ABSTRACT

Water under glaciers and ice streams often flows under pressure over an erodible substrate. Do the channel patterns produced by such flows resemble those of their free-surface counterparts? We studied pressure-driven water flow over an erodible, noncohesive bed using a physical model with a transparent, rigid lid. The dominant channel pattern produced in the model is a widely distributed, braided network of broad, shallow channels. Relative to braided networks formed under equivalent free-surface conditions, those formed under pressure show higher braiding intensity, greater channel curvature and variability of flow direction, and more sharply defined channel margins. Increasing discharge increases braiding intensity, maximum channel size, and variablility of flow direction. Downstream pressure gradients are insensitive to changes in discharge, which may in part reflect a tendency to maintain constant shear stress in the channels, as observed in rivers. Lateral pressure gradients measured in the pressurized model indicate that pressure surfaces are highly variable in both magnitude and direction over time and space. When converted to equivalent topographic slope, these pressure gradients represent much larger lateral slopes than are typically produced in rivers, accounting for the wider range of channel directions in the pressurized-flow experiments.

Keywords: subglacial environment, braided channels, ice streams, flume studies, drainage patterns.

INTRODUCTION

The formation of a network of highly dynamic, unstable, and interconnected channels—a braided network—is the fundamental instability of unconfined free-surface flows over noncohesive sediment (Murray and Paola, 1994). Is this instability connected in some essential way to the presence of a free surface, or could it also occur in a flow confined by a rigid lid and driven by a pressure gradient rather than a slope? This question becomes physically relevant when one considers the likely form of pressurized water flow under a glacier. Here we report a study initially conducted to investigate the types of channels that form beneath glaciers and ice sheets that overlie deformable beds and the physics of flow in these channels.

Ice streams in West Antarctica move at speeds much greater than the surrounding ice because of the highly pressurized hydraulic system that lies beneath them (Engelhardt and Kamb, 1997). The morphology of this drainage system is still subject to debate. Theoretical work by Walder and Fowler (1994) suggests two end member drainage-system types for glaciers and ice sheets resting on an unconsolidated bed. The first consists of an arborescent network of channels cut up into the ice (Röthlisberger channels), characterized by low water pressures \( P_w \) (high effective pressures) and a positive relationship between \( P_w \) and water discharge \( Q \). The alternative to this is a distributed network of wide, shallow “canals” cut into the sediment floor. These channels have high water pressures and an inverse relationship between \( P_w \) and \( Q \). On mountain glaciers, where ice surface slopes are high, both systems can exist but higher \( P_w \) in canals make them unstable in the presence of Röthlisberger channels (Walder and Fowler, 1994). On ice sheets, small surface slopes and low effective pressures favor the formation of a distributed system of canals. A basic tenet of the distributed-network scenario is that pressure-driven flow over a noncohesive bed will braided, analogously to free-surface flow. We investigate this idea experimentally here. The experiments reveal striking similarities and differences between pressure-driven and gravity driven braiding. Although important effects associated with ice cover are missing, the experiments suggest that braided networks could exist beneath ice sheets but may have substantially different form and dynamics than their free-surface counterparts.

EXPERIMENTAL DESIGN AND PROCEDURE

The model consisted of a shallow, horizontal, wooden box (1.73 m long × 1.13 m wide × 0.20 m deep) filled with sediment and covered with a rigid lid that fit inside the box (Fig. 1). The lid was continuously attached to three sides of the box by a metal frame that sandwiched a sheet of rubber between the frame and the lid. This seal provided a watertight boundary while allowing the lid to move in response to water-pressure changes. The lid was not attached to the box at the downstream end, allowing water and sediment to escape across the entire width of the model. A variable-discharge, constant-head tank supplied water to a 1 m long perforated plastic tube placed 0.07 m below the sediment surface at the upstream end of the box (Fig. 1). Well-sorted silica sand with a medium grain size, \( D_{50} \), of 0.25 mm was used in the experiments. Sediment was not recirculated, so sediment supply came from erosion at the upstream end. We did not use data from this area in our analyses. Recirculating sediment would have resulted in slower channel growth, which would not have significantly affected the overall results of this work.

