologic profile from the coastal plain to the marine shelf and down into the deeper water slope and basin-floor environments, so we are using the term 'clinoform' in a broader sense (i.e., for the entire length of the time line) than first defined by Rich (1951). The characteristics of some of these clinoforms have been discussed by Steel et al. (2000), Plink-Björklund et al. (2001), Plink-Björklund and Steel (2002), Mellere et al. (2002), and Steel and Olsen (2002).

A 30-km-long, dip-oriented transect exposes a series of coastal plain shelf–slope–basin-floor clinoforms, with de-compacted amplitudes of 200–400 m (Fig. 1). The exposed and accessible length of individual clinoform complexes is up to 15 km, and our preliminary estimates of the time duration of these complexes is 200,000–400,000 years (i.e., produced during 4th-order relative sea level falls and rises).

Mapping over distances of 40 km out from the western edge (foredeep) of the basin demonstrates that the shelf to basin-floor clinothems occur in a shingled stratigraphic pattern, with successively younger clinothems being offset basinwards through time (Fig. 2). Most of the basin-infilling clinothems in the Central Basin of Spitsbergen are shale-prone throughout (Type 4), or are sand-prone only as far out as their mid-shelf reaches, and shale-prone from outer shelf to slope and basin-floor reaches (Type 3, Fig. 3). However, there are

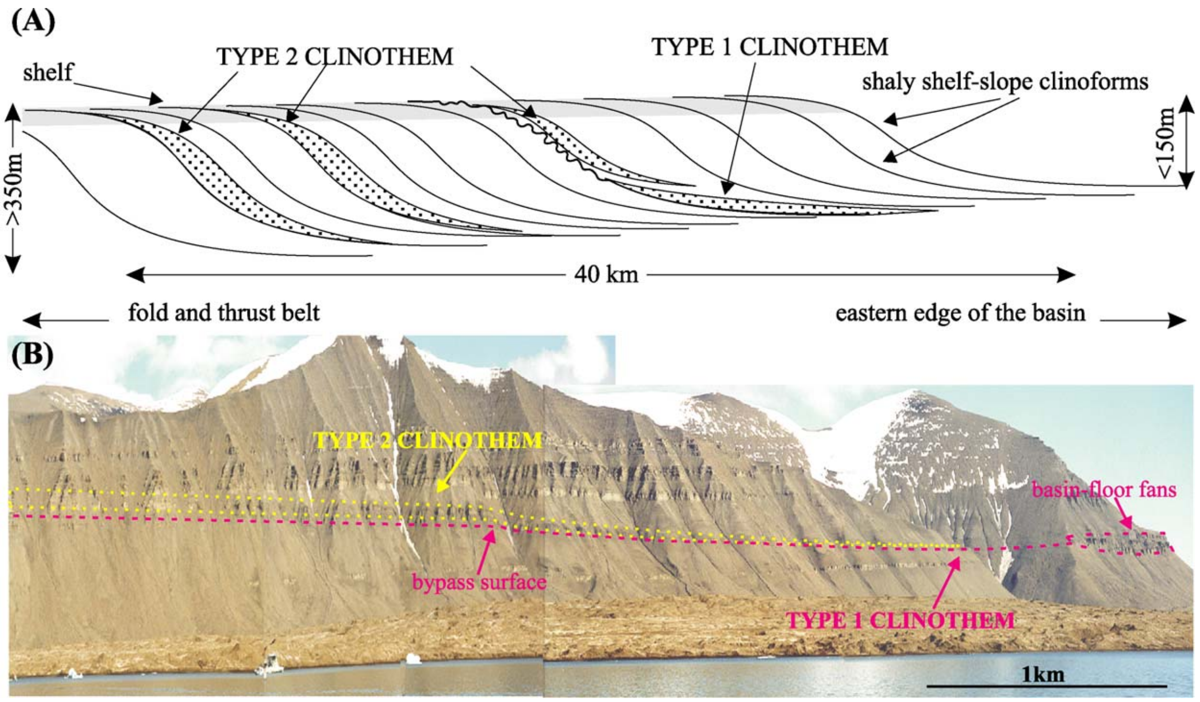

Fig. 2. (A) The shelf to basin-floor clinothems occur in a shingled stratigraphic pattern, with successively younger clinothems being offset basinwards through time. Type 1 clinothems show mainly bypass from the shelf edge to lower slope but have thick sand-prone turbidite accumulations from the base of slope out onto the basin floor. Type 2 clinothems develop a sandy slope turbidite wedge but have no significant basin-floor extension. (B) Examples of Type 1 and Type 2 clinothems on Storvola (redrawn from Plink-Björklund et al., 2001).

occasional clinothems that are sand-prone across the entire shelf, bringing sands onto the shelf edge and slope (Type 2), and even rarer clinothems that provide evidence of sand delivery out beyond the slope, onto the basin floor (Type 1, Fig. 3). The topsets of clinoforms are lithostratigraphically assigned to the Aspelintoppen Formation, and foresets and bottomsets to the Battfjellet Formation. The clinoforms overlie sharply the Gilsonryggen Member shales, and are themselves overlain by marine shales in younger downlapping shaly clinothems of the Battfjellet Formation. This communication focuses mainly on the turbidite-producing, Types 1 and 2 clinothems.

3. Evidence for hyperpycnal flows in Spitsbergen outcrops

The linkage between the onshore to offshore facies tracts and the nature of the deepwater depositional systems were studied by lateral mapping, measuring of tightly spaced detailed vertical sections, and walk-out of individual beds down the palaeoslope, along the seismic-scale, dip-oriented mountainside exposures. The excellent quality of the outcrop belts, which expose eastward-younging clinoform complexes, enabled "walk-out" of internal architecture, as well as individual beds or packages of beds where the mountainsides were not too steep, otherwise detailed lateral relationships were determined from helicopter photomosaics.

The following observations strongly suggest that hyperpycnal flows, generated directly by river or distributary effluent, deposited most of the Spitsbergen turbidites: (1) physical connection between fluvial channels on the shelf edge and turbidite channels on the upper slope; (2) general abundance of thick turbidite sandstone beds; (3) generally sand-prone character of the turbidite systems; (4) downslope changes of individual turbidite beds and identification of their collapsed pinch-outs; (5) great abundance of relatively delicate, terrestrial material (leaves, coal) in

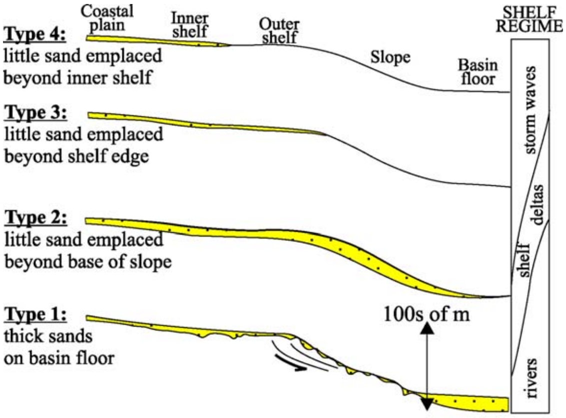

Fig. 3. Most of the basin-infilling clinothems in the Eocene Central Basin of Spitsbergen are shale-prone throughout (Type 4), or are sand-prone only as far out as their inner-shelf reaches, and shale-prone from outer-shelf to slope and basin-floor reaches (Type 3). There are occasional clinothems that are sand-prone across the entire shelf, bringing sands onto the slope (Type 2), and even rarer clinothems that give evidence of sand transport far out beyond the slope, onto the basin floor (Type 1) (redrawn from Mellere et al., 2002).

turbidite beds; (6) low abundance of associated slumped or debris-flow beds; and (7) the occurrence of turbidites in systematically accreted shelf-margins. Each of these observations is discussed separately, although it is collectively that they most strongly indicate hyperpycnal flow.

3.1. Physical connection between fluvial and turbidite channels

There are two distinctly different settings where the fluvial system at the shelf edge has been correlated into turbidite channels on the slope. One is where incised river channels at the shelf edge feed down into basin-floor fans (Type 1 clinoform) via small canyons on the slope; the other is where the distributary channels of shelf-edge deltas feed into small slope channels and chutes, without developing basin-floor fans (Type 2 clinoforms).

3.1.1. Rivers feed into canyons

Type 1 clinoforms have turbidite systems that were fed by rivers that debouched into channels and small canyons cut into shelf edge and slope (Steel et al., 2000). Lateral to and within the canyons are bypass surfaces that lead down into basin-floor fans. The transition from river deposits to turbidites at the shelf edge has been documented in several clinoform complexes along the Van Keulenfjorden transect (Fig. 4). The dip extension of individual clinoform complexes is up to 15 km, and the thickest sands occur in the basin-floor fans (up to 65 m thick).

The fluvial facies is dominated by medium- to coarse-grained, massive, flat-stratified, planar/trough cross-stratified sandstones. Individual fluvial units commonly show an upward-fining trend or motif, but can also be complex in their vertical organization. Basal erosion surfaces of units are usually overlain by intervals of flat-stratified or trough/planar cross-stratified pebbly sandstone, with some interstratified sets of clast-supported conglomerate, covered by stacked, cross-stratified sandstone sets. These sand-rich, fluvial units are interpreted as fluvial channel belts (Steel et al., 2000; see also Nemec et al., 1988; Nemec, 1992; Gjelberg and Steel, 1995) on the basis of the cross-stratified facies, the upward-fining character of units, the lower erosional contacts of individual units, the dominantly unidirectional, basinward-directed nature of the palaeocurrents, and the lack of marine fossils. At some localities, it appears that the channel belts are tidally influenced, as judged by sigmoidally shaped cross-stratified sets, and bi-polar cross-stratal directions. Wave-influenced, upwards-coarsening mouth-bar complexes are recognized below and at the mouth of the fluvial channels. The fluvial channel belts cut down into older shelf deltas or upper-slope mudstones.

