MODELLER FOR

ØSTERSØEN - STOREBÆLT - KATTEGAT

Seminar afholdt på Danmarks tekniske Højskole

16. Marts 1987

DANSK VANDBYGNINGSTEKNISK SELSKAB

DANISH SOCIETY OF HYDRAULIC ENGINEERING

v/ H. F. Burcharth, AUC, Sohngårdsholmsvej 57, 9000 Aalborg. Tlf. 08 - 142333

FORORD

Nærværende indlæg hidhører fra et seminar afholdt den 16. marts, 1987 af Dansk Vandbygningsteknisk Selskab om emnet:

'Modeller for Østersøen - Storebælt - Kattegat'

Enkelte tidligere udgivne publikationer er tilføjet som supplement til indlæggene fra seminaret.

På selskabets vegne bringes hermed en tak til alle, der ved skriftlige bidrag eller deltagelse i diskussioner har medvirket til en orientering om et vigtigt felt inden for vandbygningsinstitutternes område.

Helge Gravesen

Helge Gravesen
Sekretær, Dansk Vandbygningsteknisk Selskab

INDHOLDSFORTEGNELSE

ØSTERSØENS FORBINDELSE MED NORDSØEN

Fl. Bo Pedersen, ISVA, DTH.

DIVERSION OF THE RIVER NEVA.
How will it influence the Baltic Sea, the Belts and Cattegat.

Fl. Bo Pedersen, Jacob Steen Møller, ISVA, DTH.

FLERLAGSMODELLER I HYDROGRAFIEN.
Styrker og svagheder ved eksisterende modeller.

Jacob Steen Møller, DHI.

MECHANISMS RESPONSIBLE FOR OXYGEN CONDITIONS IN THE DEEP WATER OF THE KATTEGAT AND THE BELT SEA.

Hans Schrøder og Jacob Steen Møller, DHI.

ÅRSAGEN TIL OG EFFEKTEN AF EUTROFIERING I KATTEGAT OG BÆLTHAVET.

Gunni Ærtebjerg, Miljøstyrelsens Havforureningslaboratorium.

Deltagerliste for VBS's "aftenmøde" mandag den 16. marts 1987.

Modeller for Østersøen - Storebælt - Kattegat.

Kampsax, Dagmarhus, 1553 København V.	L. Steding Jessen J. A. Jensen
Civilingeniør Karsten Mangor, Stampetoften 14, 2970 Hørsholm.	
Stud.ing. Henrik Kofoed-Hansen, Jægersborgvej 27, 2800 Lyngby.	
Christiani & Nielsen A/S, Vester Farimagsgade 41, 1501 København V.	Civ.ing. Ole Jensen Civ.ing. Ole A. Madsen Peter Lange
B. Højlund Rasmussen Rådg. Civilingeniører, Nørregade 7A, 1165 København K.	Kaj Vaaben Flemming Schaarup Aravinda Joshi Claus Højlund Rasmussen Palle Sejer Larsen Kim Nielsen
Rambøl og Hannemann Rådgivende ingeniører A/S, Teknikerbyen 38, 2830 Virum.	Lars Erik Storck Lars Eriksen Lars Walh Andersen Vang Henning Poulsen
Miljøstyrelsens Havforureningsl. Jægersborg Alle 1 B, 2920 Charlottenlund.	Torben Jacobsen
COWIconsult Rådgivende ingeniører A/S, Teknikerbyen 45, 2830 Virum.	Klaus Ostenfelt Ole Damgaard Larsen Søren Degn Eskildsen Jens L. Jensen
Rådg. Ingeniørfirma Steensen & Varming Århus ApS, Krabbesholms Allé 1, 8260 Viby J.	Paul Fusager
Københavns Teknikum, Ingeniørhøjskolen, Prinsesse Charlottes Gade 38, 2200 København N.	Civiling. Birthe Rodevang
DSB, Proj.org. Faste Forbindelser, Møntergade 1, 1116 København K.	Aing. B. Juul-Jensen
ISVA, bygning 115, DTH, Lundtoftevej 100, 2800 Lyngby.	Lic.techn. Fl. Thunbo Christensen Civiling. J. Skourup
Ove Bundgaard, Dalstrøget 23, 2860 Søborg.	
Akademiingeniør Knud Christiansen, Louisehøj 3, 2880 Bagsværd.	Ole Juul Jensen Mogens Hebsgaard Anders Søvsø Michael Arnskov Jørn Juhl Doris Mükletén

