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하도 주변부 퇴적층의 구성요소에 대한 연구 : 마이오세 포항 분지의 도음산 선상지－삼각주

A Study on the Architectural Elements of Channel Margin Deposits : Doumsan Fan-Delta in the Miocene Pohang Basin

1996. 2

한 국 해 양 연 구 소

제 출 문

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본 보고서를 '하도 주변부 퇴적층의 구성요소에 대한 연구: 마이오세 포항 분지의 도음산 선상지-삼각주' 사업의 최종보고서로 제출합니다.

1996년 2월

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요 약 문

Ⅰ. 제 목

하도 주변부 퇴적층의 구성 요소에 대한 연구: 마이오세 포항 분지의 도음산 선상지-삼각주

Ⅱ. 연구 내용 및 결과

마이오세 포항 분지의 도음산 선상지-삼각주는 바다에 형성된 길버트 형태의 역질 선상지-삼각주로 그 높이가 150 m 이상, 경사가 20^0 이상 되는 대규모 전면층으로 특징지워진다. 정밀 야외 조사에 따르면 전면층은 여섯 개의 구성 요소 즉 퇴적상으로 구성된다. 퇴적상 A는 다양한 형태의 점이 층리와 층의 3차원적 모양 (Bed Geometry)을 보이는 두꺼운 사질역으로 구성된다. 퇴적상 B는 두께가 비교적 얇고, 역점이 층리가 잘 발달한 사질역으로 이루어진다. 퇴적상 C는 두께가 역 서너개의 높이 보다 얇은 판상역층이며, 퇴적상 D는 수평 연장성이 불량한 렌즈 형태의 역층이다. 퇴적상 E는 매우 크기가 큰 역을 포함하는 얇은 모래층이며, 퇴적

상 F는 매우 두꺼운 괴상의 사질역층이다. 이러한 6개의 퇴적상은 주로 점성이 없는 쇄설류 (퇴적상 A와 B), 쇄설물 낙하 (퇴적상 C와 D), 난류성 유체 (퇴적상 E) 및 사태 (퇴적상 F) 등에 의한 퇴적을 지시한다.

6개의 퇴적상이 서로 밀접하게 연관된다는 사실은 퇴적상들이 중력류가 발생하여 경사면을 따라 흐르는 동안 여러 형태의 유체로 변해간다는 점을 시사한다. 점성이 없는 쇄설류가 선상지-삼각주의 topset과 전면층의 경계 지역 또는 전면층의 경사면에서 발생하여 전면층의 경사를 따라 흐르게 되면 쇄설류는 내부에 특성이 서로 다른 두개의 유체로 나뉘게 된다. 하부 유체는 주로 역에 의해 지배되나, 상부 유체는 사질 입자로 주로 구성되며 역이 일부 포함된다. 상부 유체는 쉽게 변형되어 사질의 난류성 유체 또는 쇄설 낙하를 형성하며, 주로 퇴적상 D 또는 E를 초래한다. 점성이 없는 쇄설류는 때로 일련의 얇은 유체로 변형되어 퇴적상 A층 주위에 퇴적상 B가 형성되는 현상을 야기한다. 경사면을 따라 보다 낮은 곳으로 흐르는 무점성 쇄설류 내에서는 포함된 모래가 역 사이로 빠져 유체 하부로 떨어지거나, 주변 물에 의해 유체 밖으로 제거된다. 결과적으로 모래 기질이 제거되어 역만으로 구성된 쇄설 낙하로 변형된다. 이러한 쇄설 낙하는 유체가 지속되는 시간에 좌우되며, 두께가 다양한 판상의 역층의 퇴적을 초래한다. 일반적으로 크기가 큰 역은 운동량이 크며, 기저층과의 마찰이 작아 작은

역에 비해 먼거리로 이동되어 경사면의 낮은 곳에 퇴적된다. 이러한 유체 변형에 따른 진화 모델은 선상지-삼각주 앞에 발달하는 Prodelta 퇴적층이 삼각주와는 달리 주로 사질층에 의해 지배되며 판상의 역층이 협재되는 입도의 양극성 현상을 설명하는데 도움을 줄 수 있을 것이다.

SUMMARY

I Title

A Study on the Architectural Elements of Channel-Margin Deposits: Doumsan Fan-Delta in the Miocene Pohang Basin

II. Results

The Doumsan fan-delta in the Miocene Pohang Basin, SE Korea is a marine gravelly Gilbert-type fan-delta with large-scale foresets that are more than 150 meters high and dip at about 20^0. The foreset deposits consist of six sedimentary facies: medium- to thick-bedded sandy gravel deposits with variable grading patterns and bed geometries (Facies A), thin- to medium-bedded, commonly inversely graded sandy gravel deposits (Facies B), less than a few grain-thick and sheet-like layers of densely packed pebble gravel (gravel sheets; Facies C), less than a few grain-thick and lensoidal layers of cobble to boulder gravel (gravel lenses; Facies D), thin-bedded sand with outsized clasts (Facies E), and very thick disorganized sandy gravel deposits (Facies F). These facies indicate deposition from cohesionless debris flows (Facies A and B), debris fall (Facies C and D), turbulent flows (Facies E) and slumps (Facies F).

Intimate relationships among these facies suggest that they resulted from an evolving sediment gravity flow. A cohesionless debris flow generated at the topset-foreset boundary or on the middle of the foreset slope segregates its sediments into a pebble-supported lower

division and an upper sandy division with outsized clasts. The sediments in the upper division are easily remobilized into a sandy turbulent flow and bouldery debris fall, resulting in Facies D and E deposits downslope. The cohesionless debris flow occasionally transforms into a series of thinner flows, producing Facies B deposits around a Facies A bed. When the cohesionless debris flow moves further downslope, interstitial sand percolates downward or is stripped away by ambient fluid. The flow then transforms into grain-assemblage debris fall and single-grain debris fall, depending on the flow duration, producing gravel sheets of variable thicknesses. Larger clasts are transported farther downslope due to their larger momentum and smaller friction with the bed. The above model for flow evolution explains the origin of "textural bimodality" and other characteristics of prodelta deposits, which consist of isolated clasts or lensoidal deposits of cobble-to-boulder gravel in sandy materials.

CONTENTS

LIST OF FIGURES

INTRODUCTION

Gilbert-type deltas, consisting of topsets, foresets and bottomsets, are produced by progradation of alluvial or fluvial systems into a standing body of water, either lacustrine or marine, usually in a basin with steeply inclined margin (Gilbert, 1885). They are characterized by very steep depositional surfaces, i.e., foresets that are inclined near the angle of repose and dominated by gravity-driven processes. Sand-dominant Gilbert-type deltas include thin-bedded, graded to planar- or ripple-cross-laminated foreset beds that are laterally continuous, distinctly bounded and intercalated with silt layers. These features suggest deposition from dilute underflows or turbidity currents (Gustavson et al. 1975; Stanley and Surdam 1978; Dunne and Hempton 1984; Flores 1990). On the other hand, gravel-dominant Gilbert-type deltas comprise crudely stratified foreset beds of which the individual layers are difficult to discern because of their lateral pinching-out and abrupt changes in grading pattern and grain fabric. These features have hampered detailed facies analysis of the gravelly foreset deposits, although their general characters suggest important roles of high-concentration flows such as cohesionless debris flows, modified grain flows, high-density turbulent flows, and grain avalanches (Postma 1984; Postma and Roep 1985; Colella et al. 1987; Postma and Cruickshank 1988; Hwang and Chough 1990; Martini 1990; Massari and Parea 1990; Mastalerz 1990; White 1992).

Small-scale gravelly foreset deposits which are less than a few decameters high exhibit similar features to those of sandflow cross-strata in aeolian and subaqueous dunes (e.g., Hunter 1977, 1985; Buck 1985), such as predominance of inverse grading, lack of mud matrix, common presence of well-sorted or openwork gravel layers and lenses, and coarsening of clast size toward the downdip margin of a bed or the base of the foreset slopes, and have been interpreted basically in terms of the grain flow theory of Bagnold (1954). A straightforward application of the sandflow facies and processes is, however, of limited relevance on large-scale gravelly foreset slopes (> 100 m in height) where flow characters are liable to be more complex and spatially variable due to larger chances of flow transformations. In this paper, we describe a large-scale gravelly foreset deposits in an attempt to reveal how foreset processes transport, segregate and deposit sediments, evolve spatially and produce various sedimentary features on and beyond the foreset slopes.

The Doumsan fan-delta in the Miocene Pohang Basin, SE Korea is a marine gravelly Gilbert-type fan-delta, of which the foreset deposits have a thickness of at least 150 meters (Hwang and Chough 1990; Hwang et al. 1995). It is supposed that the foreset slope has extended for several hundred meters with a slope angle of about 20^0. The foreset deposits consist of monotonous succession of steeply inclined, well to crudely stratified, laterally pinching-out, commonly amalgamated gravel deposits with very few intercalations of traceable fine-grained layers. Detailed clast-by-clast and bed-by-bed observations reveal, however, that the foreset deposits can be represented by six sedimentary facies. We suggest that these facies are genetically related with one another and can be interpreted in terms of evolving sediment gravity flows. We also suggest that the characteristics of prodelta deposits are strongly affected by the foreset processes.