Beyond the shelf edge, the fluvial channels connect into slope-channel or small canyon systems (Fig. 4). The canyons can be followed from the shelf edge downslope for several kilometers. The canyon walls cut abruptly into older slope shales and prodelta deposits. The canyon fills are up to 25 m thick in upper-slope settings and consist largely of a complex of turbidite-filled channel systems that stack progressively basinwards and downslope through time, or show a lateral accretion. Each channel unit has a marked basal erosion surface (Fig. 4), paved by up to 20-cm-thick lag of medium-grained sandstone, with shale rip-up clasts and coal fragments. The overlying sandy deposits consist of 50–150-cm-thick beds with

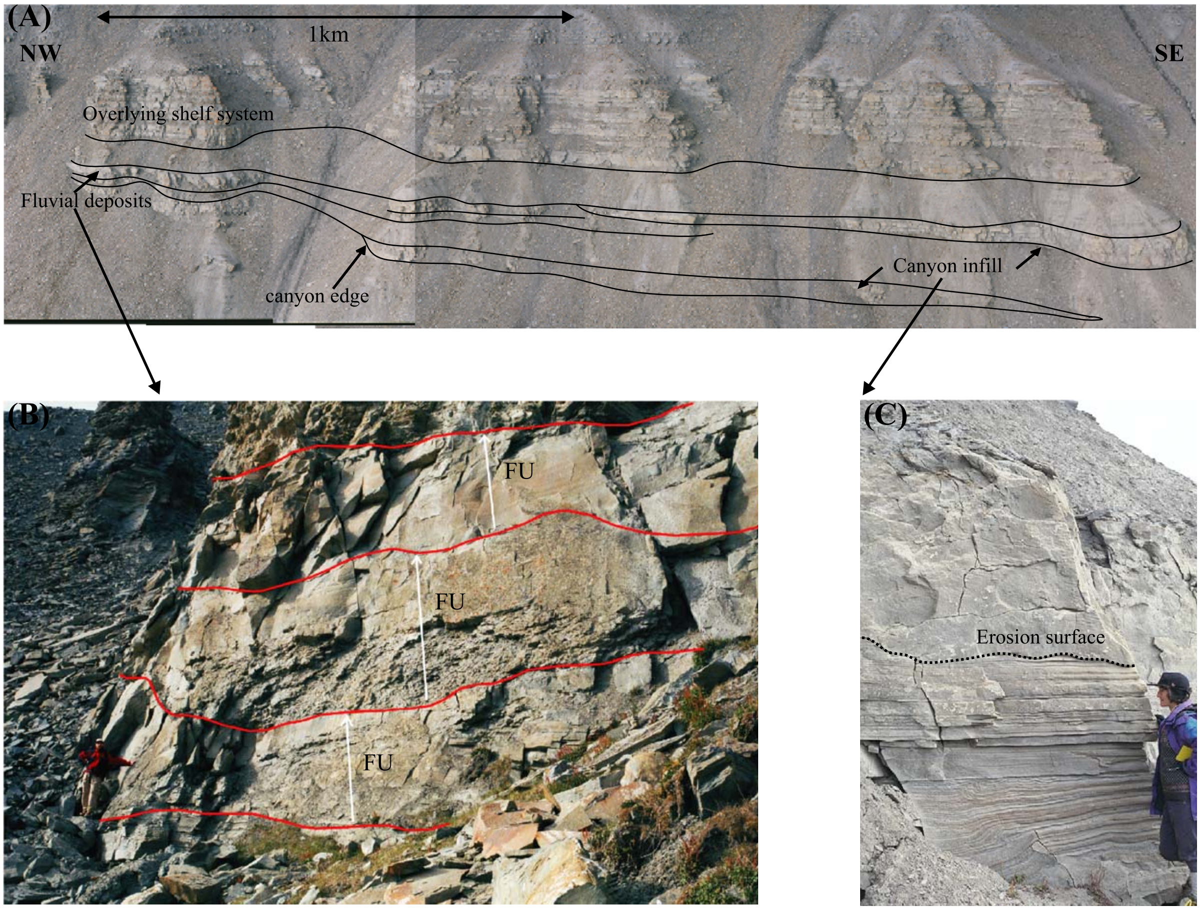

Fig. 4. (A) Transition from river deposits to turbidite deposits in a small upper-slope canyon on Brogniartfjella (Clinoform complex 9 in Fig. 1). The transition can be walked-out on the steep mountainsides (D. Mellere, personal communication). (B) Detail of (A) shows fluvial deposits that in this locality consist of upwards-fining erosionally based packages. (C) Detail of (A) shows thick-bedded plane-parallel laminated and ungraded sandy turbidite deposits.

a characteristic plane-parallel lamination (Fig. 4). In places, the sandstones overlying the basal lags grade upwards from massive sandy beds (< 2 m thick) to somewhat finer-grained intervals of alternating massive/plane-parallel and current-ripple laminated beds. In places, the basal unit of the infill consists of chaotic slumps, forming a wedge-like body (up to 1 m thick, 25 m long) that pinches out both updip and downdip on the canyon floor.

3.1.2. Delta distributary channels feed into slope channels

The turbidite systems of Type 2 clinoforms are wedge-shaped bodies that are up to 80 m thick at the shelf edge and pinch-out on the lower slope or just beyond the base of slope (Plink-Björklund et al., 2001; Plink-Björklund and Steel, 2002). The normal dip extension of the sandstone wedges is within a distance of ca. 3500–5500 m (Fig. 5A). The slope clinoform complexes consist of a series of sand-prone, partially amalgamated turbidite lobes, sheets and channels that can be traced from shelf edge to basin floor (Steel et al., 2000; Plink-Björklund et al., 2001; Plink-Björklund and Steel, 2002; Mellere et al., 2002). Individual clinothems within the complex are 5–10 m thick, channelled and sand-prone. Muddy or heterolithic intervals (< 0.5 m thick) that thicken basinwards separate the individual clinoforms. The stacking pat-

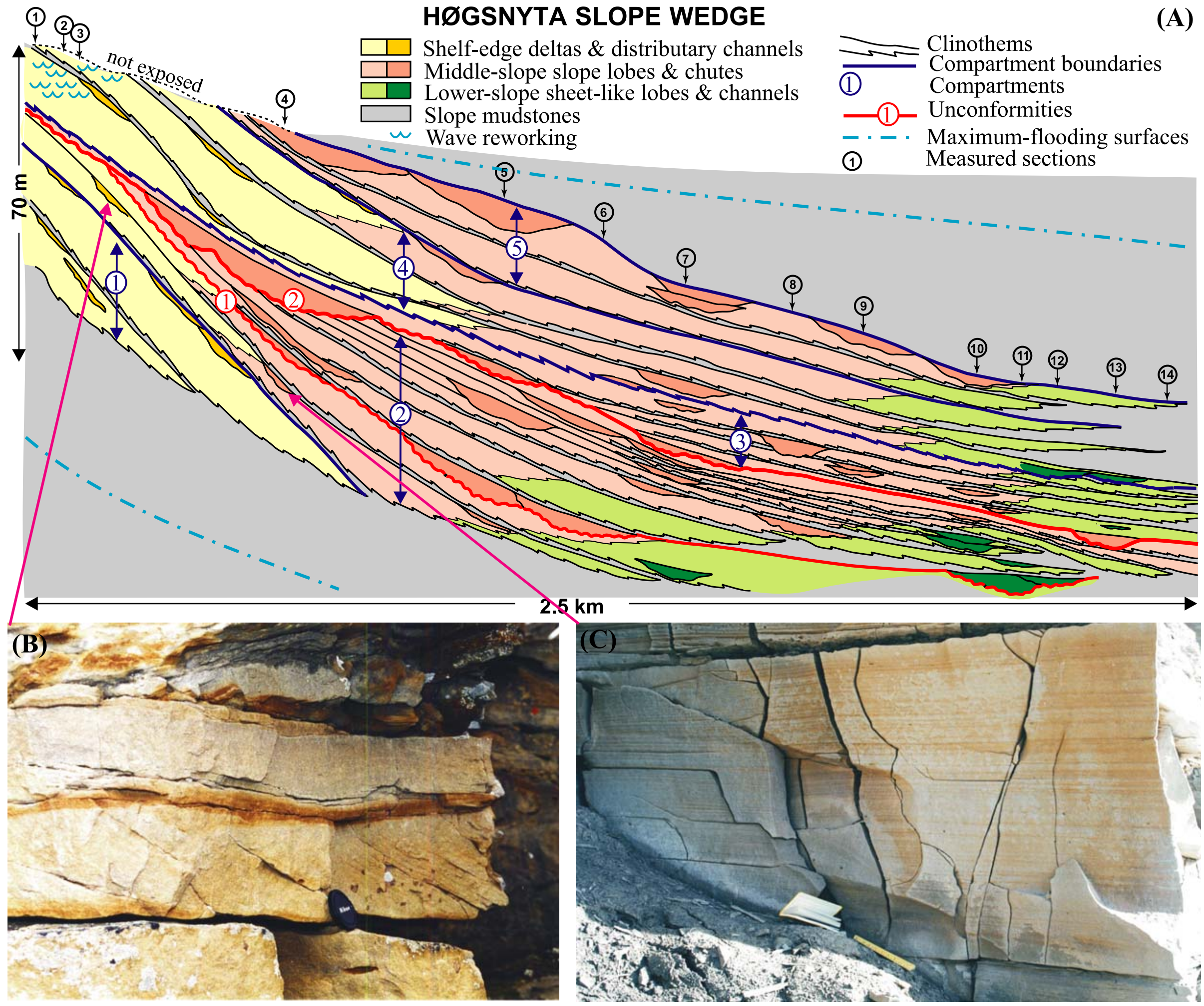

Fig. 5. (A) Høgsnyta slope wedge (Clinoform complex 26 in Fig. 1) consists of a series of clinoforms. Fluvial distributary channels (B) can be followed into turbidite beds in slope channels (C) within individual clinoforms. The Høgsnyta diagram is redrawn from Plink-Björklund and Steel (2002).

tern of clinothems in the complex varies slightly, but typically consist of (1) a lower seaward-stepping segment that downlaps onto the slope mudstones, (2) a seaward-shifting segment with multiple erosion surfaces (unconformities), (3) and an aggradational segment (Fig. 5A).

