Dudley Chelton video
Satellite altimetry
2 satellites (upper)

1 satellite
Day 3

1. Rotating, stratified fluids: oceans and atmospheres
   - Vorticity: a vector-tracer in classical homogeneous fluids
2. Wave dynamics: fundamentals, group velocity, energetics, ray theory
3. Potential vorticity (PV)
   - Vortex stretching, Prandtl’s ratio, geography of PV
4. Rossby waves
5. Case study of topographic effect on atmospheric circulation: Greenland and Atlantic storm track.
6. Teaching young undergraduates about the global environment?
7. Seminar: subpolar climate dynamics observed from above and below: altimetry and Seagliders
Fig. 4. One-point correlation plot for CCM3 mean Dec–Feb (DJF) 300-mb nondivergent $u$ wind component internal variability for a base point at (28.9°N, 112.5°E). Contour interval is 0.1.
zonal wind (m/sec) at 45W with Greenland topography

JFM 1993

“Svairdrups”

\[ 1 \text{ Sv.} = 10^6 \text{ tonnes sec}^{-1} \]

Kuroshio \( \sim 50-100 \) Sv.

A.C.C. \( \sim 180 \) Sv.
Topographic flows: we now introduce the effects of mountainous topography, which owing to the ‘vertical stiffness’ imparted by rotation, is greatly influential. On an f-plane, the Rossby number, U/\( f \), topographic height and Froude number Nh/U are key parameters.

(i) 2D stratified flow over bump of height \( h \) (buoyancy freq N upstream flow U)
\[ \text{Nh}/U \]

(ii) 3D stratified flow …over or around
\[ \text{Nh}/U \text{ crit} \sim 2 \quad \text{(Schaer+Durran, JAS 1997)} \]

(iii) f-plane rotation, unstratified: Taylor column dynamics
flows ‘around’ if \( h/H > \text{Ro} \) (Ro = U/\( f \)).

(iv) f-plane rotation weakens blocking (Nh/U crit =>3) Taylor column => Taylor cone Nh/U > 1 again! (independent of Ro)

(v) \( \beta \)-plane flows
- large-scale potential vorticity (PV) gradient
- contours of constant background PV \( (f/h = \text{const}) \)
  - bend Equatorward over a ridge, locating a \textit{cyclone} over its crest for very slow flow
- dispersive lee waves, semi-circular wavecrests
- hydraulic structure in some limits (planetary geostrophic) where the waves are non-dispersive
- strong upstream blocking (‘Lighthill mode’) for \( \beta L^2/U > 1 \)
pressure drag occurs when the anticyclone shifts upstream as wake vorticity is shed, placing the Equatorward flow over the lee downslope

Impulse (the time-integrated force on the solid Earth) is summed up by meridional PV flux: 
\[ \langle q'v' \rangle \] which expresses the x-averaged force on the Eulerian fluid along a latitude circle

\[ \sim (f/H) \langle h'v' \rangle \]

\[ Nh/U = 1, 2, 2.5, 3, 4, 6 \]

Petersen, Olaffson, Kristjannson JAS2003
Lee Rossby-waves in the wake of a cylindrical mountain  
(*McCartney JFM 1976*)

Rossby waves are ‘one-way’: their phase propagation has a westward component relative to the fluid: thus they exist as lee waves for an eastward flow but not a westward flow. Wave drag peaks at: 

\[ 8.2 \frac{\delta}{\varepsilon} \] 
times the ‘naive estimate’, 

\[ \frac{\rho f U L^2 \delta H}{\varepsilon} \]

where \(\delta = h/H\) is the fractional mountain height, \(\varepsilon\) the Rossby number, \(U\) the mean flow, \(L\) the radius and \(H\) the total fluid depth.

Note strong correlation of meridional velocity and topographic height.....wave drag. Also note the beginnings of jet formation southeast of the mountain. even in the linear theory.
wavenumber diagram for Rossby waves in steady zonal westerly flow ($U>0$)

\[
\omega - U k = - \frac{\beta k}{k^2 + l^2}
\]

\[
\psi = R \exp\left(ikx + ily - i\omega t\right)
\]

\[
\omega = 0 \quad \text{STATIONARY WAVES}
\]

\[
k \left( U = \frac{\beta}{k^2 + l^2} \right)
\]

\[
\begin{cases}
  k^2 + l^2 = \beta / U \\
  k = 0
\end{cases}
\]

$\frac{k}{U} \text{ Rossby wave number}$

Extend downwind faster than $U$... at 2xU
A westerly (prograde, cyclonic) zonal flow encounters a small mountain (at 2 o’clock). Rossby wave dynamics produces standing waves downwind, a convoluted lee cyclone, intense jet structure wrapping round the mountain, and a ‘Lighthill block’ upstream. This stagnant blocking region is (in linear, yet finite topography, theory, a Rossby wave with vanishing intrinsic frequency and upstream group velocity for merid. wavenumbers \(< \left(\frac{\beta}{U}\right)^{1/2}\).