DOUMSAN FAN-DELTA

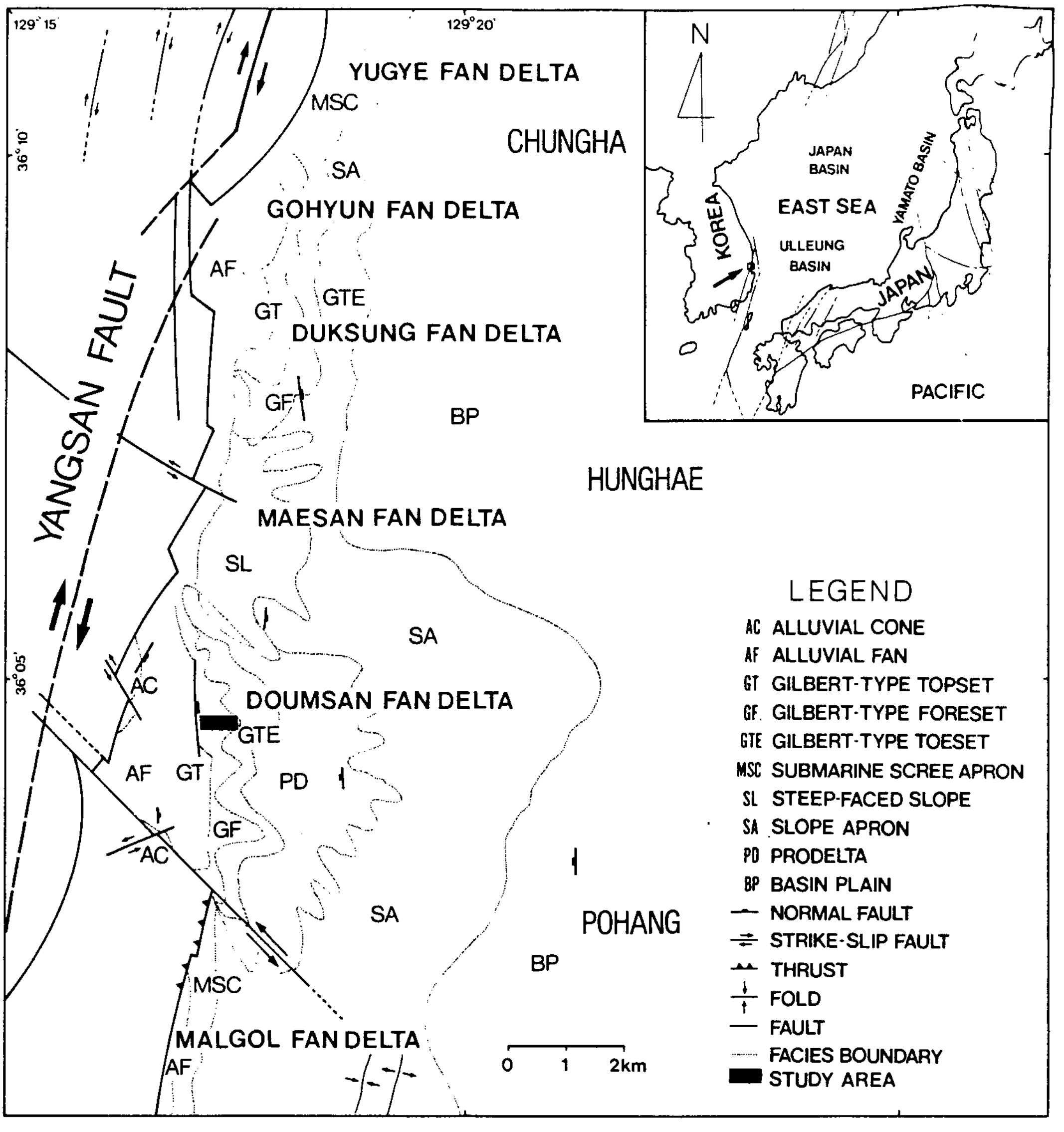
A number of small Tertiary sedimentary basins occur along the southeastern margin of the Korean Peninsula (Fig. 1A). The Pohang Basin is the largest of these basins and has been a target of intense sedimentological research (Choe and Chough 1988; Chough et al. 1990; Hwang and Chough 1990, Choe 1990; Hwang 1993; Hwang et al. 1995). The basin was produced by pull-apart opening along the Yangsan Fault which experienced dextral strike-slip movements during Miocene time (Yoon and Chough 1995). The basin was filled by more than 1-km-thick sequence of gravel, sand and mud deposits, of which coarse clastic sediments are spatially organized into six fan-delta systems (Hwang 1993; Hwang et al. 1995). The Doumsan fan-delta, the largest of these systems, comprises an entire spectrum of basin margin facies, deposited in alluvial fan, braided stream, transition zone, submarine Gilbert-type foreset, toeset, prodelta (bottomset), slope apron and basin plain environments (Fig. 1A).

The foreset deposits of the Doumsan fan-delta are characterized by steeply inclined (ca. 20^0) beds of massive, graded and stratified sandy gravel deposits. Individual sedimentation units are generally thin and laterally impersistent although they are difficult to discern in many cases. The foreset deposits are more than 150 m thick, prograded for about 1 km with a slope angle of about 20^0 and show a radial paleoflow pattern from the point of sediment origin located to the west of the Doumsan mount. The foreset deposits were formerly interpreted as resulting from small-scale slides (gravity slide or grain flow) with minor influence of density-modified grain flows, debris flows and high-density turbidity currents (Hwang and Chough 1990). At the base of the foresets and toward the toesets, outsized boulders, which probably rolled and slid along the steep

Fig. 1. A) Map showing the locations and paleoenvironmental zonations of six fan-delta systems in the Pohang Basin, SE Korea. The location of the Pohang Basin is indicated by an arrow in the inset figure.

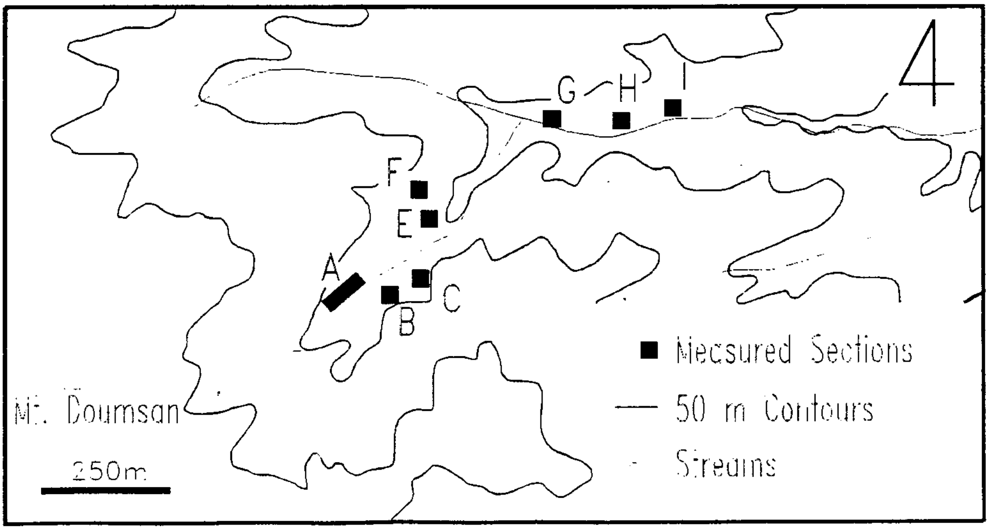
B) Location map of measured sections along an E-running valley in the central part of the Doumsan fan-delta.

A



129° 15'
129° 20'
36° 10'
36° 05'
YUGYE FAN DELTA
MSC
CHUNGHA
SA
GOHYUN FAN DELTA
AF
GT
GTE
DUKSUNG FAN DELTA
GF
BP
YANGSAN FAULT
HUNGHAE
MAESAN FAN DELTA
SL
SA
AC
DOUMSAN FAN DELTA
GTE
AF
GT
PD
GF
AC
SA
POHANG
BP
MSC
MALGOL FAN DELTA
AF
0
1
2km
N
JAPAN BASIN
YAMATO BASIN
EAST SEA
KOREA
ULLEUNG BASIN
JAPAN
PACIFIC
LEGEND
AC ALLUVIAL CONE
AF ALLUVIAL FAN
GT GILBERT-TYPE TOPSET
GF GILBERT-TYPE FORESET
GTE GILBERT-TYPE TOESET
MSC SUBMARINE SCREE APRON
SL STEEP-FACED SLOPE
SA SLOPE APRON
PD PRODELTA
BP BASIN PLAIN
NORMAL FAULT
STRIKE-SLIP FAULT
THRUST
FOLD
FAULT
FACIES BOUNDARY
STUDY AREA

B



4
G
H
I
F
E
A
B
C
Measured Sections
50 m Contours
Streams
Mt. Doumsan
250m

slope, are commoner; thin muddy layers due to hemipelagic settling are intercalated; and thick gravel deposits related to slope failures are also commoner.

FORESET FACIES

Interpretation of depositional processes and spatio-temporal evolution of a sediment gravity flow from ancient deposits is facilitated when vertical and lateral organizations of facies characteristics within a single sedimentation unit are documented. Identification of a single sedimentation unit is therefore a starting point for many sedimentological interpretations (e.g., Clifton 1984). Unfortunately, however, it is not easy to identify a sedimentation unit in a thick succession of amalgamated gravel deposits. To overcome this problem, we have selected several well-exposed outcrops in the foreset deposits of the Doumsan fan-delta (Fig. 1B). Most outcrops, observed along an E-running valley, are oriented roughly parallel to the dip directions, which probably coincide with the transport directions. The selected exposures were observed in very detail to discriminate every sedimentation unit present. After that, vertical and lateral variations in sedimentary characteristics within each unit were sketched clast-by-clast.

Six sedimentary facies are classified based mainly on grain size and gravel content, layer thickness and bed geometry. Sedimentary structures are not a useful criterion because most deposits show variable grading patterns and structures laterally within a single bed. Facies A and B refer to gravelly sand or sandy gravel deposits with variable bed continuity in which sand matrix accounts for a major portion. They are distinguished from each other mainly by their bed thicknesses. On the other hand, Facies C and D refer to gravel layers less than several-grain-thick, in which the amount of sand matrix is small and insignificant. They are distinguished from each other by subtle differences in dominant clast size and lateral continuity. Facies E refers to thin sand deposits with sparse gravel clasts. All of these facies are neither discrete nor mutually exclusive, and more than one facies units can occur within a single sedimentation unit. For example, a thick layer of Facies A, overlain by a gravel lens of Facies D or thin sand of Facies E, may pass downdip into several thin layers of Facies B (see FACIES RELATIONSHIPS for more discussion).

Facies A: Medium- to thick-bedded sandy gravel deposits

Description. This facies is the most abundant in the Doumsan foreset deposits and refers to sandy gravel deposits that are several decimeters to one meter thick (Figs. 2-5). Beds thicker than one meter are rare. The gravel clasts are mostly pebble-size and supported by themselves. Cobble- to boulder-size clasts are also present, but in subordinate amounts. In many beds, a(p)a(i) imbrication of clasts is well developed. The matrix consists of poorly to well-sorted, fine- to coarse-grained sand. Beds of this facies are variously graded and commonly comprise more than one type of grading within a single bed (Figs. 2, 3A). Inverse grading is predominant, however. Inversely graded beds show either an upward increase in clasts content and size (Figs. 2, 3A, 4B) or concentrations of coarsest clasts along the top (Figs. 3A, 6A). In the former, inverse grading is laterally impersistent and the gravel concentration zones appear as discontinuous gravel lenses of Facies D on poorly exposed outcrops. In the latter, protruding clasts are either isolated (arrows 1 and 3; Fig. 3A) or in contact with one another, forming single grain-thick gravel clusters (arrow 2; Fig. 3A). The inverse grading results in irregularities of upper bed

Fig. 2. Sketch, columnar logs and photograph of a relatively thick and continuous bed at section A (see Fig. 1B for location), characterized by vertical and lateral variations in clast size and content. The pebble-supported lower division (Facies A) consists of two wedge-like subunits and is overlain by a thin sandy layer with outsized clasts (Facies E). The wedge-like subunits are inversely graded and fining toward the downdip margins. Downdip margin of the overlying subunit (wedge 1) is stratified. The bed dips at about 20^0 toward the east. Orientation of the outcrop relative to the dip direction is indicated in the upper left.