The seaward-stepping and -shifting segments (1 and 2) of clinothems consist of the steep-fronts of shelf-edge deltas that prograded onto the upper slope, just below the shelf edge (Fig. 5). The tops of the deltas are severely eroded. The sandy prograding delta front is steep (up to 20°) and consists of 2–10-m-thick, medium- to coarse-grained sandy sheets, made of ungraded and laminated turbidites, separated by thin (1–10 cm) very fine sandstone to siltstone beds. The delta topset consists of distributary channels (25–50 m wide, 1–5 m deep) and minor mouth bars (a few meters thick). The distributary channels contain unidirectional trough cross-stratified, medium to coarse-grained sandstones (5–25-cm-thick sets) at the base, typically overlain by sets of plane-parallel laminated sandstone (Fig. 5B). Mouth bars consist of upwards-coarsening units of alternating traction deposits (unidirectional trough-cross stratified or plane-parallel stratified sandstones) and grain-flow deposits (inversely or inversely-to-normally graded pebbly coarse-grained sandstones 0.9–2 m thick), and are typically cut by distributary channels. The mouth bars pass laterally and downslope into steeply inclined foresets. In the aggradational segment of the clinoform complexes (see item (3) above), the deltas are wave reworked.

The delta distributary channels can be followed downslope into small turbidite-filled channels (chutes), cut into the delta-front and upper-slope deposits (Fig. 5C). Some of the chutes terminate in minor sandy or heterolithic slope lobes within the upper-middle-slope reaches, though some chutes continue onto the lower-slope areas (Fig. 5A). The chutes are 50–100 m wide and 1–3 m deep, filled with fine- to medium-grained sandstones. The bases of channel fills typically expose multiple erosion surfaces aligned with coal or clay clasts (up to 5 cm in diameter). The channel fills typically contain a combination of ungraded or normally graded, parallel-laminated or ripple-laminated, and crudely upward-thinning and -fining sandstones. Some channel fills have up to 25-cm-thick siltstone or mudstone capping. Most of the chute-mouth lobes are sand-prone, and the heterolithic deposits occur mainly as a fine-grained fringe to the lobes.

3.1.3. Interpretation

The physical connection from river channels into canyons, and from delta distributary channels into slope channels (chutes), indicates that at times, the fluvial feeder system was dumping sediment directly onto the slope (see also Steel et al., 2000; Plink-Björklund et al., 2001; Plink-Björklund and Steel, 2002; Mellere et al., 2002). Closely spaced chutes from prodelta slopes and hyperpycnal flow events through such chutes have been well documented (Prior et al., 1987; Kostaschuk and McCann, 1987; Bornhold and Prior, 1990; Phillips and Smith, 1992; Carlson et al., 1992). However, some workers have ascribed chutes to small slides, or local liquefaction generated through a combination of wave-induced cyclic loading and oversteepening following progradation (e.g., Prior et al., 1981; Syvitski et al., 1987). The chutes in the Spitsbergen turbidite systems extend up to the mouth of distributary channels just as the canyons link up to river mouth; there is no evidence for large-scale sediment instability in the slope turbidite systems, and both the canyon-and chute-fills indicate deposition from sustained flows (see below, also Plink-Björklund et al., 2001). The Spitsbergen data set also shows that where there is a significant wave regime on the outer shelf, there are few sandy turbidites on the slope (see Steel et al., 2000). We conclude that both river and distributary effluent created hyperpycnal flows that deposited their load in deepwater slope and basin-floor (see below) areas below the shelf edge.

3.2. Abundance of thick sandstone beds

Laminated, ungraded, inversely and/or normally graded sandstone beds (0.5–4 m thick) are abundant in Spitsbergen turbidite systems. Even the ungraded and graded thick sandstone beds typically display intervals with traction structures, like plane-parallel lamination, ripples, rounded aligned clay clasts, and internal scours. The grain size changes from medium-coarse at the shelf edge to fine close to the base of slope, but many of the beds pinch-out sandy, i.e., they do not grade into muds. The following types of thick sandstone beds have been observed in Spitsbergen turbidite systems.

(1) Laminated sandstone beds. The beds consist of 0.5–2.5-cm-thick sandstones with a characteristic sub-horizontal lamination (Fig. 6A,B,C). The lamina in the beds are 0.5–2 cm thick, are plane-parallel to slightly wedging, and record subtle grain-size variations from lower fine- to upper fine-grained sand, or from upper-fine to medium-grained sand. In several places, the beds contain internal diffuse and discontinuous scour surfaces. The beds show a crudely developed grain-size profile that is overall upwards fining. The thick laminated beds have a marked basal erosion surface, in some places paved by up to 20-cm-thick lag of coarser-grained sandstone, with shale rip-up clasts and coal fragments. The thick laminated beds occur in upper-middle-slope channels (chutes) and in proximal canyons.
(2) Laminated and ungraded sandstone beds. These beds contain fine- to medium-grained sandstone and are up to 4.5 m thick, with sharp and flat or scoured bases (Fig. 6B,D,E). Thinner beds (0.5–2