Dansk Hydraulisk Institut, Agern Allé 5, 2970 Hørsholm.	Asger Kej Ross Warren Erland Ramussen Ole Jensen Lars Amdisen Hans Jacob Vested Morten Rungø
Aalborg Universitetscenter, Postbox 159, 9100 Aalborg.	Torben Larsen
A/S Dansk Geoteknik, Granskoven 6, 2600 Glostrup.	5 personer
Carl Ringe, Egedalsvænge 17^{II}, 2980 Kokkedal.	
Københavns Havnevæsen, Nordre Toldbod 7, 1259 København K.	K. Holm Jørgensen Ole Bertelsen
Dansk Geoteknik A/S, Granskoven 6, 2600 Glostrup.	A. Buhl Petersen N. Danielsen
Fyns Amtskommune, Amtsgården, Ørbækvej 100, 5220 Odense SØ.	Gorm Larsen + 1 person mere

Professor H. Lundgre n.

ØSTERSØENS FORBINDELSE MED NORDSØEN

v. Fl. Bo Pedersen, ISVA

24. marts, 1987
HG/be
4001/01-OD8

ØSTERSØENS FORBINDELSE MED NORDSØEN

Foredrag afholdt i Dansk Vandbygningsteknisk Selskab, 16. marts, 1987.
Fl. Bo Pedersen, ISVA
Tema: Fast forbindelse over Store Bælt

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Pl 1

1. ØSTERSØEN I GEOLOGISK PERSPEKTIV

Den Baltiske Issø blev dannet langs den smeltende iskap. Det var en ren ferskvandssø, idet den lå højere end havniveau med et udløb gennem midt-Sverige. Efterhånden som isen smeltede på den nordlige halvkugle, steg havet - og lidt forsinket fulgte landet med. Da havniveauet passerede landniveauet fik vi en helt ny situation med et saltvandsbassin, nemlig YOLDIA-HAVET. I løbet af de næste ca. 1000 år overtog landhævningen imidlertid føringen, og igen fik vi en periode med ferskvand, den såkaldte ANCYLUS-SØ med et flodafløb gennem STORE BÆLT. Denne periode varede også ca. 1000 år, hvor den fortsatte havstigning kombineret med en sydlig nedadrettet vipning af landjorden bragte Østersøen tilbage til et saltvandsbassin - LITTORINER-Havet, som var noget saltere end den nuværende Østersø. I de næste par tusinde år foregik der en svag landhævning i syd - som nu er stoppet - og en stærkere i nord, som stadig pågår (ca. 1 cm per år). Den svage landhævning i syd bremsede op for saltvandsindstrømningen, således at Østersøen gradvis blev mere brak, med en nuværende saltholdighed på ca. 1/4 af havvandets saltholdighed. Da såvel Havniveau som Landniveau (ved udløb) har stabiliseret sig, er de variationer i saltholdigheden, vi har observeret i historisk tid, affødt af andre ydre påvirkninger, primært VINDEN og NEDBØREN.

Af denne korte geologiske redegørelse kan vi konkludere, at Østersøens saltholdighed - og dermed dens livsbetingelser - er uhyre følsom over for de geometriske forhold ved udløbet, d.v.s. overfor forholdene i Store Bælt og i Øresund. Lukker vi helt af, bliver Østersøen en ferskvandssø. Abner vi helt op, bliver Østersøen et saltvandshav.

Pl 2

2. NÅR SALTVAND OG FERSKVAND MØDES

Et af de forhold, der komplicerer de hydrografiske problemer markant i de indre danske farvande og i selve Østersøen, er lagdelingen. Denne optræder f.eks., når ferskvandet fra floderne møder saltvandet i oceanerne. Det salte vand - der er lidt tungere end det ferske - vil kile sig ind under det ferske vand. Hvis ikke der var en mærkbar blanding mellem fersk- og saltvand, ville vi få to lag, hvoraf det nederste, på grund af mangel på adgang til luftens ilt, efterhånden ville dø totalt ud. Heldigvis er ferskvand og saltvand blandbare, således at der hele tiden foregår en fornyelse af det dybere liggende vand.