Seven experiments were conducted; four with a rigid, transparent plastic lid (PL experiments) and three with a free surface (FS experiments). We measured \( P_w \) and \( Q \) approximately every hour during each run. Forty-two taps over three transects across the lid provided the pressure data (Fig. 1). We measured \( Q \) at the outlet of the model by timing the accumulation of water in a known volume.

A camera mounted above the model provided photographs of the flow pattern at hourly intervals. Because removal of the lid tended to

---

*Present address: Department of Geophysics, University of Washington, Seattle, Washington, 98195-1650 USA.

© 2001 Geological Society of America. For permission to copy, contact Copyright Clearance Center at www.copyright.com or (978) 750-8400. 
Geology; March 2001; v. 29; no. 3; p. 259–262; 6 figures.
Figure 2. Braid patterns in the transparent plastic lid (PL) experiments. Main flow direction is indicated by arrows. Dye injected into system highlights flow paths. Note faint shadows due to manometer tubing. A: PL 4, time $t = 1.5$ h, $Q = 30 \text{ cm}^3\text{s}^{-1}$, where $Q$ is water discharge. An area of sheet flow exists in the center of the model. Flow converges on headward-cutting channels. B: PL 4, $t = 18.5$ h, $Q = 76 \text{ cm}^3\text{s}^{-1}$. Entire model length is now channelized. C: PL 3, $t = 34$ h, $Q = 190 \text{ cm}^3\text{s}^{-1}$. At high $Q$, a dominant channel has developed on the right side of image.

Figure 3. Braid patterns in equivalent free-surface (FS) experiments. See Figure 2 caption for definition. A: FS 1, $t = 1.1$ h, $Q = 104 \text{ cm}^3\text{s}^{-1}$, where $Q$ is water discharge. B: Experiment FS 1, $t = 3.2$ h, $Q = 226 \text{ cm}^3\text{s}^{-1}$. Darker areas mark location of confluence scours. Arrows show direction of flow.

destroy fine channel features, measurements of channel morphology were obtained from photographs and then calibrated using postexperiment information.

CHANNEL PATTERN DESCRIPTION

The PL channel system is initially characterized by a few straight, headward-migrating channels at the downstream end of the model and a more fine-scale, sinuous set of channel features at the upstream end (Fig. 2A). Over time the model develops by an intrinsically braided, widely distributed network of channels along its entire length. These channels measure up to 20 cm wide and 0.5 to 2 cm deep, have highly sinuous flow paths (Fig. 2B) and, at the fine scale, a strikingly intricate appearance. Channels in the PL experiments regularly reach sinuosity values that would require the addition of cohesive material in free-surface flows (Smith, 1998).

As $Q$ increases, the entire channel system reorganizes and the degree of braiding intensifies. Because overbank flow is limited by the rigid lid, channel banks in the PL experiments are well defined, and new channels are created as erosional offshoots of the main ones. At high $Q$, the PL experiments are characterized by a large main channel that dominates flow (Fig. 2C). Measured velocities within the PL channels ranged from 7.5 cm s$^{-1}$ to 18 cm s$^{-1}$ giving a range of Froude and Reynolds numbers from 0.04 to 0.22 and 1125 to 2700, respectively, for an average channel depth of 1.5 cm. Thus the flows are subcritical and lie somewhere in the transition between laminar and turbulent.

For comparable discharges, channels formed in FS experiments are larger (10–50 cm wide, 1–3 cm deep) than those in the PL experiments and interconnect to form straighter, less complex patterns (Fig. 3A). As $Q$ increases in the FS experiments, water flowing over banks and bars results in an overall “smearing” of the channel pattern compared with the sharply defined channels in the PL runs. Finer features in the channel system become drowned, and channels remain very straight (Fig. 3B). The FS channel system is very stable to changes in $Q$; channel geometry remains relatively unaltered, even with large increases in $Q$.