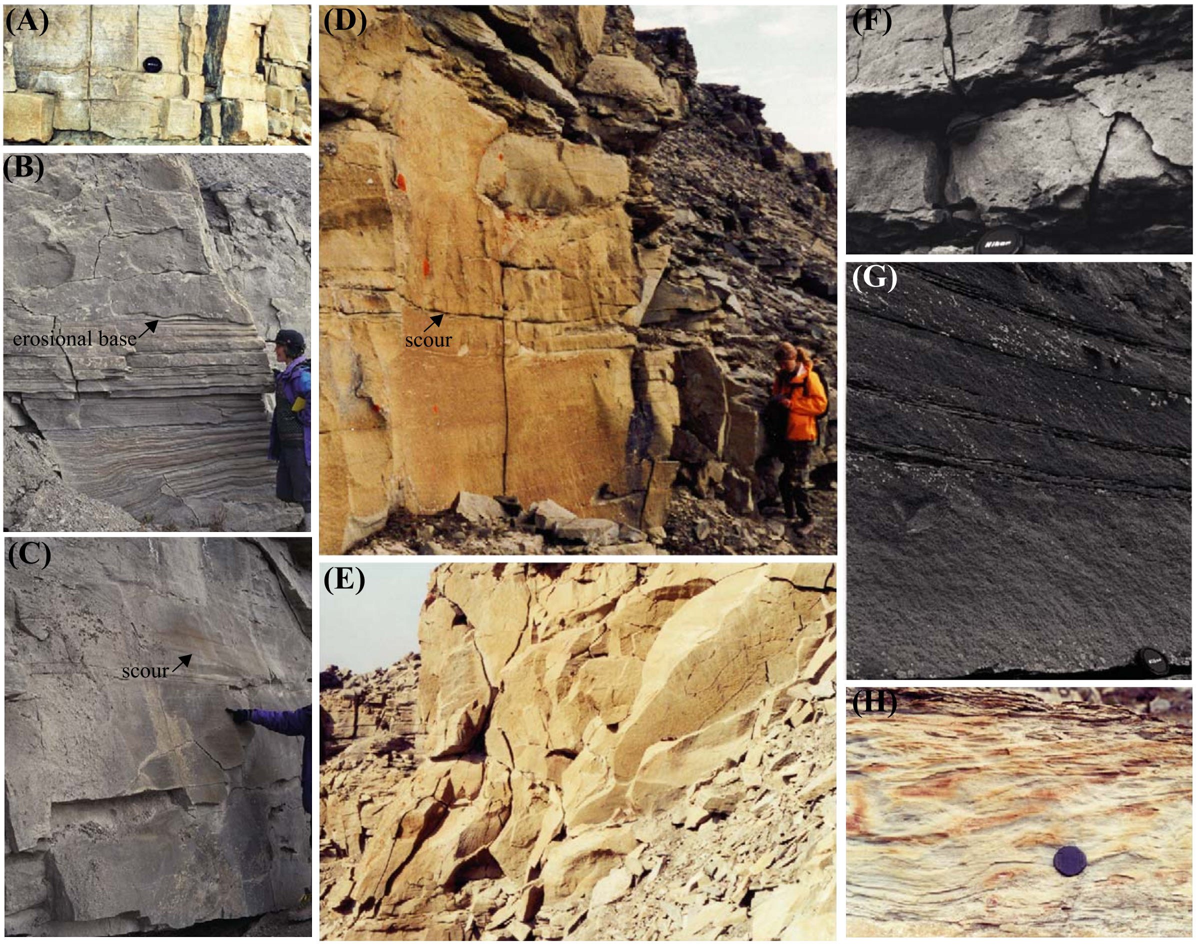

Fig. 6. Sand-prone turbidite beds (0.5–4 m thick) are abundant in Spitsbergen turbidite systems. (A) Characteristic plane-parallel lamination in the thick turbidite beds. Slope channel fill on Høgsnyta (Clinoform complex 26 in Fig. 1). (B) Plane-parallel laminated bed is erosionally overlain with an ungraded bed. Canyon fill on Brogniartfjella (Clinoform complex 8 in Fig. 1). (C) Internal scours in a plane-parallel laminated bed. Canyon fill on Brogniartfjella (Clinoform complex 9 in Fig. 1). (D) Alternating laminated and ungraded intervals, and internal scours in an almost 3-m-thick bed. Haagfjellet slope wedge (Clinoform complex 21 in Fig. 1). (E) Ungraded to normally graded 2-m-thick beds in basin-floor fans on Hyrnestabben (Clinoform complex 14 in Fig. 1). (F) Clay chips and coal fragments tend to orient parallel to the bedding surfaces. Slope turbidites on Litledalsfjellet (Clinoform complex 26 in Fig. 1). (G) A succession of ungraded beds with coarser rippled tops. Slope wedge on Haagfjellet (Clinoform complex 21 in Fig. 1). (H) Climbing ripples on Haagfjellet (Clinoform complex 21 in Fig. 1).

m thick) commonly have a sheet-like geometry, whereas the thicker beds (4–4.5 m thick) fill slope channels (chutes) 50–100 m wide and 1.5–4 m deep. The beds (average 2–2.5 m thick) consist of alternating ungraded and plane-parallel or current-ripple laminated sandstones with no vertical trend of grain size or sedimentary structures. Even where ungraded, the sandstones contain internal diffuse and discontinuous scour surfaces. In many places, aligned and rounded shale clasts occur. Clast orientation is random, or bedding-parallel and/or imbricated. The shale clasts, at bed bases or within the sandstone beds, are sometimes aligned above internal scour surfaces (Fig. 6F). These beds occur in shelf-edge delta-front sheets, canyons, in upper- to lower-slope channels, and on basin-floor fans.

(3) Ungraded sandstone beds with coarser tops. Individual beds consist of fine-to medium-grained sandstone with ungraded lower parts

and slightly coarser plane-parallel, current-ripple laminated or inversely graded tops (Fig. 6G). In places, plane-parallel or current-ripple lamination dominates throughout the beds. Climbing current ripples occur in places at the bases of slightly coarser ungraded intervals (Fig. 6H). The laminated tops may disappear laterally, scoured by an overlying sandstone bed. In most cases the erosional contacts are subtle. The individual beds, 0.2–1 m thick, have sharp, flat, or scoured bases, and sheet-like geometry. Dish structures and ball and pillow structures occur in places, most commonly at the distal pinch-out of the beds. These beds occur on upper to lower slope.

(4) Normally/inversely graded beds. These beds are inversely to normally graded, inversely graded, normally graded, or show several intervals of inverse to normal grading within individual beds. The beds, 0.5–2 m thick, grade from medium to upper-fine, or upper-fine to lower-fine grained sandstone. The beds contain traction structures, like thin plane-parallel or ripple-laminated intervals, or internal scours, in places aligned by coal fragments or clay chips. The graded sandstone beds have a sheet-like geometry, or are confined in slope channels. These beds occur in middle- to lower-slope reaches.

3.2.1. Interpretation

The abundance of traction structures, erosional bases of beds with aligned clasts, and sheet-like or channel geometry reflect deposition by turbidity flows (Hiscott et al., 1997). The thick beds with traction structures suggest deposition by gradual aggradation from sustained (quasi-steady) turbidity currents (Kneller, 1995; see also Branney and Kokelaar, 1992; Kneller and Branney, 1995), which (judged by the thickness) were sustained for days or weeks. The vertical changes in the beds reflect temporal changes in flow properties, rather than the vertical structure of the current as a whole (Kneller and Branney, 1995). Similarly, the deposit thickness is unrelated to the thickness of the current as a whole (Kneller and Branney, 1995), because deposition occurs by gradual aggradation below the moving current (Branney and Kokelaar, 1992). The alternation of stuctureless and laminated intervals, and the internal scour surfaces and subtle grain-size changes reflect temporal variations in flow velocity and sediment flux within the same current, as indicated by the discontinuity of the scour surfaces and the constant grain size above and below the surfaces. Alternation of traction-deposited (laminated) intervals with ungraded intervals indicates alternation of low and high fallout rates. The ungraded intervals reflect rapid fallout rates, whereas the laminated intervals occurred when the vertical grain flux intermittently decreased and enabled a distinct interface between the current and the substrate (Arnott and Hand, 1989; Kneller and Branney, 1995), allowing traction and the development of plane beds and ripples. Similarly, the inversely graded intervals, and coarsening from the ungraded to the laminated intervals, and the occurrence of climbing ripples at the base of ungraded beds suggest a drop in fallout rates, probably due to a temporal increase in shear velocities rather than a drop in concentration (Kneller and Branney, 1995; see also Hiscott and Middleton, 1979). This suggests that such intervals were deposited by temporally accelerating flows. The beds that display multiple inversely to normally graded intervals, then indicate multiple episodes of acceleration and deceleration, i.e., fluctuation of the shear velocity.

The roundness of aligned clast reflects bedload transport, and clast orientation suggests traction deposition (Middleton, 1993; Johansson and Stow, 1995; Kneller and Branney, 1995; Hiscott et al., 1997; cf. Shanmugam et al., 1995), suggesting that the clasts aligned otherwise unrecognizable depositional boundaries that migrated upwards during sedimentation. Alternative interpretations, such as the settling of clasts through a flow or gliding at the top of high-density inertia-flow layers (Postma et al., 1988), or the laminar flow of debris (Stauffer, 1967; Lowe, 1982; 1988; Shanmugam et al., 1995) are unlikely because of clast fabric, roundness, traction structures and internal erosion surfaces (Arnott and Hand, 1989; Kneller and Branney, 1995; Hiscott et al., 1997). The portions of beds that show water escape structures indicate rapid fallout of grains that disabled traction and created an ungraded bed. The capture of fluids within the beds suggests rapid deposition by loss of flow capacity.

Many workers agree that one of the features indicative of hyperpycnal flow is the great thickness of individual turbidite beds (e.g., Piper and Savoye, 1993; Mulder et al., 1998; Kneller and Buckee, 2000; Mulder and Alexander, 2001), because a river effluent is semi-continuous, i.e., sustained for at least hours to days, or even weeks at a semi-constant discharge (e.g., Wright et al., 1986; Hay, 1987; Prior et al., 1987; Wright et al., 1988; Nemec, 1990; Wright et al., 1990; Chikita, 1990; Zeng et al., 1991; Mulder et al., 1998; Piper et al., 1999). There are, however, alternative mechanisms for generation of sustained flows, such as (1) volcanic eruptions with their consequent remobilisation of unconsolidated material (Lipman and Mullineaux, 1981; Kokelaar, 1992), (2) seismically triggered subaerial sliding within the drainage basin (Syvitski and Schafer, 1996), (3) storm surges (e.g., Inman et al., 1976), (4) retrogressive slope failure (e.g., Andresen and Bjerrum, 1967), (5) breaching (gradual retrogression of very steep slopes) in fine clean sands (Van den Berg et al., 2002), and (6) transformation of debris flows into turbidites. We argue that the thick sandstone beds were deposited by hyperpycnal flows, and that alternative mechanisms are unlikely to have initiated sustained flows in Eocene Central Basin of Spitsbergen, because (1) there was no volcanic activity in the basin or hinterland, (2) a seismogenically triggered remobilization within the drainage basin is a catastrophic event, whereas the studied turbidite systems (especially the slope wedges) reflect a very continuous and systematic shelf-margin accretion, (3) where we observe extensive wave-reworking on the shelf, there are few sandy turbidites on the slope or basin floor; (4) retrogressive slumping would most probably generate more "pulsative" sustained flows (reflected in turbidite beds as multiple upwards fining intervals). Moreover, in the Spitsbergen slope turbidite systems, slope failures are extremely rare and their deposits volumetrically insignificant, (5) breaching occurs in well-sorted fine sands, whereas the documented turbidite beds consist of medium-grained sandstone to mudstone beds, and (6) debris-flow deposits are very rare in the Eocene Central Basin, they are observed only on delta fronts and at the base of canyons, where they form volumetrically minor sedimentary bodies.