Note the ruddy pressure features which are fine-scale evaporative convection cells, pillar-like cyclones. The edge of the block is outlined by convective rolls.

Here the controlling parameter \(\beta a^2/U > 1\) meaning that the wake is stable; smaller values of this parameter yield unstable wake and transient Rossby waves which ironically fill the hemisphere (they are not simple lee waves). See Polvani, Esler, Plumb JAS 1999 for a numerical study with some of these features.
Streak image of the same experiment (dots 2 sec apart) showing intensity of jets near mountain, lee Rossby waves and upstream block.
Greenland's effect on the Northern Hemisphere circulation: downslope winds meet the Atlantic storm track.

P.B. Rhines
Oceanography & Atmospheric Sciences
UW

Image: Petermann Glacier, NW Greenland
Kendall Sjellin
Let’s remove the ice (only temporarily)

Konrad Steffen, Univ. of Colo
R/V Knorr in Labrador Sea. At the time of this research cruise, the first deep ice cores were being drilled on the summit of Greenland. The iceberg likely calved off the Jakobshavn glacier in west Greenland. The strata, faintly visible, record climates back 120,000 years. Air bubbles in the ice accurately give us a whiff of ancient climates, showing the high correlation between Earth's temperature and the amount of carbon dioxide and methane in the air.
Red/blue = high/slow SLP

Black contours
Z250

Yellow contours
Z50
Greenland is near the ‘center of action’ of the Icelandic low, with extreme activity of jet stream, tropopause folds, storm track, meridional moisture- and heat-flux, stratospheric polar vortex, oceanic global overturning circulation and implicitly the NAM/AO/NAO principal EOF of hemispheric sea-level pressure.
The next figures are animations of the northern hemisphere circulation for a high-NAO (1989) and low NAO (1996) winter, respectively. They show many things, including strong vertical interaction between clusters of cyclones and the stratospheric polar vortex. Synoptic activity is felt in the stratosphere! A strong sudden warming is seen in 1989.

The two winters are very different. With low NAO (1996), it seems the storm track is farther south, and does not excite the stratospheric polar vortex overhead as strongly as with high NAO (1989).
1989 JFMA: colors: 1000 HPa near-surface dynamic height (blue=low pressure, red=high) contours: jet stream level 300 HPa, stratosphere 30 HPa

NAO index positive

Synoptic storm tracks are beneath stratospheric polar vortex

Greenland
1996 JFMA colors: 1000 HPa
(NAO index red contours: 300 HPa
very negative) black contours: 30 HPa

Storm tracks are usually south of stratospheric polar vortex.

Greenland
In moderate resolution simulations, Greenland's topography (but not albedo) actually weakens the Icelandic low interference pattern with hemispheric standing waves ... sea-level pressure is lower to the west and higher to the east of Greenland ... model dependent?

**SLP**

**500 hPa**

**500-1000 hPa thickness**

*Fig. 4.* (a) The mean winter (DJF) sea level pressure (hPa) in the NOGREEN simulation and (b) the mean sea level pressure difference (hPa), CONTROL–NOGREEN. The contour interval is 5 hPa in (a) and 50 hPa in (b). All the major differences in (b) are 95% statistically significant.

*Fig. 5.* (a) The mean winter (DJF) 500-hPa geopotential height (gpm) in the NOGREEN simulation and (b) the mean 500-hPa geopotential height difference (gpm), CONTROL–NOGREEN. The contour interval is 100 gpm in (a) and 20 gpm in (b). All the major differences in (b) are 95% statistically significant.

*Fig. 6.* The mean winter (DJF) 500-1000 hPa thickness (gpm) difference, CONTROL–NOGREEN. The contour interval is 20 gpm and all the major differences are 95% statistically significant.