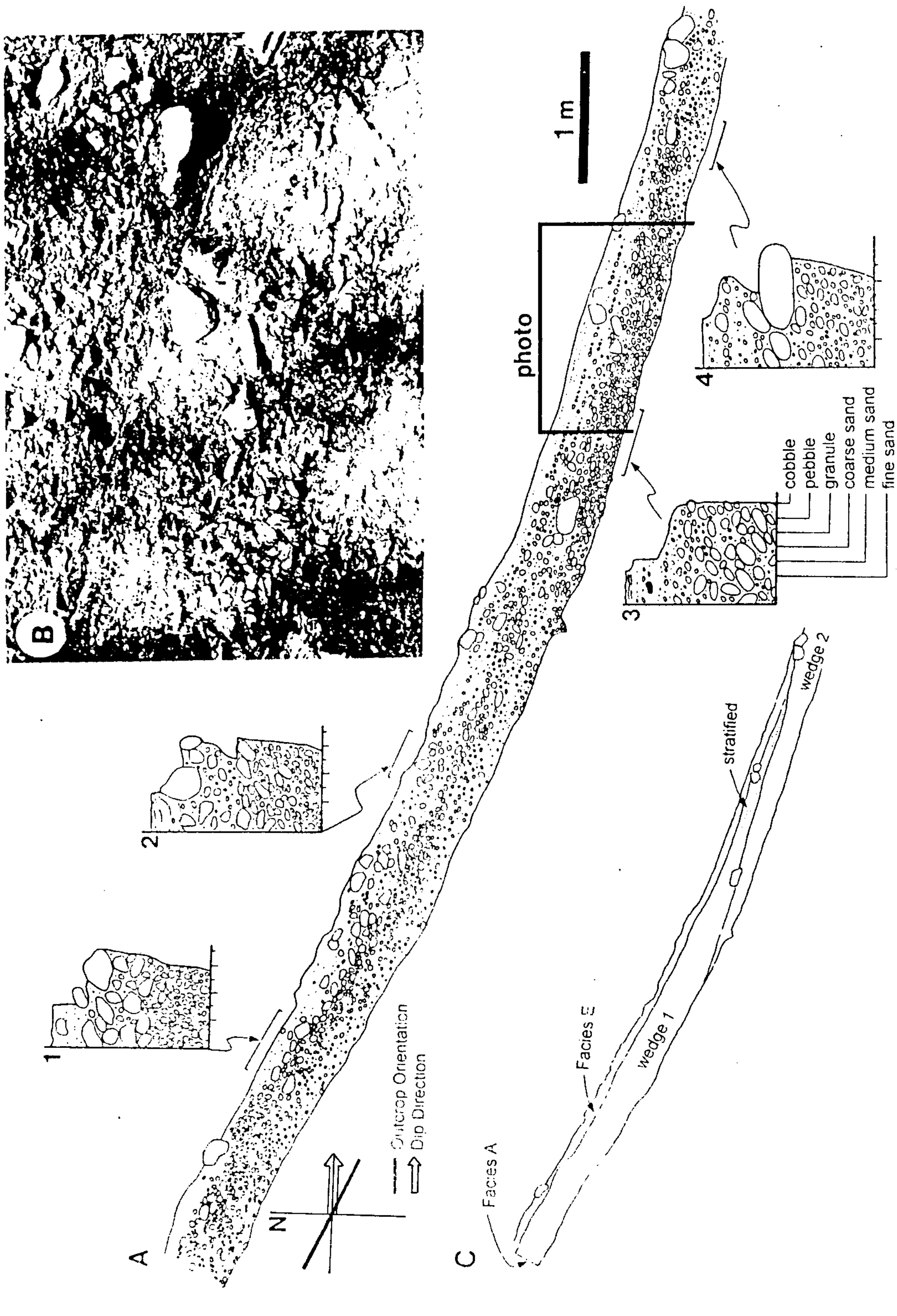
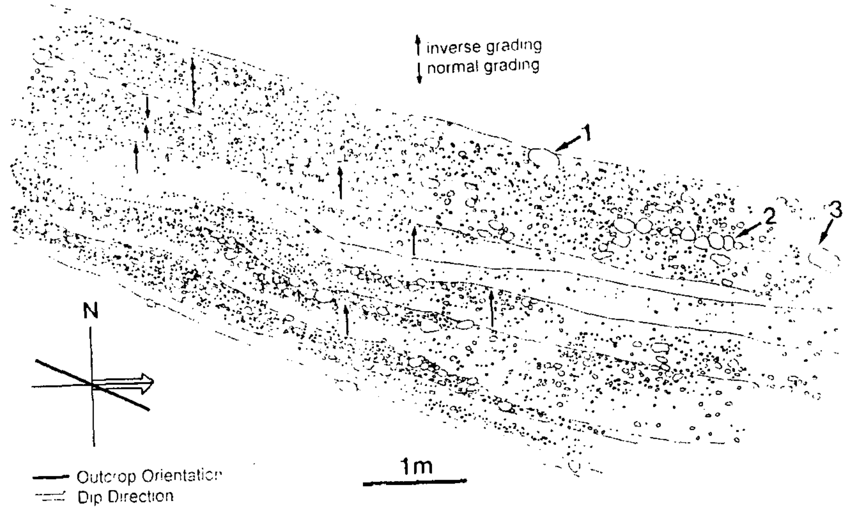

A
B
C
N
Outcrop Orientation
Dip Direction
1
2
3
4
photo
1 m
cobble
pebble
granule
coarse sand
medium sand
fine sand
Facies A
Facies E
wedge 1
wedge 2
stratified

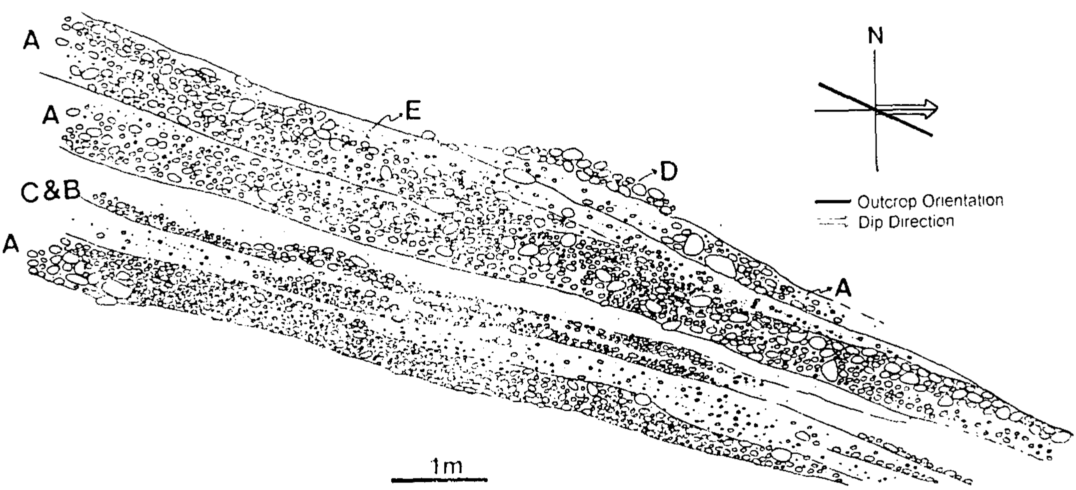
Fig. 3. Outcrop sketches at section A. A) shows laterally persistent or wedge-like beds of Facies A that are generally inversely graded and have protruding clasts, either isolated (arrows 1 and 3) or clustered (arrow 2). The bed boundaries are non-erosional but slightly undulatory due to hummocky primary reliefs of some beds. B) shows several beds of Facies A in the upper part and gravel sheets of Facies C in the lower part. The gravel sheets are thin layers of densely packed gravel clasts with minimal amounts of sand matrix, whereas the beds of Facies A comprise abundant interstitial sand. The lowermost unit is intermediate in character between Facies A and C, showing densely packed gravel clasts but with interstitial sand matrix. Capital letters indicate facies codes.

A



inverse grading
normal grading
1
2
3
N
Outcrop Orientation
Dip Direction
1m

B



A
A
C&B
A
E
D
A
N
Outcrop Orientation
Dip Direction
1m

Fig. 4. A) Sketch of section E, showing several beds of Facies A, which pass downslope into stratified deposits (Facies B). B) Sketch of section I, consisting of relatively continuous and thick bouldery beds of Facies A which are intercalated with sand beds containing outsized clasts (Facies E). Capital letters indicate facies codes.

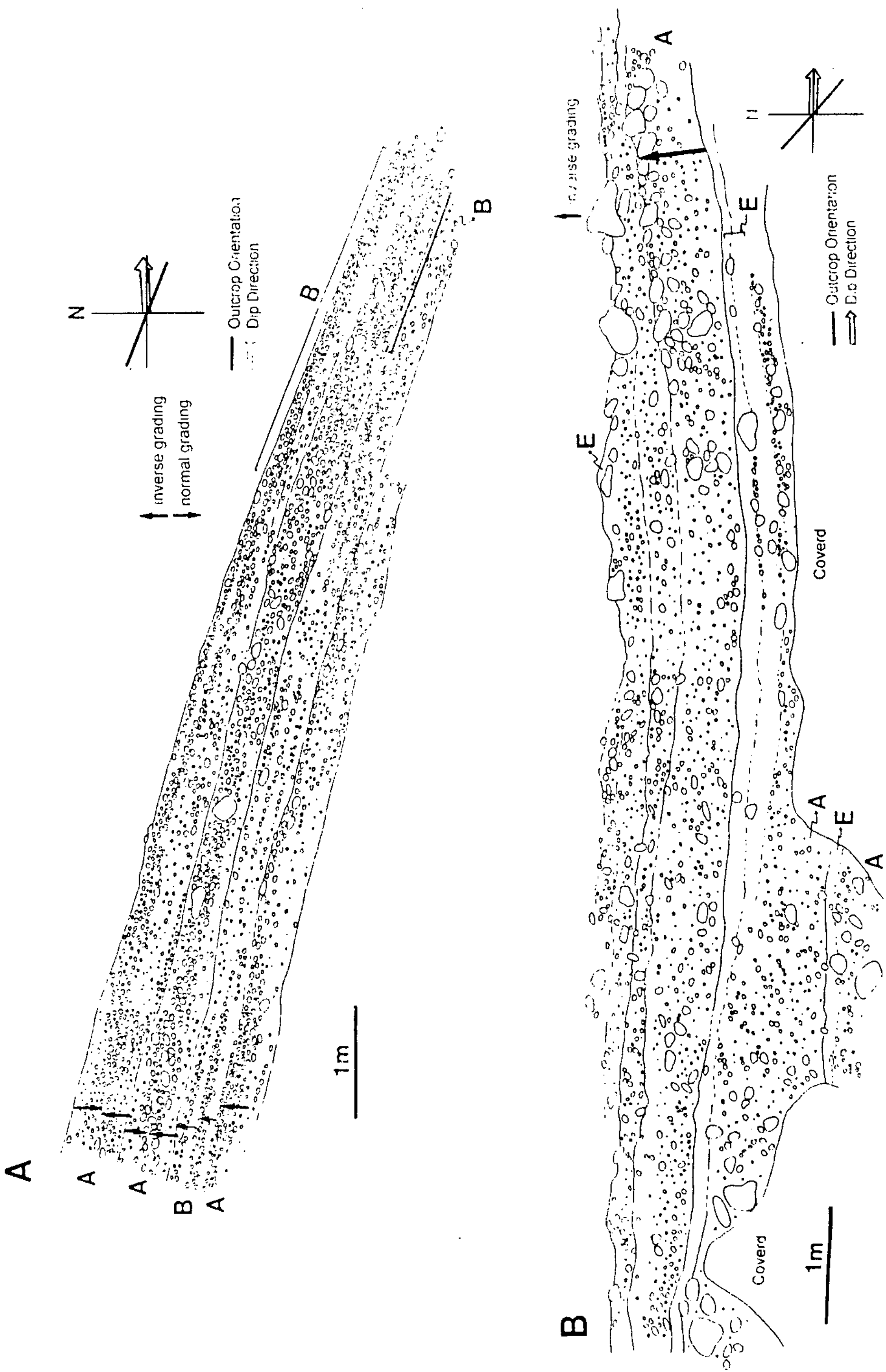

A
inverse grading
normal grading
N
Outcrop Orientation
Dip Direction
A
A
B
A
B
B
1m
B
inverse grading
N
Outcrop Orientation
Dip Direction
E
E
A
E
A
A
Coverd
Coverd
1m

Fig. 5. Sketch of transverse outcrop at section C, showing laterally pinching-out beds of Facies A and some gravel sheets (Facies C). Beds of Facies A have lateral margins that are inclined-stratified relative to the underlying bed boundaries.