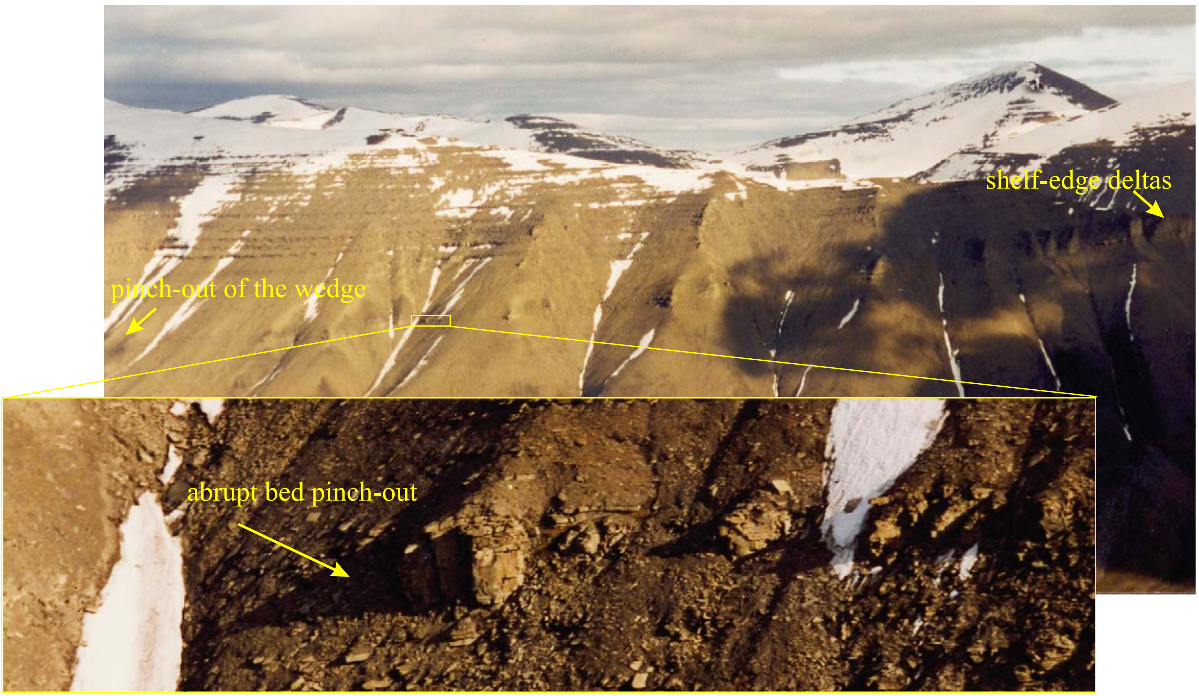

Fig. 7. Following the thick sandstone beds downslope reveals that they maintain their thickness for long distances, and pinch-out abruptly. Litledalsfjellet (Clinoform complex 26 in Fig. 1).

3.3. Abrupt pinch-out of thick sandstone beds

Characteristically, many of the thick sandstone beds described above fine and thin basinwards, but maintain their thickness for long distances, e.g., they thin from 3 to 1.5–2 m across 2–3 km and then they pinch-out abruptly in a few hundreds of meters (Figs. 7 and 8A). The pinch-out segments of such beds display structureless sandstones with occasional dish and ball structures, deformed by slightly younger soft-sediment folds (Fig. 8A,F). Walk out of the thick sandstone beds downslope from upper slope to lower slope shows that the proportion of lamination in the bed decreases distally, i.e., the fully plane-parallel laminated beds change into beds showing alternating laminated and structureless intervals, and then into structureless or graded beds with only occasional plane-parallel laminated and rippled intervals (Fig. 8B,C,D,E,F). In contrast, the thin surge-type flow (or waning flow) beds thin gradually downslope,

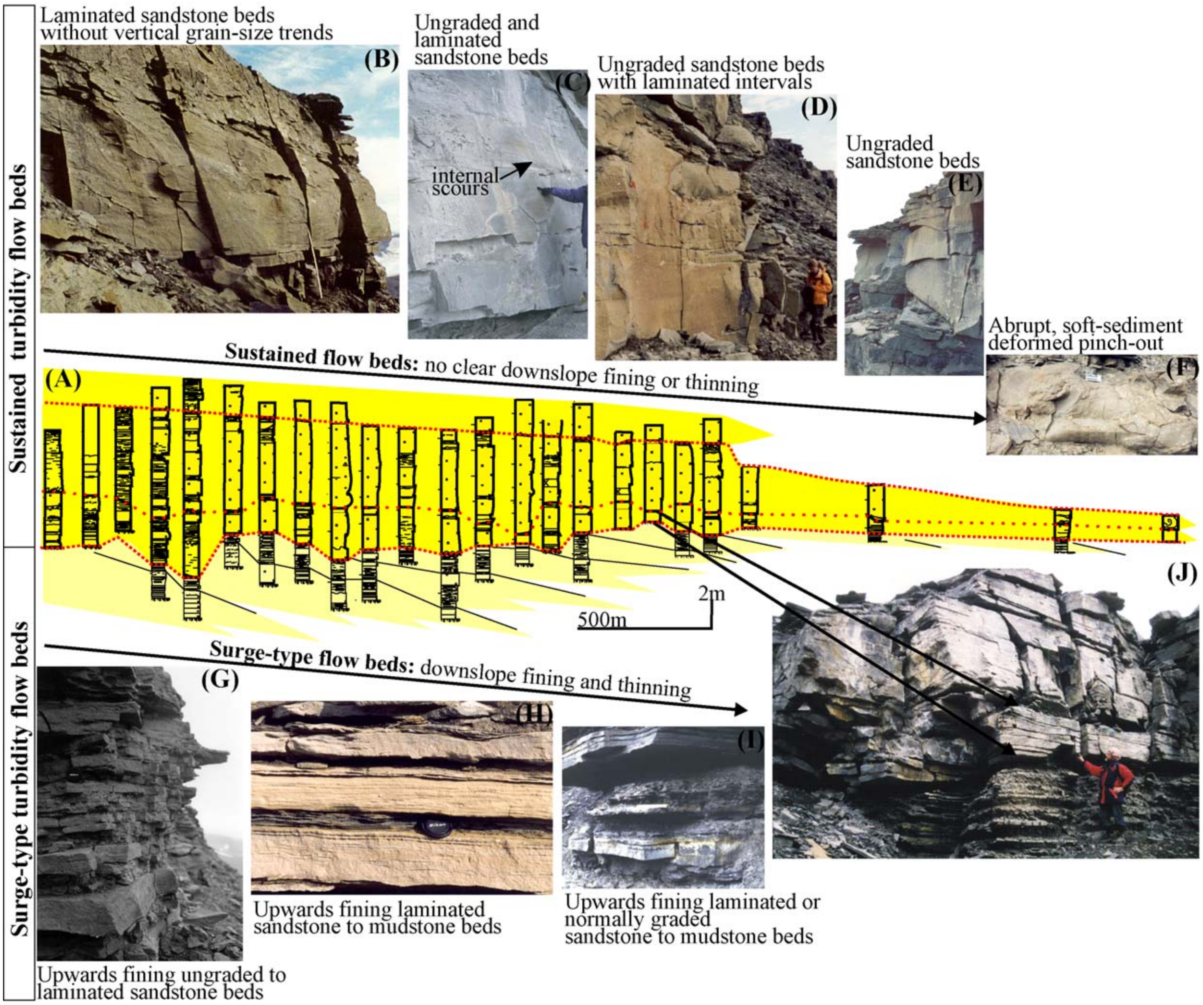

Fig. 8. (A) Finsenfjellet slope wedge (Clinoform complex 23 in Fig. 1) consists of thin surge-type turbidity flow beds in the lower portion, and of thick sustained turbidity flow beds in the upper portion (J). The sustained flow beds characteristically maintain their thickness for several kilometers, and pinch-out abruptly (B–F). In contrast, the thin surge-type flow beds thin and fine downslope (G–I). The thick sustained flow beds tend to be more laminated in their upslope reaches (B, C), and they become more structureless or graded in their downslope reaches (D, E). The pinch-outs are typically sand-prone and soft-sediment deformed (F).

and fine distally from sand-prone to heterolithic and then mud-prone beds (Fig. 7G,H,I).

3.3.1. Interpretation

Deposition from density flows results in a reduction in bulk density, with a subsequent decrease in the flow velocity, when deposition is not countered by erosion and sediment entrainment. In hyperpycnal flows, the interstitial fluid is less dense than the ambient fluid, although the bulk density of the current is greater, i.e., the flows are driven by the excess density of the suspended particles. After a sufficient volume of particles has been deposited, the current will become buoyant, cease its lateral motion, and ascend to form a buoyant plume (Sparks et al., 1993; Kneller and Buckee, 2000; Mulder and Alexander, 2001). The experiments (Sparks et al., 1993) show that there is little difference between a sediment-laden current with neutrally buoyant interstitial fluid and one with buoyant interstitial fluid until sufficient sediment has been lost to cause the latter kind of current lift-off. A marked deceleration is then observed and a plume is generated, with lift-off occurring along the length of the current. The deposit from a current with buoyant fluid shows a fairly abrupt decrease in thickness beyond the lift-off distance and has a flatter profile than that of a simple sediment current (Sparks et al., 1993). The internal segregation of dense particles to the base of the flow would eventually lead to the generation of a convective plume rising off the top of the flow, leading to decoupling of the fine and coarse particles (McLeod et al., 1999). The fine sediment lofted from the fluid may become much more widely dispersed than the coarser population. Sparks et al. (1993) suggest that the spreading of the buoyant cloud may result in an enhancement in grain size contrast between the "e" and lower divisions of Bouma sequence (Bouma, 1962), and deposition of hemiturbidites (Stow and Wetzel, 1990). The coarse fraction after the collapse of the current may produce debris-flow-like or collapsed deposits with abrupt fronts (Lipman and Mullineaux, 1981; Twichell et al., 1995; Kneller and Buckee, 2000). The proximal to distal changes from laminated to ungraded within the beds (Fig. 7) suggest an increase in sediment fall-out rates from proximal to distal, which then culminated with collapse of the distal end of the bed.