En blanding af to vædsker kræver en energitilførsel. I naturen kommer den for blandingen nødvendige energi fra forskellige kilder, hvoraf vinden er den vigtigste i Østersø-systemet. Nu er det imidlertid langt fra al den tilførte energi, der bruges til blanding. F.eks. ved vi, at størsteparten af den energi vinden tilfører vandet, går til bølgedannelser, til vindstuvning og strømninger - og dermed til produktion af turbulent kinetisk energi, der igen omdannes til varme. Når vinden holder op, forsvinder bølgerne, vindstuvningen og strømningerne, men på grund af blandingen har vi fået løftet lidt af det tunge saltvand op i det lette brakvand - vi har altså vundet lidt potentiel energi, svarende til det arbejde, der er præsteret mod tyngdekraften. Virkningsgraden af denne blanding - altså forholdet mellem den vundne potentielle energi og den tilførte energi - kaldet Flux Richardsons tal - har vist sig at være meget nær 5% i mange typer af lagdelte strømninger. Som et konkret eksempel kan jeg f.eks. nævne blandingen i Kattegat som følge af vinden. Med kendskab til vindforholdene og de deraf affødte strømninger, er det en smal sag at beregne den tilførte effekt (PROD) - og dermed ved brug af den veldokumenterede empiriske virkningsgradsfaktor - at beregne den af vinden forårsagede medrivning (POT) gennem skillefladen mellem det tunge og det lette vand.

I Østersø-sammenhæng bliver vi imidlertid også nødsaget til at se på andre energiinput for blandingen, nemlig de meget kraftige strømninger gennem Bælthavet samt de strømninger, der foregår langs bunden af tungt vand fra bassin til bassin.

Pl 3

3. STRØMNINGEN GENNEM STORE BÆLT

De meteorologiske forhold over Østersøområdet er kendetegnet ved en forholdsvis regelmæssig passage af lavtryk og højtryk. Et lavtryk over Skandinavien betyder vestlige vinde, som dels giver opstuvning i Skagerrak, dels vandspejlssænkning i den vestlige Østersø, altså en vandspejlshældning fra N til S, hvilket genererer en strømning ind i Østersøen. Strømningsmodstanden - og dermed vandspejlshældningen - vil være størst gennem de lavvandede snævre passager, altså Bælterne og Øresund. Ved højtryk over Skandinavien får vi tilsvarende østlige vinde, der skaber højvande i den vestlige Østersø og lavvande i Skagerrak og dermed en udstrømningssituation, igen med det største vandspejlsfald gennem Bælthavet og Øresund. Den periodiske ind/udstrømning gennem Store Bælt og den dertil hørende vandspejlsforskel mellem Gedser og Hornbæk kan vi f.eks. illustrere for en 3 måneders periode i 1976, hvor vi takket være Bæltprojektet har kendskab til vandføringerne. Vi bemærker, at der er en stor korrelation mellem vandføringerne og vandspejlsdifferenserne, hvilket betyder, at Store Bælt i store træk opfører sig som en almindelig kanalstrømning. En nøjere beregning ud fra kendskab til de geometriske forhold i denne "kanal" viser da også, at hvis man sammenlægger friktionstabene fra bund- og skilleflade med de enkelttab, der forekommer ved bratte arealvariationer, så får man god overensstemmelse mellem teori og målinger. Dette forhold spiller en afgørende rolle for en beregning af konsekvenserne af at ændre på de geometriske parametre i Store Bælt - et emne, som Ottesen Hansen vil komme nærmere ind på.

Det næste forhold, der skal bemærkes, er, at nettovandføringen over en længere periode er meget lille i forhold til de dagligt observerede vandføringer. Nettotransporten ud gennem Store Bælt udgør ca. 60% af ferskvandstilførslen til Østersøen, altså ca. 12.000 m^3/s ($\sim$

1 km^3/dag), medens en typisk ud/indstrømning er af størrelsesordenen 100.000 m^3/s (8,6 km^3/dag). Dette forhold har stor betydning for blandingen i Store Bælt (effekt input $\sim \rho g Q \Delta H$), men også for bundvandsfornyelsen i de dybe bassiner inde i Østersøen, hvilket netop er de to parametre, der er truffet politisk beslutning om ikke at ændre i forbindelse med udførelsen af den faste forbindelse over Store Bælt. Vi skal vende tilbage til dette kravs konsekvenser, men inden da skal vi se på Østersøens vandskifte.

pl 5

4. ØSTERSØENS VANDSKIFTE

Et længdesnit gennem Østersøen viser de typiske karakteristika for en fjord, nemlig et system af bassiner adskilt af tærskler. For det første blokerer tærsklerne for en fri udveksling mellem havets og fjordens dyberëliggende vandmasser. For det andet virker tærsklerne som ensretterventiler for det tungere vand, der passerer dem under indstrømningssituationer. Denne ensrettereffekt er særlig markant ved Darss-tærsklen i Store Bælt og Drogdentærsklen i Øresund. Mekanismen kan vi illustrere for Darss-tærsklen:

De kraftige strømninger og deraf affødte blandinger i Bælthavet resulterer i, at saltholdigheden har store langsgående variationer, i gennemsnit fra 7,8 0/00 til 18 0/00 i overfladelaget og fra 16 0/00 til 33 0/00 i bundlaget, begge voksende fra Darss til Kattegat. Når vandet strømmer ind i Østersøen, vil saltholdigheden ved Darss være stadig stigende, og da det salte vand er tungere end det vand, det møder i Arkonabassinet, vil det søge mod bunden og fortsætte som en tung bundstrøm mod Bornholmerbassinet. Observationer ved Bornholm viser tydeligt eksistensen af denne bundstrøm - forøvrigt kraftigt påvirket af Corioliskraften.