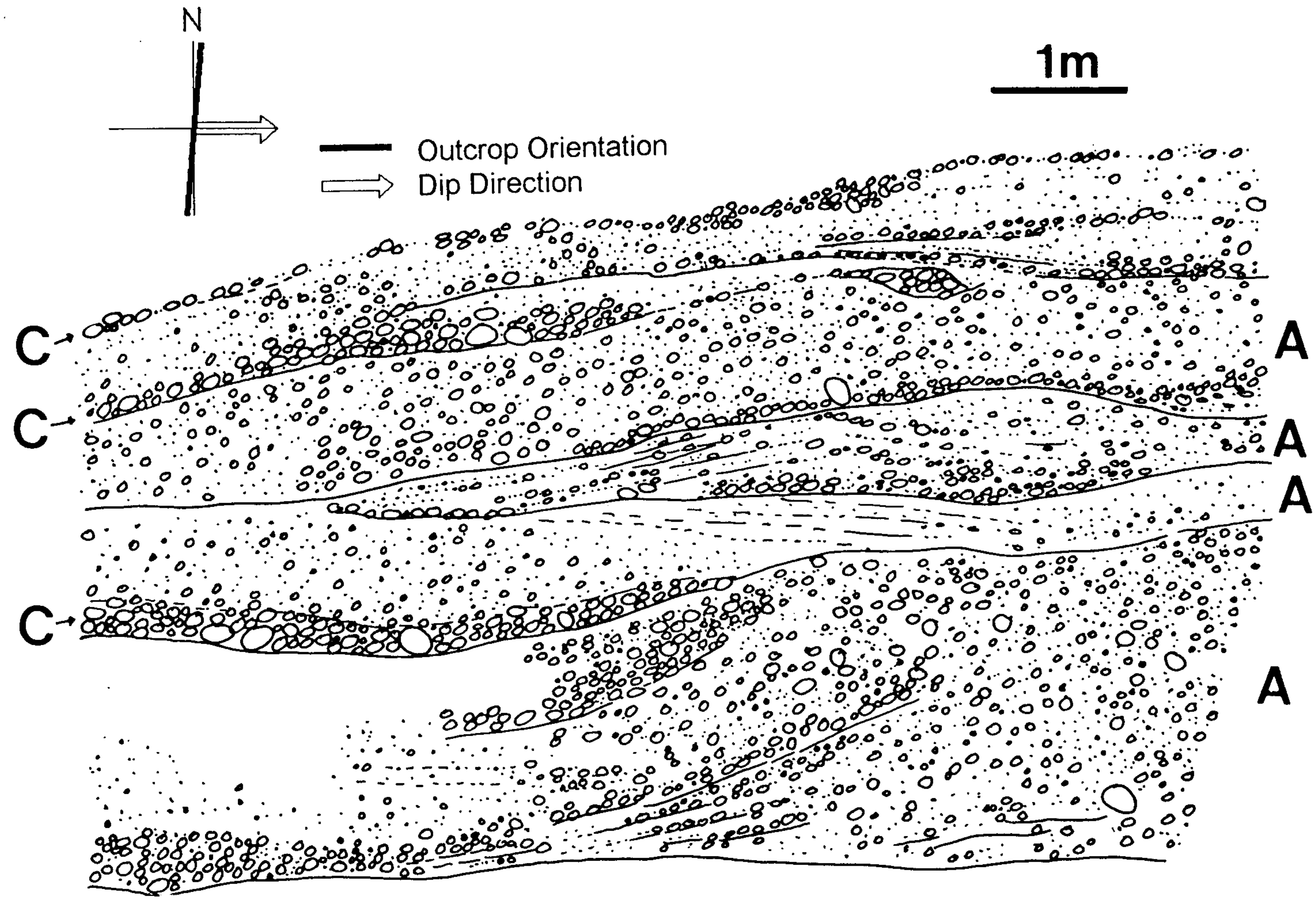
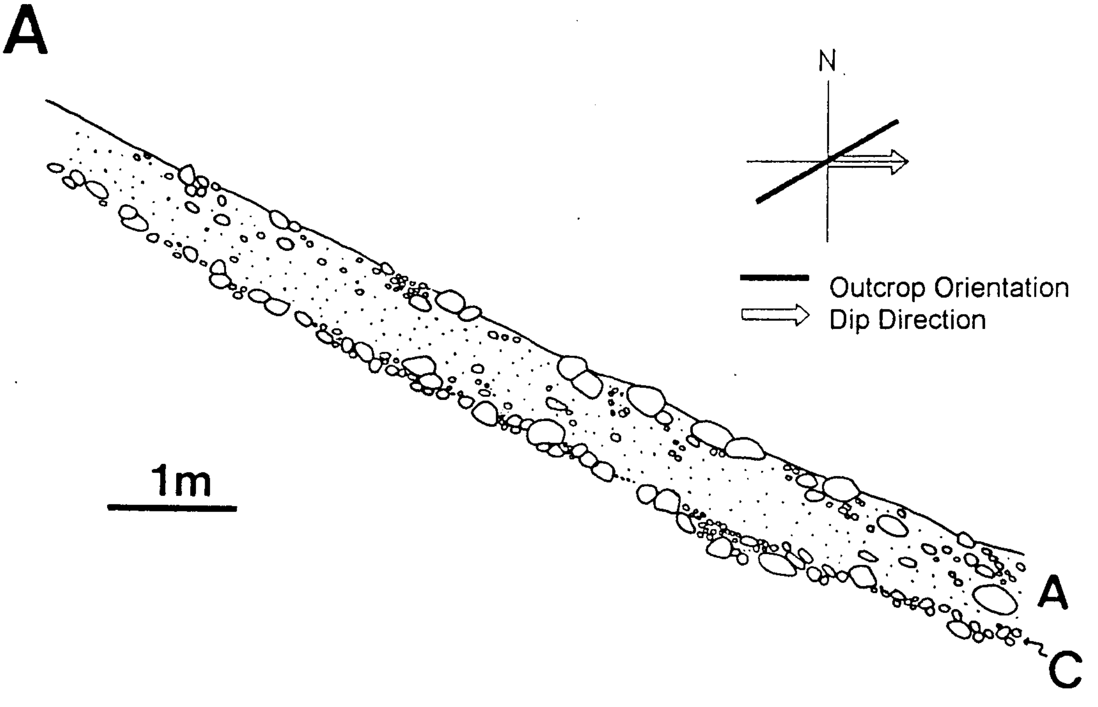
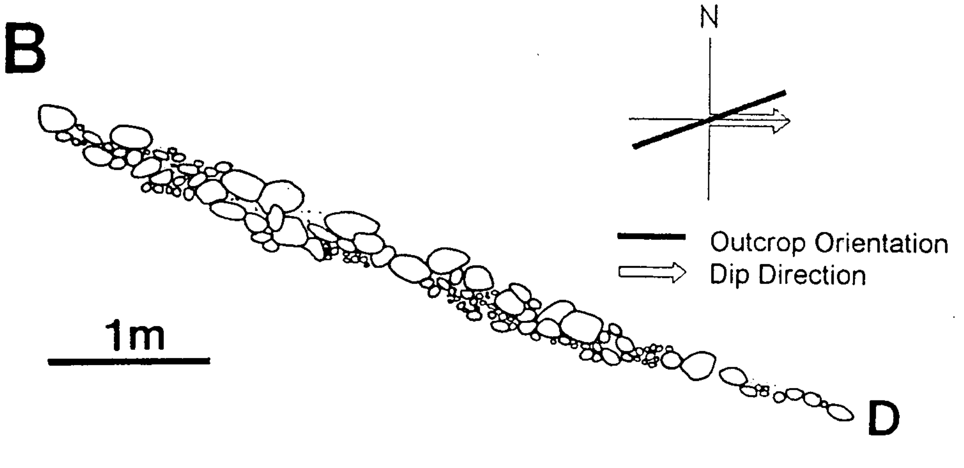
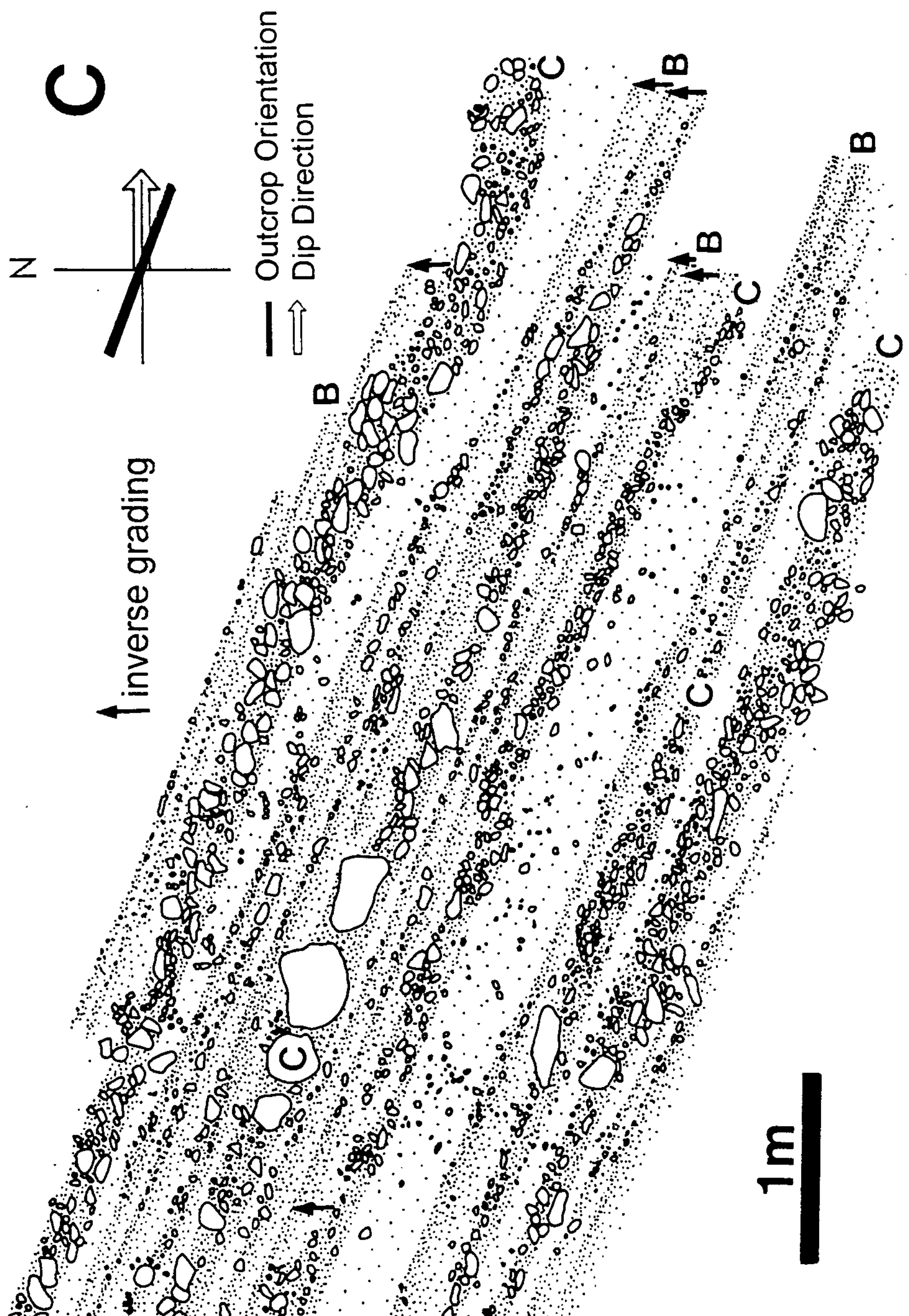

N
1m
Outcrop Orientation
Dip Direction
C
C
C
A
A
A
A

Fig. 6. A) Sketch of a Facies A bed (section G) which is inversely graded by large floating clasts. The underlying gravel sheet (Facies C) is thin and continuous and comprises some cobble- or boulder-size clasts. B) Sketch of a gravel lens at section H, characterized by lensoidal geometry. C) Sketch of well stratified deposits at section F which consist mainly of thin inversely graded beds of Facies B and gravel sheets of Facies C. A cluster of bouldery gravel clasts (Facies D) is present in the middle left part. Photograph of the section is shown in Fig. 7B.