Petersen, Kristjansson & Jonsson Tellus 04
• Numerical simulations (Petersen, Kristjansson & Olafsson Tellus 04 T106; Kristjansson & McInness QJRMS 99) suggest that Greenland topography reduces the strength of cyclonic systems in the Atlantic storm track, blocking cold-air outbreaks from Arctic Canada: Interference pattern with hemispheric standing waves.

• Is this model dependent? Paradoxically, tip-jets and gap-jets (downslope winds) are frequently observed there and are very intense. Winds in the Labrador and Greenland Seas can be very strong.

We are not contemplating removing Greenland until it melts.
potential vorticity and $\Theta$

Figure 9. Vertical cross-sections along a SW-NE line through the Icelandic low on 12 April 1983. The first section (a) shows the results obtained by simple finite differences applied to the routine ECMWF analysis at 12Z on that day. The thin lines indicate isentropes at 10 K intervals and the thick lines PV at 1 unit intervals. The 0.75 PV unit contour is shown by a heavy dashed line. The second section (b), from M. A. Shapiro (personal communication), is the result of detailed measurements using aircraft and dropwindsondes. Isentropes every 2 K are indicated by light solid lines, isotachs every 10 m s$^{-1}$ by dashed lines and the estimated tropopause position by a heavy continuous line. The scales and orientation of the two sections are approximately the same.

As far as gross features are concerned, everything is the opposite way round from Figs. 8-10. The tropopause is high, the isentropes in the troposphere bow downwards, and those in the stratosphere bow upwards. The very low potential vorticity just under the tropopause is particularly striking. Once again we shall see that...
• despite this no-Greenland model study, winds are extraordinarily there; pressure drag on Greenland’s slopes can propagate upward, block westerly flow, and affect the subpolar (and thus global-) ocean beneath
from Dudley Chelton: Greenland tip jet
pressure drag and the length of the day

Hide et al. JGR 1997

Figure 1. Time series of irregular fluctuations in the length of the day (LOD) from 1963 to 1992 (curve a) and its decadal, interannual, seasonal, and intraseasonal componentis (curves b, c, d, and e, respectively). The decadal (curve b) component largely reflects angular momentum exchange between the solid Earth and the underlying liquid metallic outer core produced by torques acting at the core-mantle boundary. The other components (curves c, d, and e) largely reflect angular momentum exchange between the atmosphere and the solid Earth, produced by torques (proportional to the time derivative of the LOD time series) acting directly on the solid Earth over continental regions of the Earth's surface and indirectly over oceanic regions (adapted from Hide and Dickey [1991]).
\[-\frac{\partial}{\partial t} \left[ \int_0^P m_r \frac{dp}{g} \right] = \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \int_0^P m_r \frac{v \, dp}{g} \cos \phi \right) + \frac{\partial h}{\partial \lambda} \]

\[+ a \cos \phi[T_{ss}^e] + a \cos \phi[T_{bl}^e] - f a \cos \phi \left[ \int_0^P \frac{v \, dp}{g} \right].\]

\[\text{resolved torque \sim} 5 \times 10^{18} \text{ N m}\]

Brown, QJRMS 2004
6 month ECMWF model integrations at T95-T159 L60 and T255 L40, and case studies
Pressure drag on Greenland (using T511L60 ECMWF model)

'normal' drag events (Greenland pushes atmosphere westward, low pressure to the east)
1000 hPa fields correlated with high pressure drag events at lags -6 days to +4 days. ERA-40 reanalysis data (T159L60) 1982-2001
250 hPa
50 hPa hemispheric response with precursor: the SPV is pulled toward Baffin Bay during high normal pressure drag events
1000 hPa fields correlated with high normal pressure drag across Greenland: high-pass filtered eddies (2-10
Z1000 associated with ‘abnormal’ pressure drag (high pressure on eastern slopes of Greenland)
effect of model resolution on resolved pressure drag, surface friction and parameterized gravity wave stress
Storm tracks
splitting of storm track, some lows moving northward on west side of Greenland (short lived), most moving north on eastern side. Cyclones are also split vertically by the 'knife edge'.
occurrence of cyclones in ERA-40 data, and at various model resolutions

(a) ERA-40

(b) T95L60

(c) T159L60

(d) T255L40

200 km

120 km grid

80 km
east-side storm tracks launched Irminger Sea (left) and Greenland Sea (right)
Two case studies, winter 2004/5 (strong, extraordinarily cold, elongated stratospheric polar vortex arrived early in the season)