Efter en opholdstid på ca. 4 måneder i Bornholmerbassinet vil det omtalte vand passere den næste tærskel ind til den centrale Østersø, hvor det får lov til at opholde sig i ca. 30 år, før det genser Store Bælt. Den faldende saltholdighed undervejs er et resultat af, at de tunge bundstrømme river vædske med sig fra det lettere omgivende vand. Dermed øges mængden af det vand, der "lufter" Østersøens dybere beliggende vandmasser. For at lukke vandets kredsløb kræves der en opadrettet transport gennem Østersøens haloklin (salt-skilleflade). Den primære energikilde til denne medrivning er vinden.

Hvis vi vender tilbage til Bornholmerbassinet, bemærker vi, at der er en markant større saltholdighed helt nede ved bunden i sammenligning med den saltholdighed vi finder i det gennemstrømmende vand (mellem kote -45 m og -60 m). For at forny dette næsten stillestående bundvand kræves der en indstrømning af meget salt vand, som kan løfte det gamle vand op i gennemstrømningsområdet. Sådanne saltvandsindbrud forekommer med års mellemrum, som illustreret for Gotlandsbassinet, hvor vi kan se, at de medfører en stigning i saltholdigheden og samtidigt en øgning af vandets iltindhold. Man bemærker imidlertid også, at der er tale om en stakket glæde, hvilket har sin forklaring i, at ilten forbruges til bakteriologisk nedbrydning af dødt plankton m.m. Stagnationsperioder med svovlbrinte i bundvandet har eksisteret gennem flere århundreder i Østersøen. Det beklagelige er, at de områder, hvor iltsvind forekommer, er rykket længere udad i systemet, således at vi nu også mærker den i Kielerbugten, Bælthavet og sidste år selv i Kattegat. Disse iltsvind og medfølgende fiskedød er hovedårsagen til den store offentlige interesse, der er for alle indgreb i naturen, der kan give anledning til forandring af forholdene, inklusive bygningen af en fast forbindelse over Store Bælt - og Øresund.

Medens man efterhånden er blevet enige om, at en reduceret udledning af de for planktonproduktionen nødvendige næringssalte til de indre danske farvande vil have en gunstig effekt på miljøet, så er det langt vanskeligere at nå til enighed om, hvilken effekt en ændring af blandingsforholdene i Bælthavet og /eller en ændring af vandføringen gennem Bælthavet vil have på Østersøen. Teknisk set ville det være muligt at styre såvel blandingen i som vandføringen gennem Store Bælt, som foreslået af prof. Lundgren, men den politiske beslutning er, at vi skal finde en nulløsning.

Pl 10

5. NULLØSNINGEN FOR DEN FASTE FORBINDELSE OVER STORE BÆLT

Nulløsningen er ensbetydende med, at den faste forbindelse ikke må være årsag til forandrede forhold i Østersøen. Som vi har set, er Østersøens vandskifte afhængig af vandføringen og saltholdigheden ved Darss-tærsklen. Uforandret vandskifte kræver altså såvel, at vandføringen i Store Bælt som blandingen i Store Bælt ikke påvirkes af den faste forbindelse.

Vedr. vandføringen Q

Friktionsbestemt Q

For uændrede blandingsforhold vil vandføringen gennem Store Bælt i størsteparten af tiden være entydigt relateret til vandspejlsforskellen mellem Kattegat og Østersøen gennem den specifikke modstand K, der er bestemt af de geometriske forhold i Store Bælt. Vandspejlsdifferensen ΔH er primært bestemt af de meteorologiske forhold, d.v.s. næsten uafhængig af hvilket indgreb, vi foretager i Store Bælt. For at holde Q uforandret kræves der altså et uforandret modstandstal. Da såvel tunnel som tilslutningsramper og bropiller giver anledning til indsnævringer af tværsnittet med tilhørende ekstra energitab, skal der foretages nogle afgravninger, som kan kompensere for disse tab.