A
N
Outcrop Orientation
Dip Direction
1m
A
C


B
N
Outcrop Orientation
Dip Direction
1m
D


C
N
Outcrop Orientation
Dip Direction
inverse grading
B
C
1m

boundaries, either hummocks of many gravel clasts or protrusions of single clasts. The lower contacts are mostly non-erosive, following the irregular upper surfaces of underlying beds. Beds of this facies show variable continuity, either laterally persistent over an entire outcrop (> 10 m in lateral extent) (Figs. 2, 4B) or pinching out within a few meters (Fig. 3A,B). On transverse sections, beds show more pronounced discontinuity and biconvex or planar-convex bed geometry (Fig. 5). This suggests that many of the beds have a tongue-like geometry in three dimensions.

Detailed observation of a relatively continuous bed (Fig. 2) reveals that the bed is composite, consisting of two overlapping wedge-shaped units. The wedges are inversely graded and become finer-grained and stratified toward the downdip margins. Superposition of the finer-grained frontal part of the overlying wedge above the coarse-grained rear of the underlying one produced a 'false' normal grading (column 3; Fig. 2). A large clast in the middle part of the bed is not one floating in the middle of the bed, but one protruding above the upper surface of the underlying wedge.

Beds of this facies commonly pass downdip into thin-bedded or stratified deposits of Facies B (Fig. 4B). On transverse sections, lateral margins of some beds show inclined stratification relative to the lower contacts, resembling cross-stratification (Fig. 5). Some beds comprise thin upper sandy divisions (Facies E) with sparse pebble- to boulder-size clasts (Figs. 2, 4B).

Interpretation. Common inverse grading of gravel clasts in a sand matrix suggests a cohesionless sediment flow in which clast interactions were dominant. Steep bed attitudes (ca. 20^0) suggest that the flow was mainly driven by gravity-induced shear stress. Undulatory upper surfaces with large floating clasts and positive reliefs suggest plastic flow behavior with debris static strength, which is related to frictional rather than cohesive strength. The sediment flow can therefore be described as a cohesionless debris flow or a modified grain flow (Lowe 1976, 1979, 1982; Nemec and Steel 1984; Postma 1986; Postma et al. 1988). The term 'modified' is labeled because the sandy interstitial material must have provided a supplementary grain-support mechanism via increasing buoyancy, viscosity and excess pore pressure (Lowe 1976; Pierson 1981). The lateral variations of grading patterns suggests flow unsteadiness and changes in flow rheology along the length of the flow (Kim et al. 1995).

Although similar deposits to this facies have been interpreted as involving high-density turbulent flows by some authors (e.g., Clifton 1984; Colella et al. 1987; Postma and Roep 1985; Postma and Cruickshank 1988; Postma et al. 1988), we do not find any reliable evidence of turbulent flows such as traction structures, well-developed normal grading or scour surfaces between beds. The thin sandy divisions, occasionally present above some beds, may have been deposited from a turbulent flow, but it is not likely that the flow provided sufficient shear stress to the underlying coarse-grained flow.

Relative thinness and common lateral pinching-out of individual beds, especially on transverse sections, suggest that the flow was of small-volume and moved only for short distances on the foreset slope with an overall tongue-like geometry, as envisaged by Colella et al. (1987), Postma and Cruickshank (1988) and White (1992). It is also noteworthy that even a continuous-looking bed consists of wedge-like subunits (Fig. 2), suggesting emplacement by a series of closely spaced small-volume flows or a surging debris flow. The surging of a debris flow can be caused by several mechanisms. One is the development of roll waves along the upper surface of the debris flow (Davies 1986, 1990) because roll waves can develop easily in a debris flow which moves on a steep slope with high velocity. Another possibility is the development of internal waves along the upper surface of the debris flow due to different fluid velocities or Kelvin-Helmholtz instability. A debris flow can also have a surging behavior if the flow originates from a retrogressive slide or fluctuating bedload discharge at the feeder mouth. The overlapping of wedge-shaped units (Fig. 2C) may form when a deposit of a later-arrived surge stacks up behind the deposit of a preceding one before the suspended sand is settled. When individual roll waves or surges split into small-volume independent flows, single units of Facies A will be produced.

Occasional transition of massive beds into stratified deposits, either in longitudinal or in transverse sections, can be produced by gravity winnowing of material from the frozen debris flow with a positive relief (Postma 1984). Otherwise and more preferably, materials transported along the top and margin of a flow may have been spilled and transformed into thinner flows as the main part of the flow came to a halt (discussed later).

Facies B: Thin- to medium-bedded sandy gravel deposits

Description. This facies refers to sandy gravel deposits that are generally less than a few decimeters thick (Fig. 7A,B). The gravel clasts are mostly fine pebble-size and supported by themselves in a granular sand matrix. Cobble- to boulder-size clasts are scarce or absent. Individual beds show either sheet-form or gradually pinching-out wedge-form geometry. They are commonly inversely graded, characterized by an increase in clasts content and size toward the top of beds (Fig. 7A). This facies commonly occurs in repetition of several units, providing an overall appearance of thin and good stratification. The bed boundaries are generally non-erosional and parallel-sided. At some localities, thick massive beds of Facies A pass downdip into several layers of this facies (Fig. 4A).

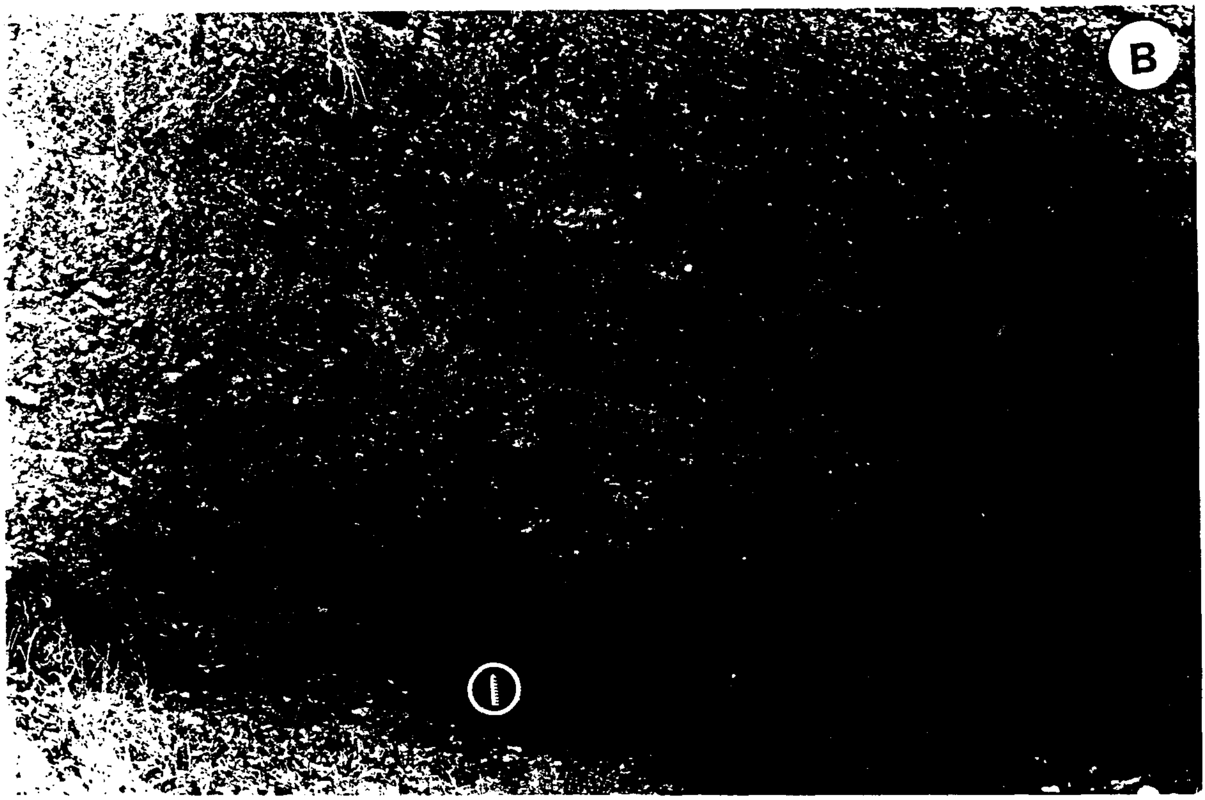
Interpretation. The inverse grading in relatively thin and steeply inclined gravelly sand beds suggests a modified grain flow or a cohesionless debris flow dominated by collisions of pebble clasts in a medium of viscous and buoyancy-providing sandy dispersion (Lowe, 1976). Although similar deposits to this facies were interpreted as traction carpet deposits formed under surging high-density turbulent flows (Massari and Parea 1990) or deposits of thin and small high-density turbulent flows (Postma and Cruickshank 1988), we negate the possibility of involvement with turbulent flows because of the absence of the structures such as normal grading, traction structures or scour surfaces along the bed boundaries. The common lateral transition from a single massive bed of Facies A into several beds of this facies suggests gravity winnowing (Postma 1984) or, more preferably, transformation of a moving debris flow into thinner flows.

Facies C: Gravel sheets

Description. This facies refers to less than several-grain-thick layers of gravel with relatively good lateral continuity over several meters or more (Figs. 3B, 5, 6A,C, 7B). In general, thicker units are laterally more persistent. The gravel clasts are mostly

Fig. 7. A) Several repeated beds of inversely graded pebble gravel deposits (Facies B) in section E. The photograph was taken at the lower left corner of Fig. 4A. A hammer gives the scale. B) Photograph of section F which consists mainly of thin pebbly sandy beds of Facies B and gravel sheets of Facies C as well as a boulder gravel cluster (Facies D). The scale arrow is 20 centimeters long.


A


B

pebble-size with scarce cobble- to boulder-size clasts and are in contact with one another with minimal amounts of interstitial sand matrix. Grading is not evident. The gravel sheets either terminate abruptly or gradually pass into alignments of isolated pebble clasts. In the latter case, size of gravel clasts generally decreases downslope. The lower and upper boundaries are generally parallel-sided but locally irregular when large clasts are present. Clasts are variably oriented, showing either random, flat-lying or imbricated fabric.

Interpretation. The flat-lying or imbricated fabric, paucity of sand matrix, thin sheet-like geometry and gradual downslope transition into alignments of isolated clasts collectively suggest tractional or near-bed movements of gravel clasts either individually or in an assemblage, probably driven by large downslope pull of gravity on a steep slope. Although the clast-supported fabric in the several grain-thick deposits suggests some grain interaction during emplacement, good lateral continuity of most gravel sheets relative to the layer thickness and the lack of snout, lensoidal geometry and inverse grading indicate that individual clasts were transported in a highly dispersed state with minimal grain collision, that is, the momentum of clasts was largely transferred in a streaming mode or by particles themselves (Campbell and Gong 1986; Campbell 1989). These clasts movements dominated by streaming momentum transfer are called "debris fall" (Nemec 1990), compared with a grain flow or a cohesionless debris flow which is dominated by collisional momentum dissipation.