Limpusaven et al. JGR 2007 event

'normal' drag events (Greenland pushes atmosphere westward, low pressure to the east)
Θ on PV-2 surface (~tropopause) and SLP: Christmas 2004T511L60 ....3 storms in 10 days

Cold air sweeps through the northern tip of Greenland, even at the tropopause level

25 Dec 2004

$2 \times 10^6 \text{m}^2 \text{Ks}^{-1} \text{kg}^{-1}$
25-27 Dec 2004

surface winds

lowlevel vorticity
Lagrangian back-trajectories (using software of H. Wernli)
Effect on subpolar gyre of the Atlantic, and Greenland Sea: enhance air/sea heat flux: much intensified at higher model resolution.

T95, T255, T799, T799, T799, T799
Principal eof of sea surface elevation, 1992-2006, which is mostly a simple trend, showing *deceleration* of the subpolar Atlantic gyre over 15 years

Häkkinen & Rhines 2004 Science
Tip jets and reverse tip jets:


10%. If one considers a higher threshold, 33 m s⁻¹ or hurricane-force winds, then the probability of observation near Cape Farewell drops to approximately 6%, while in the two regions along Greenland's southeast coast it drops to approximately 3%-4%. Aside from the Greenland coast, the other area where one is likely to experience high winds is in the central North Atlantic (around 57°N, 30°W). This is along the southern flank of the primary synoptic-scale storm track (Hopkins and Hodgetts 2003). The region to the northeast of Iceland as well as the northwest coast of Greenland and adjoining areas of the Labrador Sea stand out as regions where the probability of observing high winds is low.

We finish our presentation of the statistics of the 10-m wind field with a more detailed view of the winds at the three locations identified above as being ones where high wind speed events are common: Cape Farewell, Denmark Strait South, and Denmark Strait North. Wind roses derived from the QuikSCAT data for these three locations are shown in Fig. 7. With regard to Cape Farewell, the wind regime is bimodal with winds coming more frequently from the north and west, with the direction of the strongest wind speed.

Figure 2: A tip jet event recorded by Greenlandic meteorological land stations. The COAMPS model output (averaged over the region 59°-60°N, 37°-42°W) is compared with the observed meteorological time series (see key) for February 1997. The land station data are recorded every 3 hours (gaps indicate missing data); see Fig. 1 for the locations of the meteorological stations. The tip jet event is denoted by T1. a, Sea level pressure; b, zonal 10-m wind (positive is westerly). The COAMPS and station P winds are significantly correlated (r = 0.72); neither of them is correlated with the station I wind record. c, 2-m air temperature. The two meteorological station temperature records have been offset by 4°F (warmer) for plotting purposes.
Forecast effect of high-pressure drag events is short-lived, particularly in Europe; skill of forecasting drag is high.
EP-flux: hemispheric impact; anomalous SPV acceleration (barotropization rather than simple mountain drag)

note factor of 10 zoom for anomaly vectors
Egger *JAS* 2006 finds that zonal momentum tendency can be opposite to expected push by pressure drag...due to imported meridional vorticity flux in transient eddies. Covariance fields with pressure drag, barotropic model.
Gravity waves
Fig. 6. Cross section of potential temperature (K) at 0000 UTC 10 November 2001, along the line AA’ of Fig. 3.

Three-dimensional depiction of the Greenland relief.
ECMWF model runs: high-pressure drag events develop downslope winds and upward propagating gravity waves reaching the stratosphere at T511, T799....resolutions higher than that used in ‘no-Greenland/Greenland’ model studies.
Downslope winds increase wavedrag (by Bernoulli) here in a layer of CO2
AIRS Radiance Perturbation (< 500 km)

Limpasuvan et al. JGR 2007
gravity waves decelerating SPV: at rates 10-120 m sec\(^{-1}\) day\(^{-1}\)
24 Jan 2005..an ‘abnormal’ pressure drag event (would lead to westerly accel of atmos)

Figure 4. (Top row) ARPS simulations at 0600 UTC and 1800 UTC of 24 January 2005 at 200 hPa. The height field (Z, in hectometer) and temperature field (T, in Kelvin) are given as black and green contours, respectively. The vertical wind (w) is given as filled color contours: upward (downward) motion is shown in red (blue). Contour intervals are indicated. (Bottom row) As above except at 500 hPa. Greenland and Iceland are shaded in gray.