Kritisk strømning

Kanalstrømninger har den særegne egenskab, at de for en ganske bestemt indsnævring i dybden eller i bredden pludselig skifter karakter fra at være strømmende til at være strygende. Vi kender fænomenet fra Venturikanaler (breddeindsnævring) og fra overløbsbygværker (dybde indsnævringer). Disse er begge karakteriseret ved, at man for en given opstrøms vandstand får maximum af vandføring gennem det kritiske snit. I denne situation er vandføringen bestemt af forholdene i det kritiske snit - og altså ikke af friktionsforholdene.

Tiden tillader ikke, at vi kommer nærmere ind på kritisk strømning i lagdelte vædsker, men vi kan dog illustrere de strømningssituationer, man kan komme ud for i det simpleste tilfælde, nemlig hvor bredde- og dybdeindsnævringen foregår i samme snit vinkelret på strømmen, og hvor vædskerne er ublandbare og friktionsløse. Man kan af illustrationerne fornemme, at blot det at introducere en lagdeling, øger løsningsrummet kolossalt. I Store Bælt har vi så yderligere, at vædskerne er blandbare og er friktionspåvirkede, samt at dybde- og breddeindsnævringerne ikke forekommer i samme snit. Dette uhyre komplekse problem findes ikke behandlet i litteraturen, så for at understøtte de teoretiske beregninger, har det været nødvendigt at foranstalte modelforsøg. Disse vil blive nærmere behandlet af Ottesen Hansen.

Vedr. blandingen

Med uforandret vandføring og energitab er der samtidig sikret, at blandingen forbliver den samme, såfremt virkningsgraden i blandingsprocessen er konstant. Jeg nævnte tidligere, at denne virkningsgrad var en empirisk størrelse, hvis værdi var fastlagt på baggrund af en fysisk begrundet hypotese kombineret med målinger i naturen og i laboratoriet. Jeg kan bedst illustrere hypotesens holdbarhed for en tung bundstrøm, hvor vi kan se, at teorien passer med målinger på strømninger lige fra vandret til lodret bund, - og lige fra laboratorierender med en typisk dimension på 0,1 m til verdens største tunge bundstrøm i Danmarksstrædet, som er ca. 200 m høj, ca. 400 km bred og med et fald på godt 2000 m. Imidlertid indeholder vort erfaringsgrundlag ikke de blandinger, der optræder i forbindelse med lokale energitab, d.v.s. netop den type strømninger en fast forbindelse over Store Bælt vil give anledning til. Det har derfor været nødvendigt at foranstalte nogle modelforsøg, som enten kan bekræfte blandingshypotesen eller give svar på, hvordan vi ellers skal sikre den uændrede blanding.

6. SAMMENFATNING

Østersøen er et komplekst dynamisk system. Selv om vi ikke kan beskrive systemet i alle detaljer, afviger det selvfølgelig ikke fra andre dynamiske systemer med hensyn til, at systemets tilstand er bestemt af randbetingelserne. Den vigtigste dynamiske randbetingelse er vinden og ferskvandstilførslen. Den vigtigste geometriske randbetingelse er Bælthavet og Øresund med tilhørende tærskler. Den vigtigste biologiske randbetingelse er tilførslen af næringssalte.

Ønsker man at ændre systemet i en eller anden retning, kræver det et nøje kendskab til samtlige involverede parametre. Det har vi som sagt ikke. Ønsker man derimod kun at foretage et mindre lokalt indgreb, kan man med stor sikkerhed foretage lokale kompensationsforanstaltninger, således at indgrebet ikke mærkes i større afstand fra bygværket.

At bygge en fast forbindelse over Store Bælt eller Øresund er set med hydrografiske og biologiske øjne det samme, som en hjerteoperation er for en læge. Vi griber ind i det vigtigste organ for Østersøens liv - det skal derfor gøres med stor omhu.

Det hele nytter imidlertid ikke noget, såfremt patienten indtager så mange næringsstoffer, at han istedet får fedt på hjertet og dør af et hjerteslag. Den nødvendige slankekur til en pris af godt 10 milliarder kr. fra dansk side og de deraf affødte aktiviteter bliver efter alt at dømme startskuddet til udviklingen af nyere og bedre modelværktøjer, et emne som Jacob Møller vil komme nærmere ind på.