Facies C deposits show whole gradation from grain-assemblage debris fall to single-grain debris fall. Several-grain-thick units (Figs. 3B, 6C) would probably result from a dispersion of saltating gravel clasts, a transitional type between grain flow and debris fall (grain-assemblage debris fall), whereas single-grain-thick units (Fig. 6A) result from rolling or sliding of individual clasts with infrequent clast collisions (single-grain debris fall). The downslope thinning and eventual transition into single-grain-thick type suggest that grain-assemblage debris fall transforms into single-grain debris fall as it spreads and collapses. The paucity of sand matrix seems to be due to rapid downward percolation and stripping of sand into ambient fluid. The dominance of pebble-size clasts with rare cobble-to boulder-size clasts suggests the bypassing of larger clasts further downslope due to their larger momentum and less frictional resistance with the bed (Nemec 1990; Sohn and Chough 1993). The good lateral continuity of each layer comprising highly concentrated gravels suggests that the Facies C deposits are broadly spread on the foreset slope, analogous to the "boulder stream" of Bornhold and Prior (1990) and Prior and Bornhold (1990).

Facies D: Gravel lenses

Description. This facies refers to less than several-grain-thick layers of tightly packed gravel clasts with relatively poor lateral continuity (Fig. 6B). They terminate both up- and downslope abruptly within several meters. Several-grain-thick lenses have convex-up upper surfaces, producing a gently hummocky primary relief above depositional surfaces (Fig. 6B). Single-grain-thick lenses or gravel clusters (Figs. 6C, 7B) also produce a positive relief. Individual clasts are in contact with one another, distinguished from the single grain-thick gravel alignments of Facies C. Constituent clasts are mainly cobble- to boulder-grade in size, considerably coarser than those of the pebbly gravel sheets (Facies C). Interstices between the gravel clasts are filled with sand matrices. The clasts are variably oriented.

Interpretation. The thinness of layers, tightly packed clasts and the lack of matrices are similar characteristics to those of Facies C, suggesting emplacement by debris fall. Although similar deposits to this facies have been interpreted as resulting from modified grain flows by Colella et al. (1987) and Martini (1990), active clast collision is unlikely in a flow whose deposit is commonly single-grain-thick and occasionally comprises only several clasts. We therefore prefer debris fall processes as a more plausible means of emplacement for this facies. The dominance of cobble- to boulder-size clasts can be explained by selective remobilization and downslope movement of large clasts which have larger momentum and smaller frictional resistance with the bed. The large clasts may have been derived either from the fines-removed topset deposits (Martini 1990) or from the protruding clasts of Facies A deposits. The lobate form and large clast size suggest that the Facies D deposits are comparable with the "boulder tongue" of Bornhold and Prior (1990) and Prior and Bornhold (1990) or "elongate boulder cluster" of Kim et al. (1995).

Facies E: Thin-bedded sand with outsized clasts

Description. The beds of this facies are generally less than a few decimeters thick and are either laterally continuous or gradually pinching out. They consist of fine- to medium-grained, well-sorted sand and contain sparse pebble- to boulder-size clasts whose maximum diameters are in excess of the bed thickness (Figs. 2, 4B). They occur as thin veneers on the gravelly beds of Facies A in the proximal part (Fig. 2) or as relatively thick interbeds in the distal part with a thickness up to a few tens of centimeters (Fig. 4B). In the former case, the textural characteristics of the sand can not be distinguished from those of the matrix sand of Facies A. The unit boundaries are commonly disrupted by protruding blocks from below as well as by loading of overlying beds. Individual layers are generally massive, but are occasionally crudely laminated and contain discontinuous fine pebble trains. Lignite laminae are sometimes intercalated. This facies is very rare overall, but is commoner in the distal or downslope part of the foresets.

Interpretation. Generally massive nature rarely with crude lamination suggests suspension settling from turbulent flows with very weak tractive transport. The common occurrence above the beds of Facies A suggests the turbulent flows were produced by surface transformation of a cohesionless debris flow (Hampton 1972; Fisher 1983). The suspended sand may have been settled before long downslope transport, resulting in the thin veneers above the cohesionless debris flow deposit from which the sand was derived. Otherwise, suspended sand may form a separate flow and move farther downslope, resulting in the relatively thick sandy interbeds in the distal part. Sparse outsized clasts may have been entrained from the protruding clasts of the underlying gravelly layer, either flowing or frozen. It is also possible that large clasts were separated from the top of the freezing debris flow due to their large momentum and transported downslope and eventually incorporated into the sandy interbeds (Kim et al. 1995).

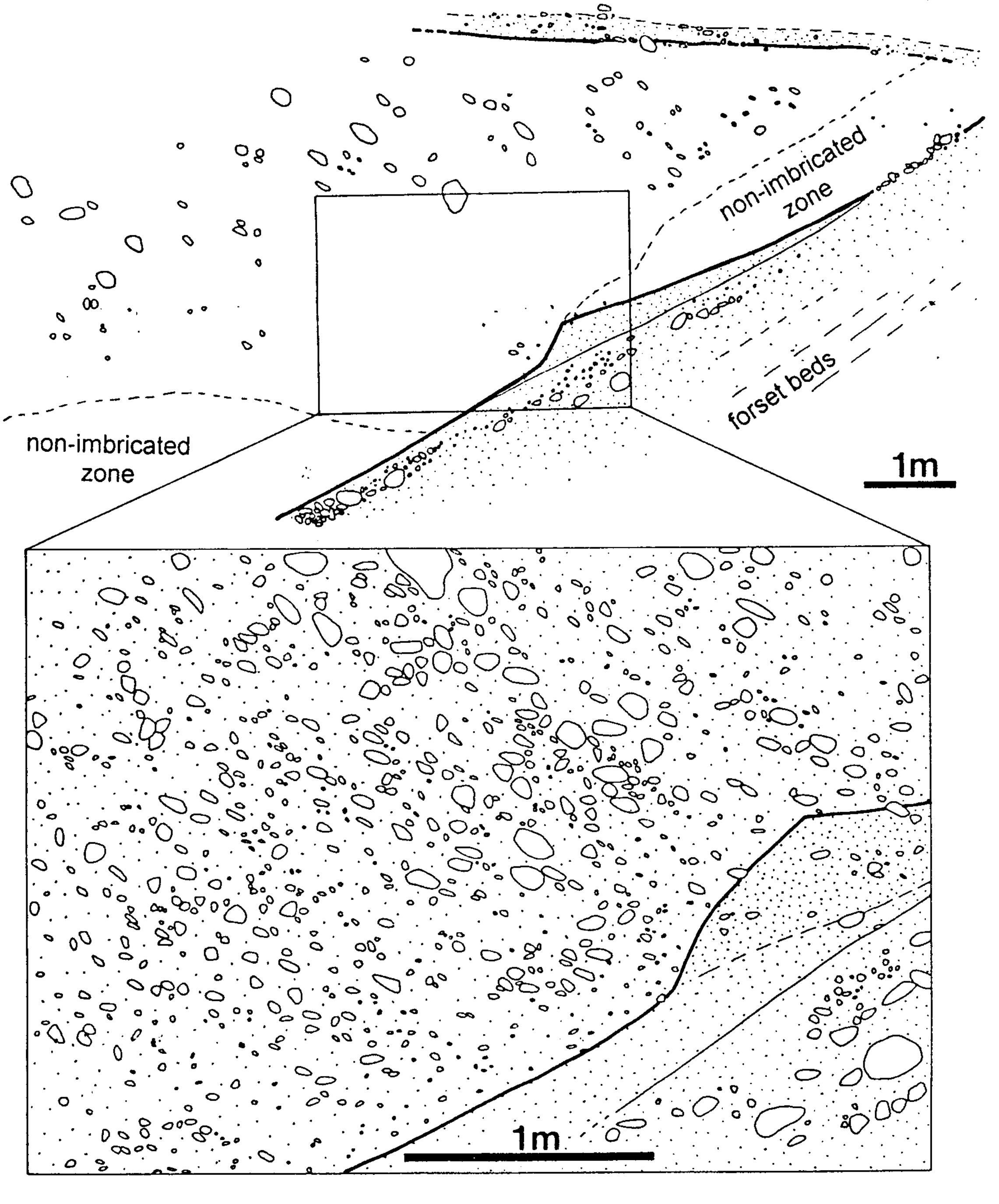
Facies F: Very thick disorganized sandy gravel

deposits

Description. This facies is very scarce in the Doumsan foreset deposits. It includes a very thick, disorganized sandy gravel deposit with very limited lateral continuity both in transverse and in longitudinal sections (Fig. 8). The unit has an irregular lower contact, discordantly overlying the stratified foreset deposits. The upper contact is flat-lying. The gravel deposit shows locally well-developed updip imbrication of gravel clasts, with an angle of 15^0 relative to the horizon and more than 40^0 relative to the lower contact. The imbrication is, however, poorly developed in most part of the bed except for several large clasts. The gravel deposit passes rapidly downslope in to a bed of Facies A (lowermost bed in Fig. 5).

Interpretation. This facies is interpreted as resulting from the collapse and limited downslope movement of the foreset deposits, i.e., a slump. The high-angle imbrication of gravel clasts suggests contractional shear strain of the slump during emplacement (Nemec 1990). Rapid downslope transition of the bed into a Facies A deposit suggests instantaneous remolding and loss of strength of the slump, followed by transformation into a cohesionless debris flow. Scarcity of this facies throughout the Doumsan foreset deposits suggests that large-scale collapse of foreset deposits was infrequent.

Fig. 8. Sketch of a thick massive deposit at section B with a discordant lower contact and showing locally well-developed high-angle imbrication of gravel clasts.


non-imbricated
zone
non-imbricated
zone
forset beds
1m
1m

FACIES RELATIONSHIPS AND FLOW EVOLUTION

The six facies described above have close relationships with one another. Relatively thick, variably graded beds of Facies A commonly pass into stratified deposits or multiple layers of Facies B either in longitudinal or in transverse sections. Beds of Facies A are also commonly overlain by thin sand layers (Facies E), which become thicker and more abundant toward the downslope part of the foreset deposits. Although lateral transition from a bed of Facies A into a gravel sheet of Facies C is not observed in the outcrops, presence of intermediate facies units suggests that there is a complete gradation between the two facies. The example is a many grain-thick and densely packed gravel sheet but with interstitial sand matrix (lowermost gravel sheet in Fig. 3B). There is also an intermediate type between the gravel sheets (Facies C) and lenses (Facies D), which is sheet-like in geometry but with many cobble- to boulder-size clasts (Fig. 6A). Protruding gravel clusters above the beds of Facies A (e.g., arrow 2 in Fig. 3A) have close resemblance to the gravel lenses of Facies D, suggesting that the origin of gravel lenses is intimately related with the protruding clasts of Facies A. Downslope transition form a Facies F bed into a Facies A bed is also observed. All these facies relationships suggest that the six facies are genetically related, i.e., these facies can be explained in terms of an evolving sediment gravity flow. In the following, we provide a model of flow evolution on the foreset slope based on the facies relationships.