Limpasuvan et al. JGR 2007
Mel Shapiro’s Greenland flights

Fig. 5. Cross section of potential temperature (K) at ~1200 UTC 29 January 1997 derived from dropsondes (numbered 12-17) from the NOAA/G-4 aircraft.
Ro = □
Nh/U = 1.5

Ro = 0.42
Nh/U = 1.5

Petersen, Olaffson, Kristjannson
JAS2003
Schär (JAS 93): PV is transported along the intersections of the Bernoulli-function and isentropic surfaces in a statistically steady flow.

\[ PV \text{ flux}: \mathbf{J} = \nabla \theta \times \nabla B \]

\[ B = \text{enthalpy} + \frac{1}{2} |\mathbf{u}|^2 + \Phi \]

\[ \approx c_p T + \frac{1}{2} |\mathbf{u}|^2 + gz \]

Schär & Durran
JAS 97

also: tip horiz vorticity to make vertical vorticity
Rottuno et al. JAS 99
Ertel potential vorticity generation by breaking lee gravity waves. The PV generation as well as the gravity-wave momentum flux alter the geostrophic circulation.

\[
\frac{\partial q}{\partial t} + \nabla \cdot \mathbf{J} = 0,
\]

\[
\mathbf{J} = \nabla \theta \times \nabla B,
\]

Chen, Hakim & Durran, JAS 2007 subm
PV and zonal flow generation in flow over a 1.5 km high mountain

dipole of PV, decelerated wake

Chen, Hakim & Durran, JAS 2007 in press
Overturning circulations
Some ‘burning’ questions for which we thought we knew the answers:

(i) What drives the global meridional overturning circulation (MOC) of the oceans — buoyancy or mechanical mixing induced by winds and tides?

(ii) Is high-latitude sinking and the deep, cold branch of the MOC a dominant member of the meridional heat and fresh-water transport?

(iii) Does the ocean circulation substantially warm western Europe? More generally, does heat transport by oceanic general circulation affect atmospheric climate?

(iv) What are the paths of upwelling of deep waters in the global oceanic MOC?

(v) Where are the crucial sites for convection and water-mass transformation?

(vi) What is the quantitative rate of water-mass production for the several components of the North Atlantic Deep Water (for example, Labrador Sea Water), and how are they altered before being ‘delivered’ to the global MOC?

(vii) How do convection and mixing drive diffusive overturning at many scales, reaching to the distant circulation.
Oceanic overturning circulations: coexisting with ‘horizontal gyres of wind-forced circulation
• MOCs have an easier time in the oceans than in the atmosphere:

A ring of air moved 1000 km north gains westerly velocity of 100 m sec⁻¹. There is not enough energy available to utilize this mode; the Hadley cell is limited in north-south extent. Forces (eddy momentum flux from PV stirring) and non-symmetric circulation are required to support extensive meridional excursion.
channels and conduits for heat- and fresh-water transport
After Schmitz (1995)
consider the differences between tropics and Arctic... (a) at 60N latitude the sunshine incident per unit area is 50% of the full intensity with the sun overhead; (b) the albedo (whiteness) is greater.
solar radiation (kilowatt-hours per square meter, per day) varies with latitude and season (here neglecting the great effect of cloudiness)
• A simple radiative calculation gives an Earth with the correct average T, but wrongly distributed meridionally (north-south)
the simple radiation back directly to space has average temp of about 250K, 290K = average surface 255K (-18°C)
Global meridional heat transport divides roughly equally into 3 modes:

1. atmosphere (dry static energy) \( c_p T + \Phi \) (Bryden & Imawaki 2002)
2. ocean (sensible heat)
3. joint atmosphere/ocean mode: water vapor/latent heat transport \( Lq \)

The three modes of poleward transport are comparable in amplitude, and distinct in character (sensible heat flux divergence focused in tropics, latent heat flux divergence focus in the subtropics) (based on Keith (Tellus 1995) climatology, similar to more modern: Trenberth et al. J.Clim 2003)

Error est.: \( \pm 9\% \) at mid-latitude; Bryden est 2.0 \( \pm 0.42 \) pW at 24N

The northern subtropics show extremely active upward air/
very similar numbers from Trenberth & Stepaniak, QJRMS 04
Flux of fresh water by the atmosphere is concentrated in the Pacific and Atlantic storm tracks.