The buoyancy effects would not be significant if the interstitial fresh water is changed out by entrainment of the seawater. However, the experiments of Hallworth et al. (1996) suggest that entrainment of the ambient fluid into the head of a gravity current fed by a constant flux is much slower than the entrainment into the head of a surge-type flow. This is due to the continuous replenishment of the fluid in the head by the constant feed of undiluted fluid from the tail. As the abrupt pinch-out segments on Spitsbergen occur only in thick sandstone beds, the depositing flows were sustained and likely to have fresh interstitial water.

3.4. Sand-prone successions

The Eocene Central Basin turbidite systems are sand-prone (Fig. 9), as the thick sandy sustained flow beds dominate volumetrically many of the turbidite systems. Such sandstone beds are only occasionally capped by thin heterolithic or mudstone layers (Fig. 8B,C,D,E,F). Even the surge-type flow beds are rather sandy in their proximal reaches (Figs. 8G and 9D). Muddy or heterolithic intervals occur as a fine-grained fringe laterally or distally to the sandy turbidites, or as thin (0.1–0.5 m) intervals between sandy clinoforms (5–10 m thick) (Fig. 9A,B).

3.4.1. Interpretation

The sandiness of the succession suggests that either the turbidites were generated sand-prone (e.g., the mud was selectively removed), or sand was entrained into the current from the bed. Any turbidity current can serve as a triggering mechanism for larger sandy flows (García and Parker, 1993). However, that (1) the turbidites deposited in proximal reaches of canyons, and in delta-front and upper-slope channels are sand-prone, and (2) the shelf-edge delta-front sheets are sand-prone, suggests that the turbidity flows were sand-rich from the start. In the slope clinoform complexes of the Eocene Central Basin, there are no significant collapse features at the shelf edge. Any documented slope instability features can be traced into minor slump deposits that pinch-out within a few to tens of meters from the initiation point. Moreover, if the flows were generated by slumps, the transformation into turbidites would probably occur first on the lower-slope or basin-floor reaches, as such a

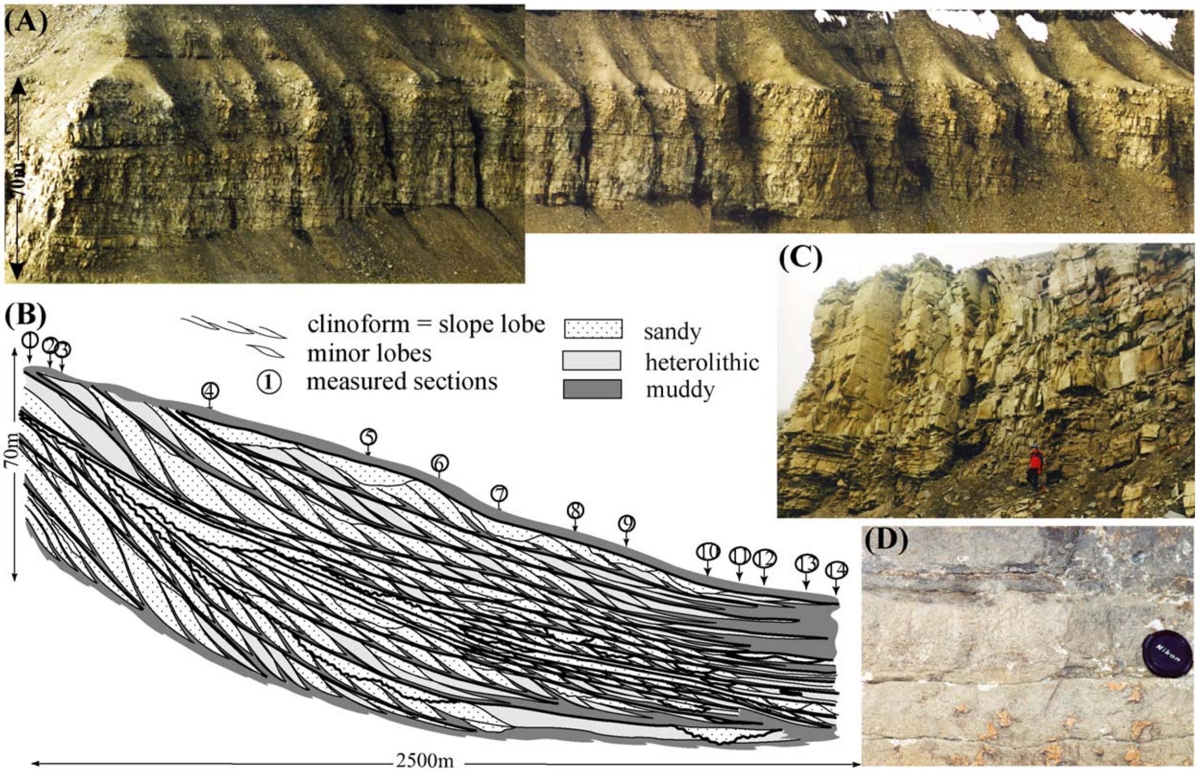

Fig. 9. Both the Type 2 slope turbidite wedges (A, B; Clinoform complex 26 in Fig. 1), as well as the Type 1 basin-floor fans (C; Clinoform complex 14 in Fig. 1) that contain thick sustained flow beds tend to be sand-prone. The muds or heterolithic intervals occur between the individual clinoforms or in the distal fringe rather than in individual beds (B, D). The diagram is redrawn from Plink-Björklund et al. (2001).

transformation from slumps to debris flows and turbidites is an inefficient process (Maar, 1999).

Also, the lack of large channels or canyons in the slope systems confirms that the flows were sand-rich when they were initiated, because sandy (i.e., non-cohesive) turbidity currents must have the ability to erode and entrain sediment from the vicinity of the bed, because turbulence alone is not sufficient to keep non-cohesive materials in suspension (Parker, 1982). The fact that the turbidity currents in slope systems on Spitsbergen were only rarely able to erode large channels, and never generated canyons, implies that the flows rarely entrained new sediment. Neither did they reach an ‘‘ignited’’ stable state, consequently, they travelled slowly and were not able to increase their density significantly.

When a river enters the ocean, it can generate a hypopycnal plume, which remains at the surface, or a hyperpycnal plume, which descends to the sea floor as a result of excess density generated by its sediment load, depending on the access density of the effluent (Parsons et al., 2001). Parsons et al. (2001) also describe that, often, both a hyperpycnal and a hypopycnal flow are generated when the fluvial discharge enters the basin. This will cause a decoupling of coarse and fine populations (Parsons et al., 2001). McLeod et al. (1999) have suggested that both, hypopycnal and hyperpycnal flows, and the decoupling of coarse and fine populations is most likely to occur when particle concentration in the fluvial effluent is relatively low, and the flow density close to the seawater density.

3.5. Abundance of leaves and plant/coal fragments

Coal fragments and fossil leaves are abundant in the Spitsbergen turbidite systems (Fig. 10). Coal occurs as randomly distributed centimeter-size frag-

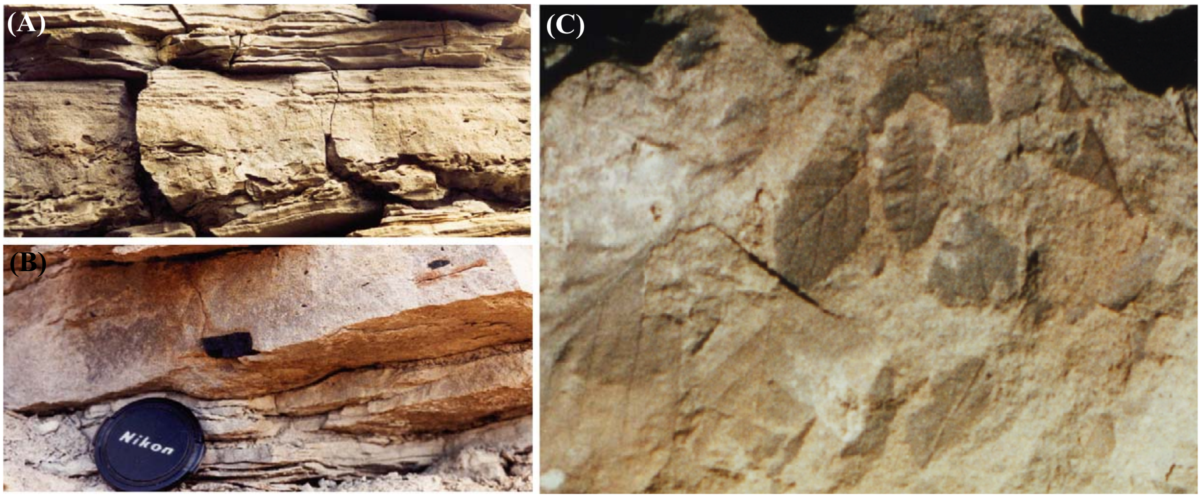

Fig. 10. Coal fragments, up to 5 cm in diameter are abundant in Spitsbergen turbidite beds (A, B). In places, the coal fragments are align basal erosion surfaces or internal scours (A). In other places, they occur randomly distributed throughout the beds (B). Plant fragments, as well as whole leaves (C) occur in several places.

ments in fine-grained beds and as pebble-cobble size particles in coarser-grained beds. Large particles are especially abundant at the bases of channels and canyon-fill units. Whole leaves and leave fragments are found in slope turbidites and in the basin-floor fans. Both the coal layers and leaves are plentiful in the coastal-plain successions, but rare in the shoreline-shelf deposits.