In the beginning, the gravelly and sandy sediments near the edge of the transitional zone or the topset-foreset boundary are thought to be resedimented by various triggering mechanisms and move in a cohesionless debris flow of variable size and volume. Some flows may originate directly from the inertia-dominated riverine input when the bedload discharge rate is high enough to generate a hyperpycnal flow. The cohesionless character of the flows is inherited from the nature of the topset sediments, which were deposited in braided streams and shallow marine environments and are generally mud-free (Hwang and Chough 1990). The common lateral pinching-out of most beds suggest that the flows were generally of small volume and moved only for short distances. Therefore, most sediments of the Doumsan foreset deposits are interpreted to have been "recycled" many times, i.e., they were repeatedly remobilized and deposited on the foreset slope. Those deposits that were derived directly from the topset-foreset boundary are thought to be restricted to the uppermost part of the foreset deposits.

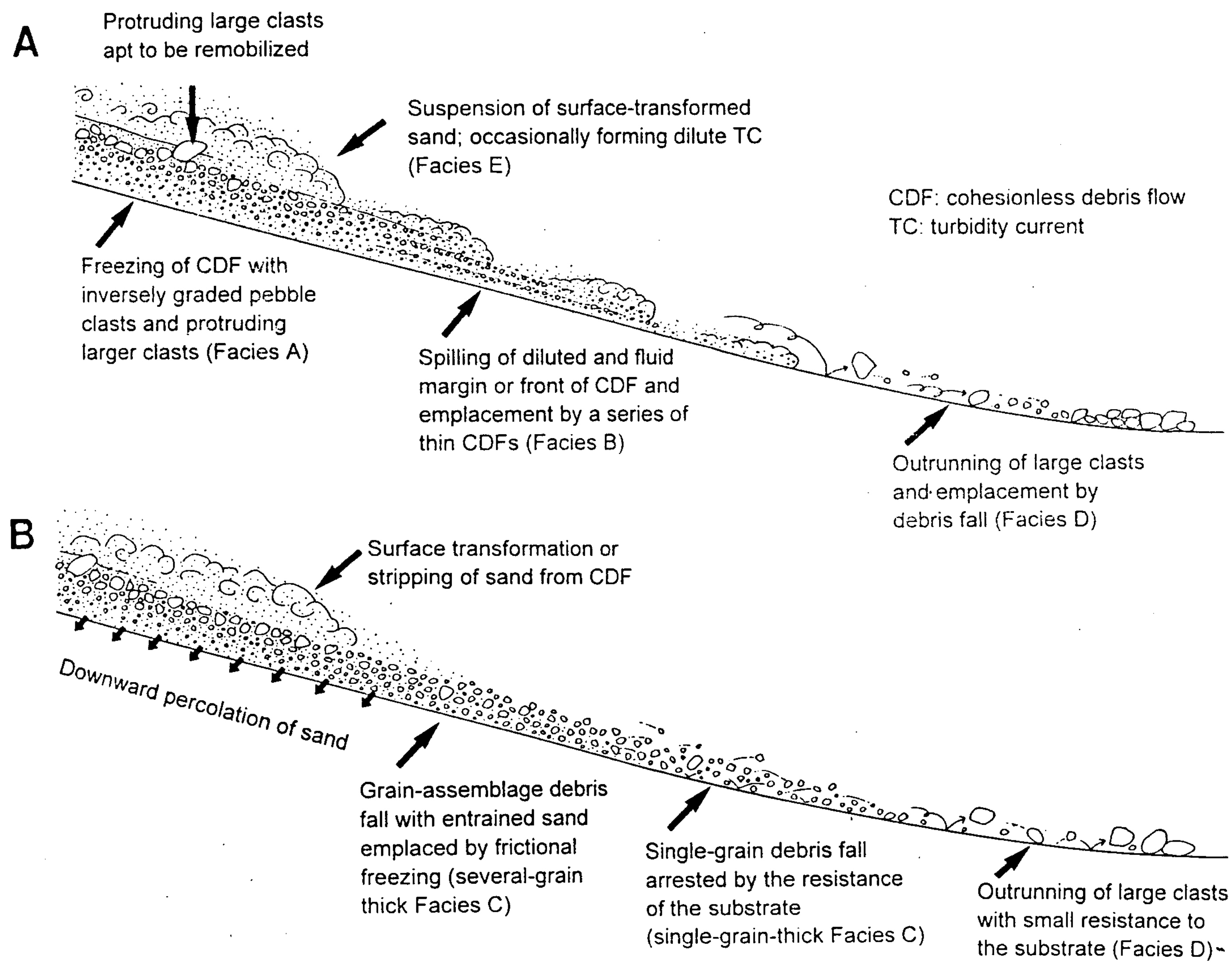
The cohesionless debris flow experiences several types of flow transformations and segregates sediments of different grain sizes during the course of its downslope movement. The gravel clasts are subject to vigorous grain interactions and are inversely graded by either dispersive pressure, kinematic sieving and/or geometrical mechanism. It is important that the pattern of inverse grading varies depending on the size of gravel clasts. The pebble-size clasts, which are the dominant clast phase, are relatively continuously inversely graded from the base toward the top of the pebble-supported division, whereas the cobble- or boulder-size clasts are abruptly inversely graded and float above the pebble-supported division (Figs. 2, 3A). This organization of gravel clasts suggests that the large clasts preferentially drifted upwards and were transported along the top of the pebble-sand mixture. We suggest that the large gravel clasts of Facies D (gravel lenses) were derived mainly from the protruding clasts of moving or deposited debris flows.

In a moving grain flow or debris flow, larger clasts tend to float to the top and move toward the flow front because the upper zone of the flow has a higher velocity. During deposition or freezing of the main body of the flow, the clasts concentrated near the top and front of the flow can be separated from the flow, rather than freeze together, due to their larger momentum and less frictional resistance with the moving flow and the substratum (Nemec 1990; Sohn and Chough 1993). The large grain size contrast between the clast-supported pebble and the protruding clasts (e.g., Fig. 2) probably facilitated continued downslope movement of the large clasts. The separated clasts are no longer a constituent of the flow and move downslope via rolling or sliding until their momentum is lost (Fig. 9A). This inference is consistent with the commoner occurrence of gravel lenses toward the prodelta areas. Another possibility of clast segregation may involve gravity winnowing processes (Postma 1984). After cessation of the flow, the deposits may produce a high positive relief and are subject to large downslope gravity force. The deposits may then collapse in the frontal part, wherein large clasts are concentrated by the conveyer-belt mechanism. The large clasts remobilized in this way can move farther downslope and form gravel lenses or clusters.

When the cohesionless debris flow stops, it produces a bed of Facies A sometimes with stratified deposits of Facies B along the lateral margins and front of the bed. This lateral facies transition is most plausibly explained by the transformation of the debris flow into several thin flows (Fig. 9A). The transformation is probably related with the rollwave phenomena (Davies 1986, 1990). When a debris flow comes to a halt, the faster-moving upper part of the flow would continue to move downslope and spill over or outrun the lower deposited part of the flow. If the spilling of material occurs

Fig. 9. Illustration of flow evolution on the Doumsan foreset slope. A) A cohesionless debris flow produces an inversely graded bed of Facies A. Occasionally several beds of Facies B are produced near the lateral or frontal margins of the Facies A bed by spilling of diluted material over the flow front and margin. Surface transformation of the debris flow generates sandy suspension, from which sand layers of Facies E can be deposited. Large clasts drifted toward the top and front of the debris flow are easily transported further downslope due to their large momentum and smaller resistance with the bed, forming debris fall deposits (Facies D).

B) If a cohesionless debris flow experiences long downslope transport, interstitial sand is removed via surface transformation and percolation. The flow then transforms into a grain-assemblage debris fall and then into a single-grain debris fall, producing gravel sheets of variable characteristics (Facies C). Coarse clasts are selectively transported further downslope and can produce gravel lenses or clusters (Facies D). See the text for more explanation of the processes.


A
Protruding large clasts
apt to be remobilized
Suspension of surface-transformed
sand; occasionally forming dilute TC
(Facies E)
CDF: cohesionless debris flow
TC: turbidity current
Freezing of CDF with
inversely graded pebble
clasts and protruding
larger clasts (Facies A)
Spilling of diluted and fluid
margin or front of CDF and
emplacement by a series of
thin CDFs (Facies B)
Outrunning of large clasts
and·emplacement by
debris fall (Facies D)
B
Surface transformation or
stripping of sand from CDF
Downward percolation of sand
Grain-assemblage debris
fall with entrained sand
emplaced by frictional
freezing (several-grain
thick Facies C)
Single-grain debris fall
arrested by the resistance
of the substrate
(single-grain-thick Facies C)
Outrunning of large clasts
with small resistance to
the substrate (Facies D)

in repetition due to the presence of rollwaves along the top of the flow, several small-volume flows will be generated successively, producing several thin beds of Facies B around a bed of Facies A.

If the cohesionless debris flow continues to move downslope, further segregation of sediments will take place. The mixture of pebble gravel and sand with floating large clasts will be more and more gravel-rich as the interstitial sand is removed from the flow. There are two possible routes of sand removal: one is from the flow base toward the bed and the other is from the flow top toward the ambient fluid. Within the gravel-sand mixture, gravel clasts behave as a granular phase and produce the majority of shear stress. On the other hand, sand behaves more likely as a fluid phase because it generates minimal amounts of shear stress (Nakagawa and Imaizumi 1992). It is therefore thought that the sand is passively entrained by the gravel clasts, providing buoyancy in one way and dampening the impact forces of colliding clasts on the other. Under this condition, part of the interstitial sand will settle or percolate downward (Fig. 9B). Once a sand grain reaches the bed, it will hardly be lifted upward but deposited because there is no fluid turbulence to resuspend the sand grain and the bed is sufficiently rough to arrest the sand grain.

Along the top of the flow, shear stress is induced by the differences in fluid velocities. The interstitial sand in the upper part of the flow will be suspended into the shearing ambient fluid (surface transformation of Fisher 1983; Fig. 9B). The amount of suspended sand may be too small to generate a separate density current. In this case, the sand is not likely to be transported very far from the site of suspension but form a thin veneer of sand above the cohesionless debris flow deposits. When the amount of suspended sand is large, the sand may form a separate flow (e.g., turbidity current) and travel farther downslope or toward the prodelta area independently of the debris flow. Many of the thick Facies E beds in the downslope or distal part of the foreset deposits resulted in this way.

As the debris flow loses its sandy interstitial material along the top and base of the flow, buoyancy and cushioning effect will be diminished. The gravel clasts will collide with one another and with the bed with more impact forces and efficient momentum transfer. The flow therefore becomes more akin to an unmodified grain flow or debris fall, comprising mainly gravel clasts. The flow will eventually decelerate, contract and freeze because the slope angle of the foreset (20^0) is much smaller than the angle of repose and probably than the dynamic angle of internal friction of the flow. The duration of the flow will be, however, very variable depending on the flow characters of individual flows, such as flow volume, velocity, particle concentration and properties (size, shape, etc.). If the flow loses its energy rapidly and freezes, it will produce a relatively thick clast-supported gravel layer with some interstitial sand and with relatively distinct frontal termination (e.g., the lowest gravel sheet of Fig. 3B). The flow can be described as "grain-assemblage debris fall", i.e., a transitional flow type between a grain flow and debris fall (Nemec 1990) (Fig. 9B). Within the grain-assemblage debris fall, both collisional and streaming modes of momentum transfer are thought to prevail.

If the grain-assemblage debris fall continues further downslope, sandy interstitial material will be completely removed and the flow will comprise only gravel clasts which roll, slide or saltate downslope individually, colliding infrequently with one another and with the bed (Fig. 9B). Deposition from the debris fall will occur clast by clast, rather than by en masse freezing, when the momentum of individual clasts is dissipated or the frictional resistance of the substratum exceeds the downslope gravitational force applied to the clasts. Because the deposition occurs clast by clast along the path of its movement, deposits of the debris fall can be as thin as one grain thick. We describe the debris fall as "single-grain debris fall". During this process, cobble- to boulder-size clasts may be selectively transported further downslope because of their larger mobility and be incorporated as a constituent of gravel lenses or as isolated clasts in prodelta deposits.