Globally it carries ~2 petawatts of latent heat flux ... which is ~0.7 Sverdrup (0.7 megatonnes/sec) of freshwater flux.

1993 JFM
1996 JFM
Trenberth & Caron JClim01

Meridional heat transport at 35N: 78% A, 22% O; 18N: 50% A, 50% O
• So, ventilation of the tropics by atmosphere + ocean MOC’s provides \( \sim 5 \text{ pW} \ (5 \times 10^{15} \text{ W}) \); distributed over the area of the Earth between 0N and 30N, averages \( 5 \times 10^{15} \text{ W/} \pi R^2 = 39 \text{ W m}^{-2} \), delivering the same amount per m\(^2\) to the Earth north of 30N.

Fully as much heat is carried in the atmosphere by 0.8 Sverdrups (megatonnes s\(^{-1}\)) moisture flux \( \sim 2 \text{ pW} \) as by dry static energy flux. (using the heat of vaporization, 2.25 MJ/kg)

(It is useful to talk about both oceanic and atmospheric mass (water or air) transports in Sverdrups (Sv):

Gulf Stream 30-120 Sv
Antarctic Circumpolar Current \( \sim 180 \text{ Sv} \)
Atlantic MOC \( \sim 16-20 \text{ Sv} \)
westerly winds/jet stream \( \sim 500 \text{ Sv} \)
atmospheric MOC \( \sim 50 -100 \text{ Sv} \)
Qnet, net atmosphere-ocean heat flux, watts/m² (Keith Tellus 95) (annual average)

It should be noted that because the sun heats the ocean, O, but does not cool the atmosphere, A, the most useful maps of Qnet for A will differ those for O by the short-wave insolation.
Where is air-sea heat flux most intense? January (W m⁻²) (SOC/NOC1.1a climatology based on COADS)
The air/sea heat flux seen by the atmosphere (latent+sensible+long-wave rad) and by the ocean (latent+sensible+long-wave + short wave solar rad)
Annual average ratio of convergence of heat flux by ocean circulation divided by annual average heating of the atmosphere by ocean: \[
\frac{(LH+SH+LW+SW)}{(LH+SH+LW)}
\]
cold-air outbreaks: a source of deep convection
(surface air temperature, 2 Jan 1993)
Fig. 8. The column-averaged diabatic heating field in Jan obtained from the NCEP–NCAR reanalysis as described in the appendix. The contour interval is 0.5 K day$^{-1}$. 
A baroclinic vortex is created by injecting water at mid-depth into a stratification. Note purple dye shows a zonal (that is, velocity exists above and below the water mass). The MOC (meridional circulation) drives 3 vortices.

cyclones
anticyclones
 cyclones
PV inversion: using a model of convective destruction of PV. The modelled or diagnosed PV field is associated with a field of azimuthal circulation, displaced mass, and interacts with the meridional overturning circulation.

**Figure 37.** PV inversion for a mixed patch with (a) inhomogeneous and (b) homogenous boundary conditions at the surface. PV distribution, isopycnals, and currents are plotted. In Figure 37a the potential density at the sea surface is specified and an idealized interior PV anomaly inverted to give the hydrography and azimuthal velocity of a baroclinic vortex. In Figure 37b an interior PV field identical to that of Figure 37a is used, but now the cold surface is represented by a sheet of high PV just beneath the upper boundary, which is prescribed to be an isopycnal surface. Note that in Figure 37b, unlike Figure 37a, the isopycnals cannot cut the upper surface, which itself is an isopycnal.
viscous overturning in a rotating cylinder:
the radial/vertical plane transmits stress from the top plate (which is at rest in the laboratory frame) and the bottom of the cylinder (which is rotating)
Overturning cells in an annulus of fluid between concentric cylinders (the inner cylinder is rotating, the outer cylinder is stationary (Taylor-Couette flow).

The cells transmit torque between the solid cylinders more strongly than would pure viscous diffusion.