7. LITTERATUR

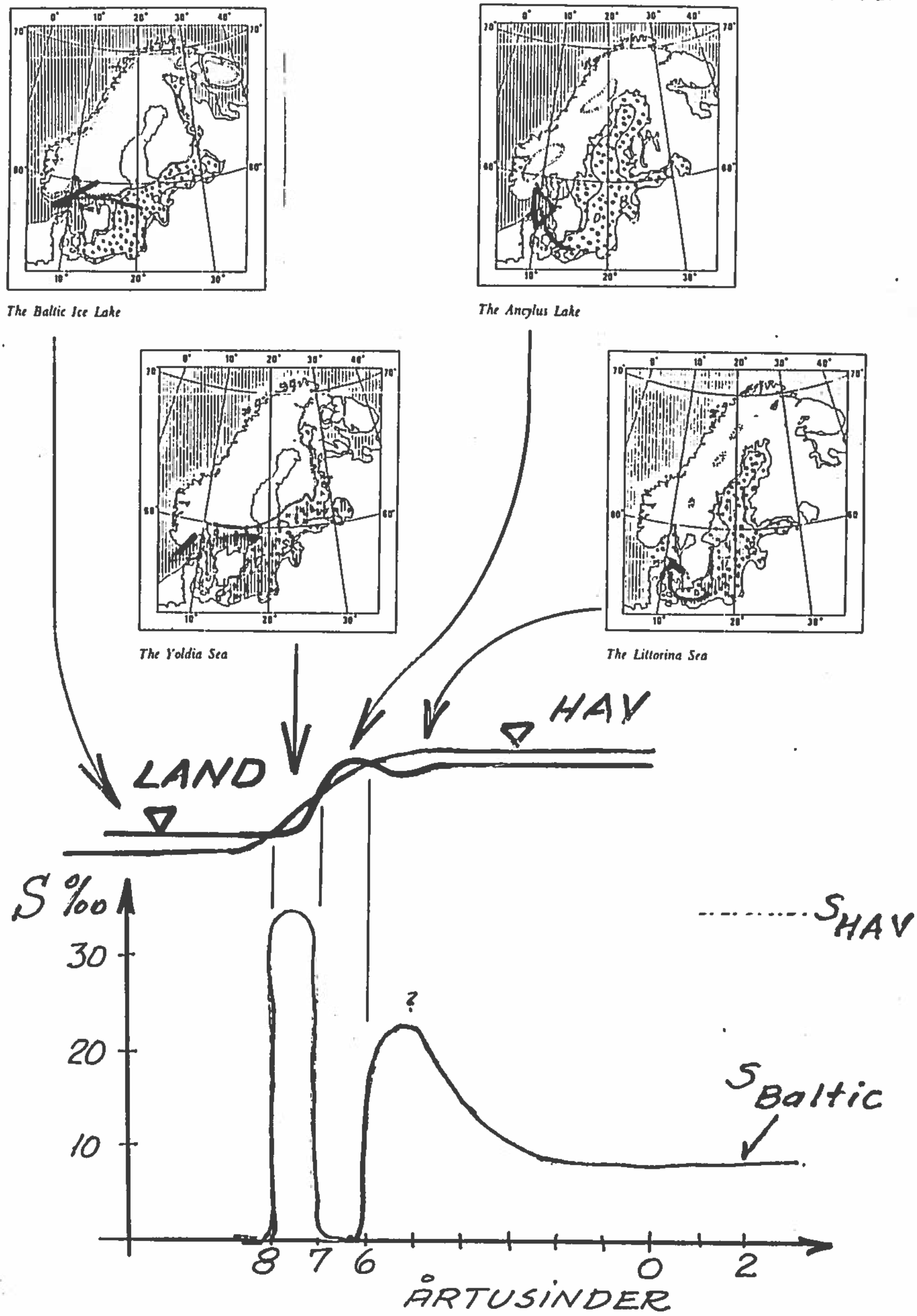
Lagdelte strømninger og blandingsprocesser:

Fl. Bo Pedersen: Environmental Hydraulic Stratified Flows. Lecture Notes on Coastal and Estuarine Studies. 18. 278 pp. Springer-Verlag, 1986.

Eksempel på konsekvensberegning for Østersøen:

Fl. Bo Pedersen, Jacob Steen Møller: Diversion of the River Neva. How will it influence the Baltic Sea, the Belts and Kattegat. Nordic Hydrology, 12, 1981, 1-20. (Genoptryk vedlagt).