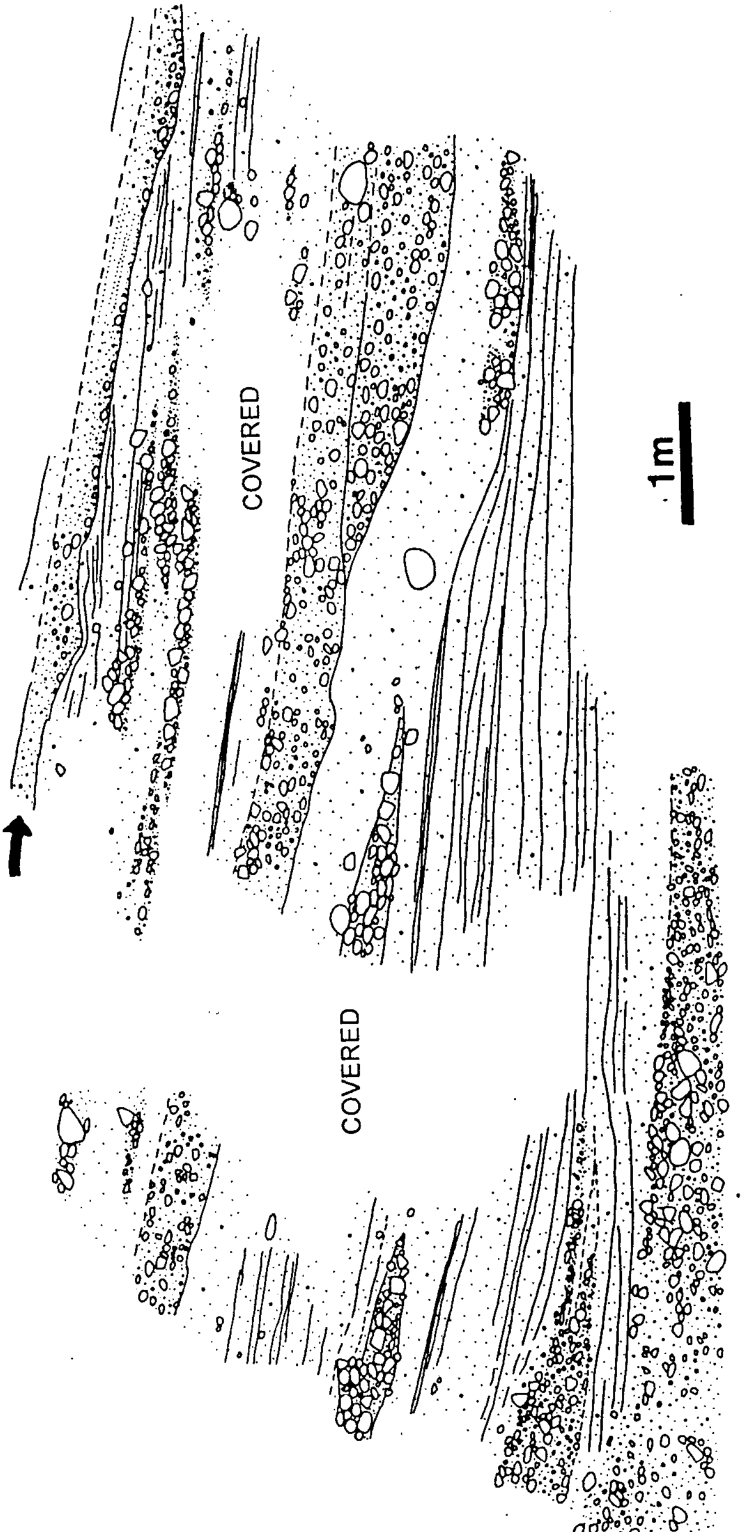
The above discussion suggests that the six sedimentary facies of the Doumsan foreset deposits can be explained by an evolving cohesionless debris flow of gravel and sand materials which transform into debris falls of gravel clasts and dilute flows of sand. This suggestion is in contrast with previous interpretations which regard the facies variations in Gilbert-type foreset deposits as resulting from the variations in feeder system characters such as short-term or seasonal variations in discharge rates and the amount and type of supplied material (Stanley and Surdam 1978; Dunne and Hempton 1984; Martini 1990; Mastalerz 1990). The cause of grain segregation into gravelly and sandy layers was also interpreted as resulting from seasonal alternation of gravel and sand supplies (Martini 1990), in contrast to our interpretation that it is caused by transformation of sediment gravity flows. On small-scale foreset slopes whose heights are of the order of several meters, facies variations due to flow transformations may not be important because there may not be sufficient time and space for a sediment gravity flow to evolve and change its flow characteristics to produce various sedimentary facies. On the other hand, sediment gravity flows have larger chances of flow transformations on large-scale foreset slopes whose heights are in excess of several tens to one hundred meters (e.g., Postma 1984). We suggest therefore that the facies variations in large-scale foreset deposits are controlled more strongly by the characters and spatial evolution of sediment gravity flows.

DEVELOPMENT OF PRODELTA FACIES

The Doumsan fan-delta system is characterized by very broad areas of toeset and prodelta environments that are a few kilometers wide (Fig. 1A). The prodelta sediments were mainly deposited by high- and low-density turbidity currents, cohesive or cohesionless debris flows and slides/slumps (Choe 1990; Hwang 1993). They were deposited in channels, chutes and lobes as well as in interchannel and interlobe areas, forming either lenticular, sheet-like or chaotic deposits. Observations of the prodelta deposits with the foreset processes in mind suggest that many characteristics of them are closely related with the foreset processes.

The prodelta deposits at section J, located about 1.8 km east of section I (Fig. 1), was chosen for detailed study. The deposits consist of thin- to thick-bedded and massive well-sorted sand deposits as well as lenticular gravel deposits of cobble to boulder grade (Fig. 10). The most prominent feature of the deposits is their "textural bimodality", i.e., they lack pebble-grade gravel clasts, comprising mainly sand and cobble-to-boulder gravel. The paucity of pebble gravel can be ascribed to the entrapment of pebble gravel in the foreset deposits. Probably, hyperpycnal flows generated at the topset-foreset boundary could not transport pebble gravel beyond the foreset slope in most cases. Pebble-rich gravity flows generated by the remobilization of foreset deposits were also incapable of

Fig. 10. Sketch of prodelta deposits at section J (Fig. 1A). The deposits comprise thin lensoidal deposits of gravel and thick gravel beds that are either concave-up or convex-up lenticular, as well as thin- to thick-bedded massive sand deposits. The deposits are characterized by "textural bimodality", i.e., the deposits consist of sand and cobble- to boulder-grade gravels, lacking pebble-grade gravel deposits in between. Many of the gravel layers are similar to the gravel lenses of Facies D in the foreset deposits.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transporting pebble clasts far beyond the foreset slope.

On the other hand, sand and cobble-to-boulder gravel could pass through the foreset slope by several mechanisms. Majority of the sand was probably transported by turbidity currents which were generated at the topset-foreset boundary as a continuation of fluvial discharge or a hyperpycnal flow. The turbidity currents probably bypass the foreset slope, leaving no traces upon it, because they are ignitive rather than depositional on steep slopes (Normark and Piper 1991). Additional sand may be added to the prodelta area by the grain segregation processes on the foreset slope. As the sediments on the foresets are repetitively remobilized and deposited by small-volume debris flows, sand is suspended frequently by surface transformation of debris flows (Fig. 9). Some of the sand may form a dilute turbidity current and eventually escape from the foreset slope, being deposited in the prodelta area.

The cobble-to-boulder gravels, which are dropped onto the foreset slope near the topset-foreset boundary, may be transported downslope either included in a cohesionless debris flow or solitarily as debris fall. When transported in a debris flow, they usually migrate toward the top and front of the flow and are easily mobilized farther downslope, either during or after emplacement, as discussed in the former section. Therefore, cobble- to boulder-size clasts can pass through the foreset slope more easily than the pebble-size clasts, resulting in enrichment of large clasts in the prodelta area. The isolated clasts and the lensoidal gravel deposits of cobble and boulder grade (Fig. 10), which are reminiscent of the gravel lenses (Facies D) in the foreset deposits, are most likely a by-product of the foreset processes, rather than produced by a process that operates solely in the prodelta environment. We suggest that the textural bimodality and the presence of gravel deposits that are 'anomalously' coarser than the surrounding deposits are hints of base-of-slope settings or adjacency of high-gradient surfaces. Interpretation of large floating clasts in turbidites as a result of transportation along a rheological interface (Postma et al. 1988) should also be restrained when the depositional setting is inferred to lie adjacent to a steep slope.

The textural bimodality of the prodelta deposits suggests that we can discriminate between a deposit of a flow that bypassed the foreset slope and a deposit whose sediments were segregated by foreset processes. There occurs a peculiar bed at the topmost part of Figure 10 (arrowed), which consists of sediments of whole size fractions, including abundant pebble gravel. The bed is normally graded and stratified, and has a scoured base and good lateral continuity. We think that the bed is 'anomalous' in that the bed has different grain size characteristics and structures, distinguished from the rest of the deposits. Most probably, the bed originated from an unusually powerful flow which comprised whole size fractions because the flow was initiated by direct riverine input. A flow transformed from large-scale collapse of pebble-rich foreset deposits can also produce such a bed.

The large breadth of prodelta area and the overall predominance of mass-flow processes are unique, compared with other prodelta environments (or bottomsets) in front of small-scale Gilbert-type foresets, which comprise fine-grained sediments that were deposited by vertical settling from plane jets or dilute turbidity currents (e.g., Dunne and Hempton 1984; Flores 1990). The uniqueness of the prodelta environment is probably related with the large size of the foreset slope and the overall fan-delta system. Any sediment flows which bypassed the large-scale foreset slope could probably attain a high flow velocity, distributing their load over broad areas far beyond the foresets.

CONCLUSIONS

The Doumsan fan-delta is a marine gravelly Gilbert-type fan-delta characterized by large-scale foreset deposits that are more than 150 m thick. The foreset deposits comprise six sedimentary facies which were deposited mainly by cohesionless debris flows and debris fall. Turbulent flows and large-scale slumps played only a minor role. Facies relationships suggest that these facies are intimately related with one another and can be explained in terms of flow evolution and grain segregation of a sediment gravity flow. Initially, cohesionless sediments were supplied from the topset-foreset boundary and moved in a cohesionless debris flow, which is responsible for Facies A beds. Grains were variously segregated according to their sizes, i.e., into a pebble-supported lower division and an upper sandy division with cobble- to boulder-size clasts. The sediments in the upper division were remobilized into a sandy turbulent flow and cobble-to-bouldery debris fall, resulting in Facies D and E deposits downslope. Occasionally, debris flows transformed into a series of thinner flows, resulting in repeated beds of Facies B around a bed of Facies A. When a cohesionless debris flow continues further downslope, interstitial sand is removed via downward percolation and stripping by ambient fluid. The flow then transforms into grain-assemblage debris fall and single-grain debris fall, depending on flow duration and producing gravel sheets of variable thicknesses (Facies C). These processes of flow evolution and grain segregation resulted in the "textural bimodality" of prodelta deposits, characterized by oversized gravel deposits in sandy background material.

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