3.5.1. Interpretation

Hyperpycnal turbidity currents generated at river mouths transport sediment directly from the terrestrial area into the ocean. Consequently, both the volume and the preservation of organic matter may be different from turbidites derived from collapse of shelf-edge or upper-slope reaches (Johnson et al., 2001). Abundance of continental fossils, coal, etc., has been suggested to be indicative for direct river input (Mulder and Alexander, 2001; Johnson et al., 2001). Moreover, Johnson et al. (2001) suggest that river-derived underflows may be biochemically significant. Particulate organic carbon may account for 1% of the suspended solids in river water at high sediment concentrations (Ludwig et al., 1996). One underflow can deliver in a day to the sea floor nearly one-half of the vertical flux of carbon from the surface to the bottom measured over an entire year (per square meter). Transport of terrestrial carbon in turbid underflows may provide much of the episodic input that is needed to resolve discrepancies between organic carbon oxidation rates measured in sediments and the rate of supply from vertical flux (Jahnke et al., 1990; Smith et al., 1992, 2001; Smith and Kaufmann, 1999).

3.6. Shelf-edge accretion

Through time, there is discontinuous but systematic accretion of turbidites on the front of the shelf platform that built out from the margin of the Central Basin (Fig. 11). We suggest that the rapid rates of such accretion, commonly up to a kilometer per 100,000 years over a distance of 30 km (Steel and Olsen, 2002), are an argument for direct river feeding of turbidity currents onto the shelf margin. This accretion by sandy turbidites happens preferentially each time the progradational delta systems reach the shelf edge, whereas during retreat of the sediment delivery system (back across the shelf), there may be continued muddy accretion and aggradation. The shelf-edge deltas deliver sand onto the slope in two main ways, through confined routes and in a nearly unconfined manner. (a) During substantial and prolonged drop of sea level below the shelf edge, deltas are cannibalized and slope channel/canyon systems link back directly to shelf distributary channels. Sand can accumulate in the slope channels, but is mainly partitioned/by-

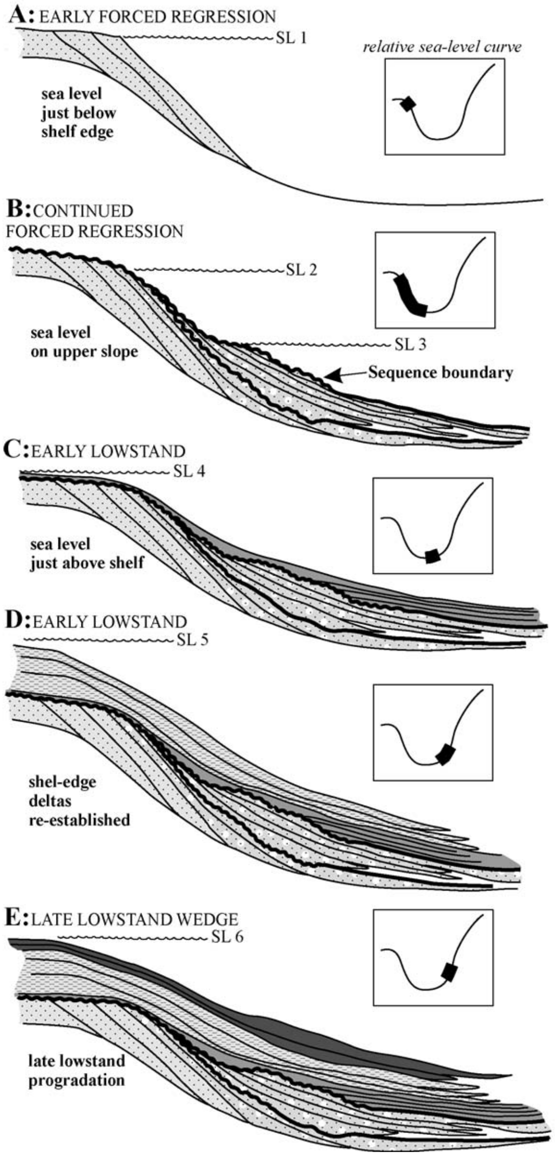

Fig. 11. The slope wedges (Type 2 clinoforms) documented in Spitsbergen, significantly contribute to slope accretion, even during the falling stages of sea level. Whereas the Type 1 systems cause basin-floor aggradation. Development of a slope wedge (Clinoform complex 26 in Fig. 1) in relation to relative sea-level changes (based on Plink-Björklund and Steel, 2002). Note that we now refer to stages D and E as early and late lowstand rather than transgressive systems tract as in Plink-Björklund and Steel (2002).

passed through the slope, and out onto the basin floor (Type 1 clinoforms). At such times there is relatively little slope accretion, but rather aggradation of the basin floor. (b) On the other hand, during times when deltas perch at the shelf edge, or even temporarily below the shelf edge (see Plink-Björklund and Steel, 2002), substantial wedges (up to 70 m thick at their proximal end) dominated by more sheetlike turbidites accumulate on the slope, and little sand reaches the basin floor (Plink-Björklund et al., 2001; Mellere et al., 2002) (Type 2 clinoforms). This is the scenario by which shelf-margin accretion occurs most rapidly. Shelf-margin accretion by the systematic feeding and progradation of slope lobes by small channels and chutes is further detailed below.

3.6.1. Interpretation

There are two principal external forces acting on turbidity currents. Gravity propels sediment-laden fluids downslope, and friction between the base of the flow and the substrate resists the downslope movement. Arrestation of turbidity current requires a decrease in the former or an increase in the latter force. As noted above, the slope chutes are too small to confine the turbidity currents, so the currents overbank the chutes and diverge from them. This lateral dispersion increases the basal area of the current, increasing also friction and thus causing the turbidity current to decelerate. When this happens, deposition starts and the turbidity current loses its density, wanes and eventually dies. However, we know that some turbidity currents on slopes do have the power to erode submarine channels, and even canyons. Sandy (i.e., non-cohesive) turbidity currents must have the ability to erode and entrain sediment from the vicinity of the bed, because turbulence alone is not sufficient to keep non-cohesive materials in suspension (Parker, 1982). The fact that the turbidity currents generated by Eocene shelf-edge deltas on Spitsbergen were only rarely able to erode large channels, and never generated canyons implies that the flows rarely accelerated or entrained enough new sediment. Neither did they reach an ‘‘ignited’’ stable state. Consequently, they did not become self-sustained, highly erosive or competent to scour out submarine canyons (Parker, 1982). Turbidity flows below ignitive state are unstable and invariably die out. The delta-fed turbidity currents on the Spitsbergen shelf margins never ignited, probably because they travelled slowly, were of low density, and sand-rich (see above). This interpretation is consistent with observations on the change in chute size and density within the slope deposits of an individual cycle of relative sea level fall and rise (the determination of this cycle is discussed below). The chutes are largest and most frequent within the steep clinothems generated during relative sea-level fall, whereas they are less frequent in the more gently sloping, aggradational clinothems developed during the beginning of relative sea-level rise.

3.7. Low abundance of slump features or debris flows

Slumped and debris-flow beds form only minor volumes within these Spitsbergen turbidite successions that contain abundant sustained flow beds. In slope clinoform complexes they are practically missing, except the small slumped beds in delta-front and mouth-bar successions (Fig. 12). These slumped or collapsed beds can laterally be followed to their pinch-out (Fig. 12), and into undisturbed beds. The deformed beds are volumetrically small (up to 1 m thick and a couple of tens of meters long) and have not been transported significantly downslope.

In the turbidite systems of Type 1 clinothems where canyons incise the shelf edge and upper slope, slumping was more frequent. In places, the basal erosion surface of the canyons has a morphology consisting of a series of 'steps' separated by steeper 'walls', suggesting that it may originally have formed by a series of slumps that migrated progressively landwards through time (i.e., retrogressive slumping). Also stratigraphically below and lateral to the canyons, slumped features and rotated blocks have been found. Nevertheless, in the canyons, the only preserved slump deposits or debris flows are volumetrically small, and pinch-out within the proximal reaches of canyons. The basin-floor fans downslope from the canyons are stratigraphically similar to the slope systems (Fig. 9C), as well as rich in sandy thick beds and coal fragments.

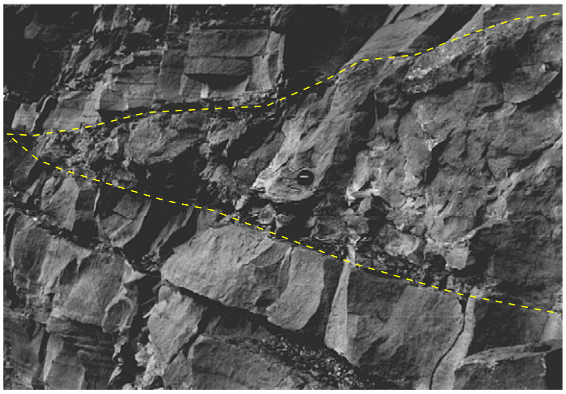

Fig. 12. Slumped or debris-flow beds in Spitsbergen are rare and volumetrically insignificant. The individual slump beds can commonly be traced to their pinch-outs.