PL 1


The Baltic Ice Lake
The Ancylus Lake
The Yoldia Sea
The Littorina Sea
HAV
LAND
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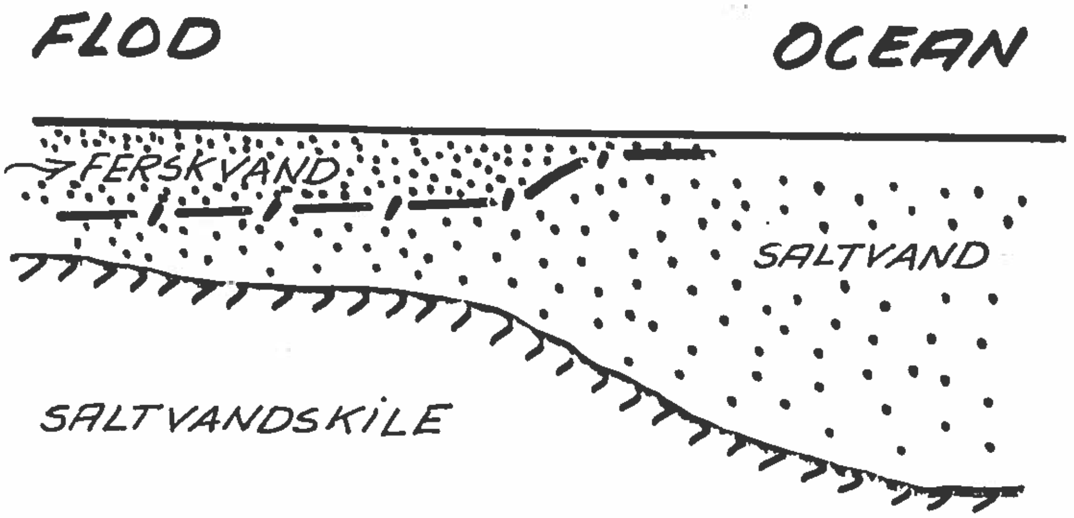
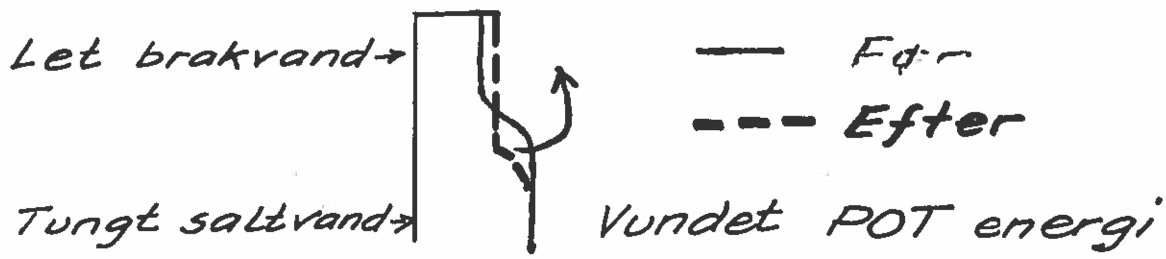


BLANDING KRÆVER ENERGI

ENERGIKILDER

1) VIND: → Bølgedannelse
+ Vindstuvning
+ Strømninger

→ PROD af Turb. Kin. Energi
→ Varme + Blanding



Virkningsgraden:

$$\boxed{R_f = \frac{POT}{PROD} \sim 5\%}$$

2 etc.) Strømning i Bælthavet
Strømning fra bassin til bassin

IN

1030 – mb

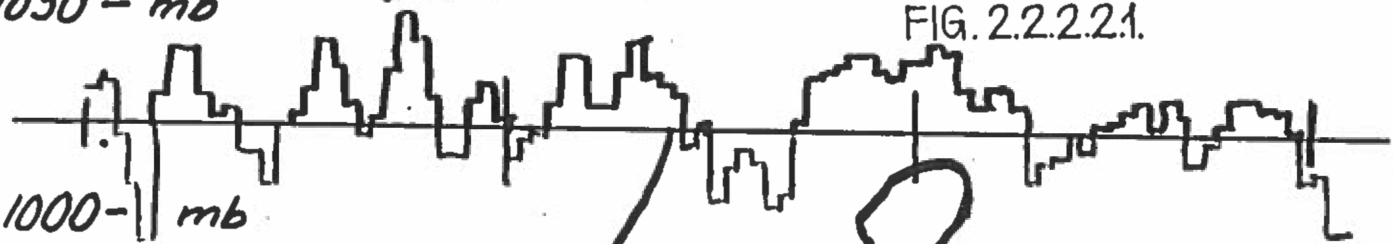
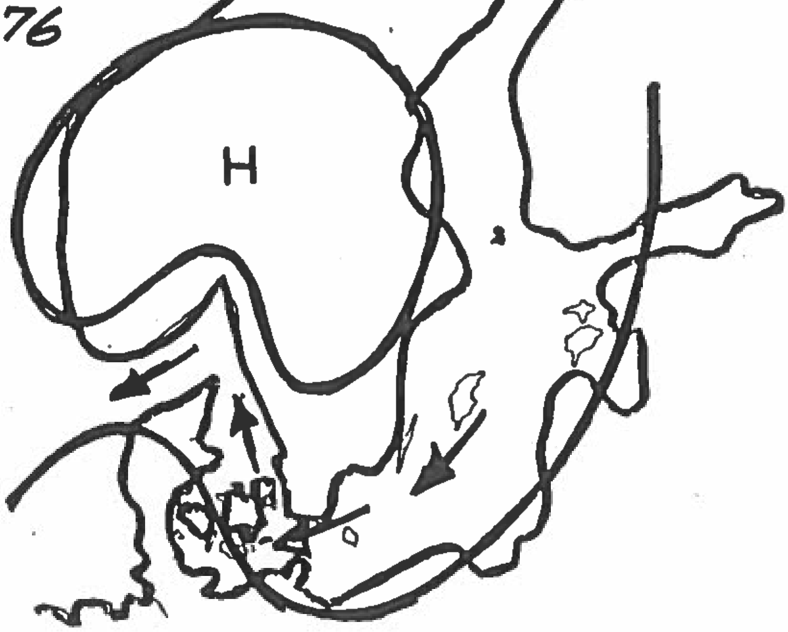