PRESSURE GRADIENTS

Pressure gradients were calculated across the flow direction ($y$) and along the flow direction ($x$) (Fig. 1). Lateral pressure gradients $dP_y/dy$ were calculated by measuring the pressure difference between pairs of manometers on each transect. A positive value of $dP_y/dy$ implies movement of water from the right to the left looking downstream. Downstream pressure gradients $dP_x/dx$ were measured between manometer taps on separate transects. Both $dP_y/dy$ and $dP_x/dx$ were averaged to give values representative of flux across the width and length of the model, respectively, at any particular time in an experiment.

Downstream Pressure Gradients

Pressure gradients were calculated for periods of steady flow, i.e., when $Q$ did not change by more than 20% over 3 or more hours. However, even during periods of steady flow, $dP_y/dx$ adjusted to accommodate channel growth, which continued to occur some time after $Q$ equilibrated. As a result of this growth, $dP_y/dx$ relaxed in accordance with standard theories of pipe flow (Catania, 1998). Interestingly, $dP_x/dx$ consistently relaxed to values of $\sim$200 Pa m$^{-1}$ for the PL experiments (Fig. 4). In part, the constant pressure gradient was forced by the fact that the overall pressure gradient applied to the system was constant because of the constant-head tank. However, the pressure losses in the inlet plumbing were significant and should have varied with $Q$. This suggests that at least part of the constant $dP_x/dx$ is due to an internal shear-stress regulation mechanism in the channels akin to that found in natural rivers. Bedload-dominated alluvial rivers with non-cohesive beds have a tendency to maintain constant values of dimensionless boundary shear stress $\tau_*$ (Parker, 1978; Paola et al., 1992; Parker et al., 1998), defined as

$$\tau_* = \frac{\tau_0}{g
m(\phi - 1)D_50},$$

where $\tau_0$ is the shear stress, $g$ is the acceleration due to gravity, $\phi$ is the porosity, $m$ is the slope, and $D_50$ is the median particle diameter.
where \( s \) is the sediment specific gravity, \( \rho_w \) is the density of water, and

\[
\tau_0 = \rho_w g z S,
\]

(2)

where \( S \) is the slope of the channel bed and \( z \) is channel depth. For the pressurized case, \( \tau_0 \) is instead given by

\[
\tau_0 = \rho_w g z \left[ \frac{1}{\frac{dP_w}{dx}} \right].
\]

(3)

Thus, for \( \tau_0 \) to remain constant at constant flow depth, \( dP_w/dx \) must remain constant. The tendency for \( dP_w/dx \) to stabilize at a constant value may simply reflect a constant-\( \tau_0 \) condition in pressurized channels analogous to that in free-surface flows.

**Lateral Pressure Gradients**

Unlike \( dP_w/dx \), lateral pressure gradients \( dP_w/dy \) are not constant during periods of steady \( Q \), rather they vary in magnitude by as much as 500 Pa m\(^{-1}\) for any particular \( Q \). This results in pressure surfaces that vary both spatially and temporally during the PL experiments.

Water pressure in the system is defined as

\[
P_w = \rho_w g h_T,
\]

(4)

where \( h_T \), total head, is measured as

\[
h_T = h_0 + h_p + h_u,
\]

(5)

where \( h_0 \), \( h_p \), and \( h_u \) are the elevation head, potential head, and velocity head, respectively. For both PL and FS experiments, \( h_0 \) is the same value for both the PL and FS experiments. Changes in \( h_T \) in the PL experiments therefore reflect only changes in \( h_0 \) in the system, and we refer to potential head simply as pressure head. Similarly, changes in \( h_T \) in the FS experiments reflect only changes in \( h_0 \). Because \( dh_T/dx \) in the PL experiments stabilizes at a constant value of 0.025 (as described previously); slope \( S \) for the FS experiments was set at 0.02.

The difference in meaning of \( h_T \) for PL and FS experiments is critical. Whereas flow in the PL experiments is steered primarily by differences in pressure, flow in the FS experiments is mainly directed by differences in elevation, i.e., by topographic slope. Thus it is much easier to change the pressure field within a pressurized fluid than to change the water-surface gradients that drive free-surface flow, since the latter are strongly coupled to bed topography. Changing bed topography requires import or export of sediment, a process that occurs relatively slowly. Changes in the pressure field are internal to the fluid and so are effectively instantaneous. There is also no physical upper limit to the magnitude of local pressure gradients, but sediment slope is limited by internal friction. Thus it is much easier to produce strong local pressure gradients than to produce equivalent gradients in water-surface topography. This difference, together with the impossibility of channel overspill in the rigid-lid model, accounts for the differences in channel patterns between the pressurized and free-surface experiments.