3.7.1. Interpretation

Lack of slumps and debris flows suggests that the systems were fed by another mechanism than oversteepening and collapse of delta front or mouth bars. Actually, the mouth bars themselves are rather small and indicate that most of the river effluent debouched directly onto the canyons/slope channels. In the Type 1 successions, the slope collapse may have generated some turbidity currents, or at least contributed with material to the hyperpycnal flows.

4. Discussion

4.1. Characteristics of hyperpycnal flow beds

The following features have been argued to be characteristic for hyperpycnal flow deposits. (1) Great thickness of individual turbidite beds (e.g., Piper and Savoye, 1993; Mulder et al., 1998; Kneller and Buckee, 2000; Mulder and Alexander, 2001), because of a persistent discharge of river effluent (e.g., Wright et al., 1986; Hay, 1987; Prior et al., 1987; Nemec, 1990). The duration of surge-type flows can be a few tens of hours in a rare large event, taking into account the period from the initial slide to final deposition in the most distal part (cf. Hughes-Clark et al., 1990; Mulder and Alexander, 2001). A hyperpycnal flow can persist for days or weeks, depending on the flood duration at the river mouth (Skene et al., 1997; Mulder et al., 1998; Mulder and Alexander, 2001). (2) Repetitive change in the angle of climb in climbing-ripple successions due to effects of the superimposing flood and ebb currents at the river mouth (Mohrig, 2001). However, similar rhythmic changes in hyperpycnal flow beds have been interpreted to indicate passage of internal waves (Nemec, 1995). (3) Hyperpycnal flow beds record variations in flood hydrograph (e.g., Mulder and Alexander, 2001). Documentation of a hyperpycnal flow bed from Saguenay Fjord, Canada (Syvitski and Schafer, 1996; Mulder et al., 1998; Mulder and Alexander, 2001), suggests that the hyperpycnal flow turbidite beds are thick, inversely graded at the base and normally graded at the top, with a uniform thick interval between, corresponding to the depositional phases during the period of increasing of flood discharge, steady flow, and falling discharge. During the period of increasing flood discharge, sediment is carried progressively seaward, and initial deposition comprises mainly sand and coarse silts. The turbidity current deposit starts out being inversely graded. At peak river discharge, basin sedimentation rate is high, reflecting the maximum suspended load carried by the river and maximum erosion of the delta front at the river mouth. Fine particles (clay and fine silt) bypass the proximal zone. Just after the flood culmination, coarser particles are deposited more and more landward and finer particles are deposited seaward, as sediment concentration and flow velocities decrease. This phase leads to the formation of a deposit that thickens and then thins seaward, in contrast to turbidites deposited from an surge-type flow, where deposit thickness simply decreases with distance. As velocity begins to decrease, deposition occurs everywhere, and a normally graded turbidite is formed. The proportion of fine particles in the deposit increases seaward. At the very end of the flood, the plume cannot carry sand, and the initial turbidite deposit is draped by fine-grained sediments. The deposits may range from various sequences in which the inversely graded unit is partly or totally eroded before deposition of a normally graded unit. Mulder and Alexander (2001) also suggest that climbing ripples may be a major sedimentary feature of hyperpycnal turbidites, as they represent the steady migration of sedimentary bedforms while sediment supply is maintained and sedimentation rate is significant. When sediment load decreases, climbing ripples can be replaced by classical asymmetric ripples. Dunes will be less common, as bigger bedforms take longer time to develop.

The Spitsbergen database also suggests strongly that great bed thickness is a likely characteristic feature of hyperpycnal flow turbidites. However, caution must be taken, as the great bed thickness combined with traction structures really only establishes sustained flows, which can be initiated by other mechanisms than direct river effluent. Additional evidence needs to be presented by, e.g., direct connection of turbidite systems to rivers or abrupt pinchouts in sand-prone beds.

The documented hyperpycnal flow beds on Spitsbergen are more complex than the described Saguenay Fjord turbidite, especially as regards the downslope variability within an individual beds. The beds can have an upwards coarsening-to-fining character in places, whereas downslope or upslope, the same bed may show several coarsening-to-fining intervals or be without any vertical grain-size trends. We have not been able to detect any rhythmic changes in the turbidite beds, except the plane-parallel lamination. We suggest that the downslope decrease of lamination in the beds that maintain their thickness for long downslope distances is characteristic for hyperpycnal flow beds.

4.2. *"Pure" hyperpycnal flows?*

It has been argued that the triggering of hyperpycnal flows takes place through the process of undrained delta-front failures linked to high rates of prodelta sedimentation (Gibson, 1958; Sangrey et al., 1979; Mandl and Crans, 1981), or floods resuspending previously deposited mouth-bar material (Prior et al., 1986, 1987; Zeng et al., 1991). In the Spitsbergen database, we have documented that slope failure does not significantly contribute to turbidite initiation in the Type 2 turbidite systems, even though erosion and entrainment of mouth-bar material by the flooding river effluent cannot be excluded.

In Type 1 systems where canyons are cut into the slope, by definition, slope instability has had significance, i.e., failure of unconsolidated sediments at the shelf edge in connection to river floods may have been important. However, the behaviour of depositing turbidites both in chutes and canyons was controlled by the river effluent, as the canyon- and chute-fills close to the river/distributary mouth display thick laminated sandstone beds that were deposited by gradual aggradation during days or weeks. The additional material eroded from moth bars or added by minor slope failures may have contributed to the initial density of river-generated flow in Type 1 systems, and enabled the flows to ignite. Whereas, in Type 2 systems, the accumulation of turbidites on the slope by itself indicates that the turbidity currents were rather slow and dilute, and never ignited (e.g., Parker, 1982).

Wright et al. (1986) suggest that there are low- and high-density hyperpycnal plumes. The low-density plumes have an excess density with magnitude of 10^{-2} to 10^{-3}, and the high-density plumes of 10^{-1}. They suggest that the high-density plumes are able to erode deep channels and canyons, and they have enough density to accelerate and ignite. Whereas the low-density plumes are widespread, they do not erode significantly, nor have they enough density to ignite. The latter flows loose their load and die out. Wright et al. (1986) also suggest that the low-density flows are initiated by "normal" river effluents, whereas the high-density plumes are created by flood-stage loads.

The Type 1 turbidite systems on Spitsbergen are fed by rivers that debouch directly into the canyons, whereas the Type 2 systems are fed by delta distributaries that debouch onto the slope. River loads are larger and may produce a more concentrated effluent compared to the distributary channels. This difference in initial density may be an alternative explanation why the turbidites fed into canyons make it to the basin floor, whereas the turbidites off the deltas are deposited on the slope.

4.3. Impact of sea-level changes

The growth of turbidite systems connected with rivers occurs when fluvial systems can discharge directly at the shelf break (Kneller and Buckee, 2000). This direct link is not limited to, but is most often associated with sea level lowstands (Mutti, 1985; Shanmugam et al., 1985; Mutti and Normark, 1991; Posamentier et al., 1991; Normark et al., 1993; Steel et al., 2000; Plink-Björklund et al., 2001; Plink-Björklund and Steel, 2002). Some workers (Burgess and Hovius, 1998) have suggested that this link may also occur during highstands if rates of delta progradation are sufficiently high.

Mulder and Syvitski (1996) have argued that during a sea level fall, failure-related turbidity current activity increases, whereas hyperpycnal flow activity decreases, because rivers merge and form large mega-rivers with huge drainage areas, and rivers with large average discharge are not likely to produce hyperpycnal plumes as a consequence of their low concentration to water discharge ratio (Mulder and Syvitski, 1995). Large rivers favour active delta construction and potential slope failures (Mulder and Syvitski, 1995). This idea is opposed by Damuth et al. (1988), who suggested that sinuous channels on the Amazon Fan, which developed during late Pleistocene lowstands, may have been formed by continuous underflows (see also Hay, 1987; Kneller and Buckee, 2000).

Our data set strongly suggests that turbidites were generated preferentially during forced regressive and lowstand times in both Type 2 and Type 1 clinothems, especially when sea level fell below the shelf edge (Steel et al., 2000; Plink-Björklund and Steel, 2002). River loads can go hyperpycnal at any time. However, the sediment has a much higher probability to make it to the slope or basin floor when the rivers unload directly onto the slope.

5. Conclusions

Following criteria documented in a detailed study of turbidite deposits from seismic-scale dip-oriented outcrops of Eocene Central Basin collectively suggests that hyperpycnal flows were significant in generating the turbidite successions:

1. Physical connection from rivers to canyons and distributary channels to slope channels (chutes) indicates that at times the fluvial feeder system was dumping sediment directly onto the slope.
2. Abundance of thick sandstone beds with traction structures reflects deposition by gradual aggradation from sustained turbidity currents.
3. Abrupt pinch-out of thick sandy turbidite beds are caused by the buoyancy effects and flow lofting due to the interstitial fresh water in turbidites.
4. Sand-prone beds were generated by decoupling of coarse and fine populations due to the separation of hyperpycnal and hypopycnal plumes.
5. High abundance of leaves and coal fragments suggests direct river input into the basin.
6. Shelf-edge accretion suggests systematic input from dilute hyperpycnal flows that did not ignite.
7. Low abundance of slumps and debris flows confirms that the turbidites were generated by a different mechanism than slope instability.

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