(The same 2D equations govern thermal convection, and the Nusselt number expresses the analogous increase in heat flux above the diffusive rate).
MoOs organized by double diffusion

Stirring disk

Center
disk drives an anticyclonic
(warmed eddy) in uniformly
stratified fluid
Sink-driven flow in a rotating, stratified fluid: the cyclonic spin of the fluid would be resisted by bottom Ekman friction (and all radial inflow concentrated there in this tornado vortex); However, stable stratification resists and forces continuing MOC within the fluid. The azimuthal velocity
Dense plume flowing down a sloping valley in a rotating fluid (model of dense downslope flows in the Weddell Sea)

Elin Darelius, Univ of Washington GFD lab

Particle paths are helical, with Ekman-driven meridional overturning transmitting the boundary stress into the fluid. (Looking up the sloping valley)

Figure 17: The "Ekman Helix" traced out by dye injected in the bottom boundary layer seen a) up the canyon and b) from above. The secondary circulation causes a particle to follow a helix-like path down the canyon.
The zonally averaged overturning streamfunction, North Atlantic/Arctic model of Håkkinen driven by NCEP winds and temperatures

This image of the ocean circulation is the usual output of climate models; many essential processes are made invisible...the east-west detail of the previous slides. These ‘details’ are likely to be essential to understanding the global ocean transports.

The tendency for dominant sinking south of Greenland in low-resolution climate models is widespread: here in density-latitude space the streamfunction reveals higher latitude sinking and dense overflows.

The difference is expected from the east-west tilt of potential density surfaces, so that equal and opposite meridional velocities at the same depth z may have very different densities.

Bailey, Håkkinen, Rhines 2003
Lumpkin & Speer’s 2003 discussion of the Atlantic MOC, here plotted against potential density and latitude. Even though we know there is much east-west structure (boundary currents, horizontal gyres as in Reid’s maps) the zonally averaged MOC ‘looks like’ the simple 2-dimensional box models.
The ACC is the only ocean current with The Problem (how to flow meridionally, given the absolute angular momentum constraint), yet it has ample topographic bottom slopes to lean on: these clearly balance the zonal wind stress that drives this greatest of all ocean currents.

This may be a dominant site of upwelling in the global MOC (with respect to and potential density).

Salinity at 24°W longitude.
A change in the MOC transport may be associated with some measurable change in the meridional density gradient. HadCM3 finds a very close correlation between Atlantic overturning rate and the S-N gradient of steric height from 30S - 60N through the W Atlantic. But, there is a possible oversensitivity of models to subpolar buoyancy/Labrador Sea.
An evacuated glass vessel with water in it illustrates the Clausius-Clapyron relation between vapor pressure of water and temperature. The water is pushed from the vessel in my hand to the ‘cold ball’, and the vapor pressure difference between the two ends is close to the hydrostatic pressure measured by the column’s vertical displacement. One can fill out the curve and see the greater sensitivity (to temperature) of water vapor production at high, ‘tropical’ temperature. This all works because we shake the vessel so that a thin film of water lies under my warm hand. It illustrates a key variable in the climate system. When shaken this water ‘clinks’ like metal, vapor cavities opening up and slamming shut.
These lectures will address the dynamics of oceans and atmospheres, as seen through theory, laboratory simulation and field observation. We will look particularly at high latitudes and climate dynamics of the ocean circulation coupled to the atmospheric storm tracks. We will emphasize the dynamics that is difficult to represent in numerical circulation models. We will discuss properties of oceans and atmospheres that are both fundamental, unsolved questions of physics, and are also important, unsolved problems of global environmental change.

**Lecture 1:**
Is the ocean circulation important to global climate? Does dense water drive the global conveyor circulation? Fundamental questions about oceans and atmospheres that are currently under debate.

- The field theory for buoyancy and potential vorticity.
- Basic propagators: Rossby waves and geostrophic adjustment.
- Potential vorticity: inversion and flux.

**Lecture 2:**
How do waves and eddies shape the general circulation, gyres and jet streams?
- Almost invisible overturning circulations.
- Lessons from Jupiter and Saturn.
- The peculiar role of mountains, seamounts and continental-slope topography.

**Lecture 3:**
Dynamics of ocean gyres and their relation with the global conveyor circulation.
- Water-mass transport, transformation and air-sea exchange of heat and fresh water.
- Ocean overflows and their mixing.
- Decadal trends in the global ocean circulation.

**Lecture 4:**
Heat, fresh-water, ice: convection in oceans and atmospheres and the texture of geophysical fluids.

**Lecture 5:**
Teaching young students about the global environment using the GFD laboratory: science meets energy and environment in the lives of Arctic natives

**Seminar:**
Exploring high-latitude ocean climate with Seagliders and satellites.
Kelvin waves, inertial waves in shallow rotating fluid