**MEASURES OF CHANNEL PATTERN COMPLEXITY**

**Braiding Intensity**

Braiding intensity is a measure of the average number of active channels per model transect (Howard et al., 1970) and represents the degree of braiding of the channel system. We measured braiding intensity along six transects when the system had become fully channelized. The influence of stream power \((Q \delta)\) on braiding intensity has been shown through theoretical and statistical analysis by several authors (Howard et al., 1970; Parker, 1976; Ashmore, 1991). For comparison between PL channels and FS channels, stream power, \( \Omega \), is defined as

\[
\Omega = Q \frac{dh_T}{dx}.
\]

(6)

Braiding intensities in the PL experiments range from 4 to 6 active channels per transect, more than twice those in the FS experiments, under comparable conditions. As \( Q \) increases, braiding intensity for the PL experiments increases 1.5 times faster than for the FS experiments. This inherent difference in the degree of braiding between the FS and PL experiments is a direct result of the variable pressure-head surfaces in the PL experiments as discussed above.

**Flow Direction**

Measurements of variability of flow direction allow us to compare the relative importance of lateral flow in the PL and FS experiments. Flow directions were analyzed using the method of Curmay (1956), weighting the measured angles by channel depth. When \( Q \) was increased in the PL experiments over a short time period, there was a noticeable increase in the relative frequency of lateral flow (manifested in higher vector standard deviation \( \sigma \); Fig. 5). Sediment eroded from channel banks choked smaller channels, diverting flow to alternative paths and reconfiguring the channel system.

Directional variability in the FS experiments was much less sensitive to discharge (Fig. 6). A \( Q \) increase much larger than the previous example produced an increase in \( \sigma \) of only 3°. The stability of the FS system is due to the slowly varying nature of bed topography, even when \( Q \) is increased. Conversely, the high degree of variability in \( h_T \) in the PL experiments sets up local pressure gradients that force flow in directions that often diverge strongly from that of the mean pressure gradient.
CONCLUSIONS

Pressurized flow over a noncohesive bed produces a highly braided channel system that is largely controlled by variation in lateral pressure gradients. The magnitude of these lateral pressure gradients is quite high, and much more variable over time and space than the downstream pressure gradient. This effect creates highly variable pressure surfaces that in turn create the intricate, braided network observed in the pressurized system.

Braiding intensity in pressurized flow is much higher than in the free-surface case because the rigid lid hinders flow over banks and bars. As $Q$ increases in the pressurized system, erosion must expand existing channels or carve out new ones. The ephemeral, but strong, lateral pressure gradients supported by the rigid upper surface drive flow laterally at significantly higher angles than in equivalent free-surface flows. In the free-surface case, the flow has the added degree of freedom of being able to flow easily over topographic barriers but loses the possibility of long-distance lateral flow associated with strong lateral pressure gradients. This gives free-surface braiding a less intricate, directionally variable network with fewer channels.

The PL experiments give us a glimpse of the possible morphology of subglacial channels. The types of channels that exist beneath the ice streams of West Antarctica are largely a result of low effective pressures and a deformable substrate, conditions that are present in these experiments. We do not know how our results would have been different if the rigid lid in our PL experiments had been deformable, as ice is. The experiments reported here show that purely pressure-driven braiding is possible, as glaciologists have speculated. And despite a strong family resemblance to its familiar free-surface counterpart, “braiding under glass” clearly has its own personality.

ACKNOWLEDGMENTS

We thank Roger LeB. Hooke, Gary Parker, and Joe Walder for their comments and contributions to this work. This research was conducted in the St. Anthony Falls Hydraulic Laboratory at the University of Minnesota and was supported by National Science Foundation grant EAR-96-28593.

REFERENCES CITED


Manuscript received June 16, 2000
Revised manuscript received November 29, 2000
Manuscript accepted November 30, 2000

Printed in USA