FIG. 2.2.2.2.1.

1000– mb

MAY 76

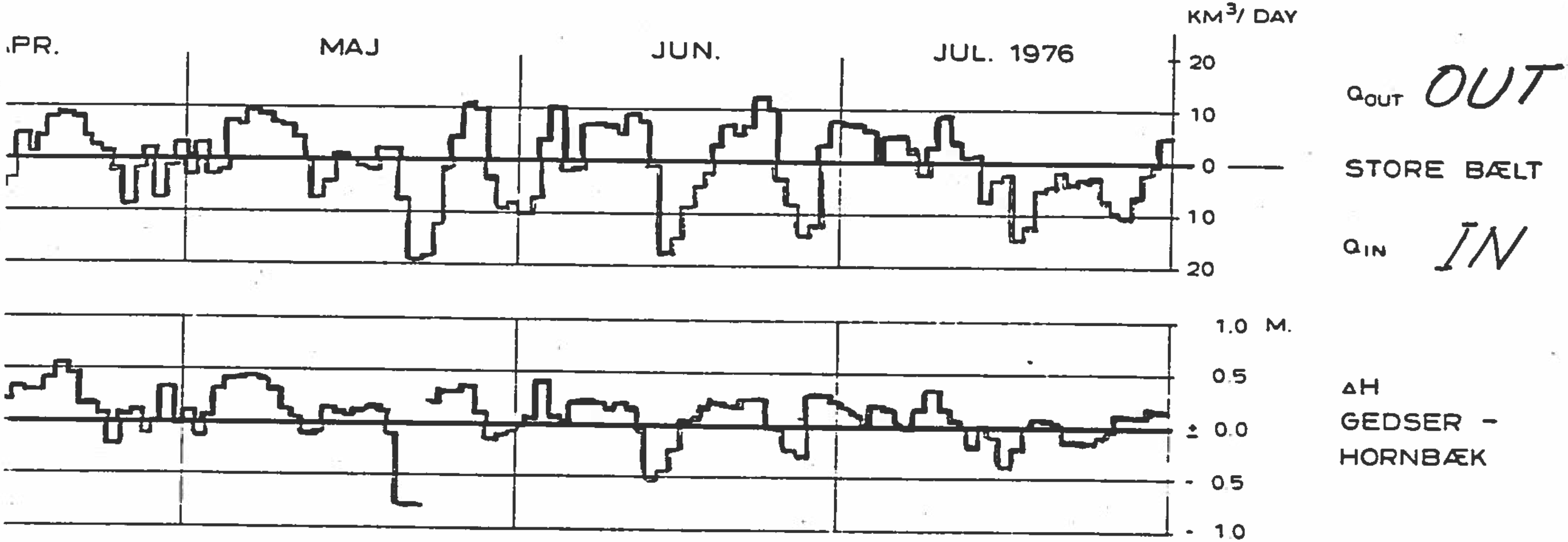
OUT



UDSTRØMNING

FIG. 2.2.2.2.2.

PL 4



$$\Delta H = K\,Q^2$$

$Q_{Netto,\,S.B.} \sim 1\ km^3/day$

$Q_{ud/ind} \sim 10\ km^3/day$

Blandingen i S.B.

Bundvandsfornyelsen i Østersøen

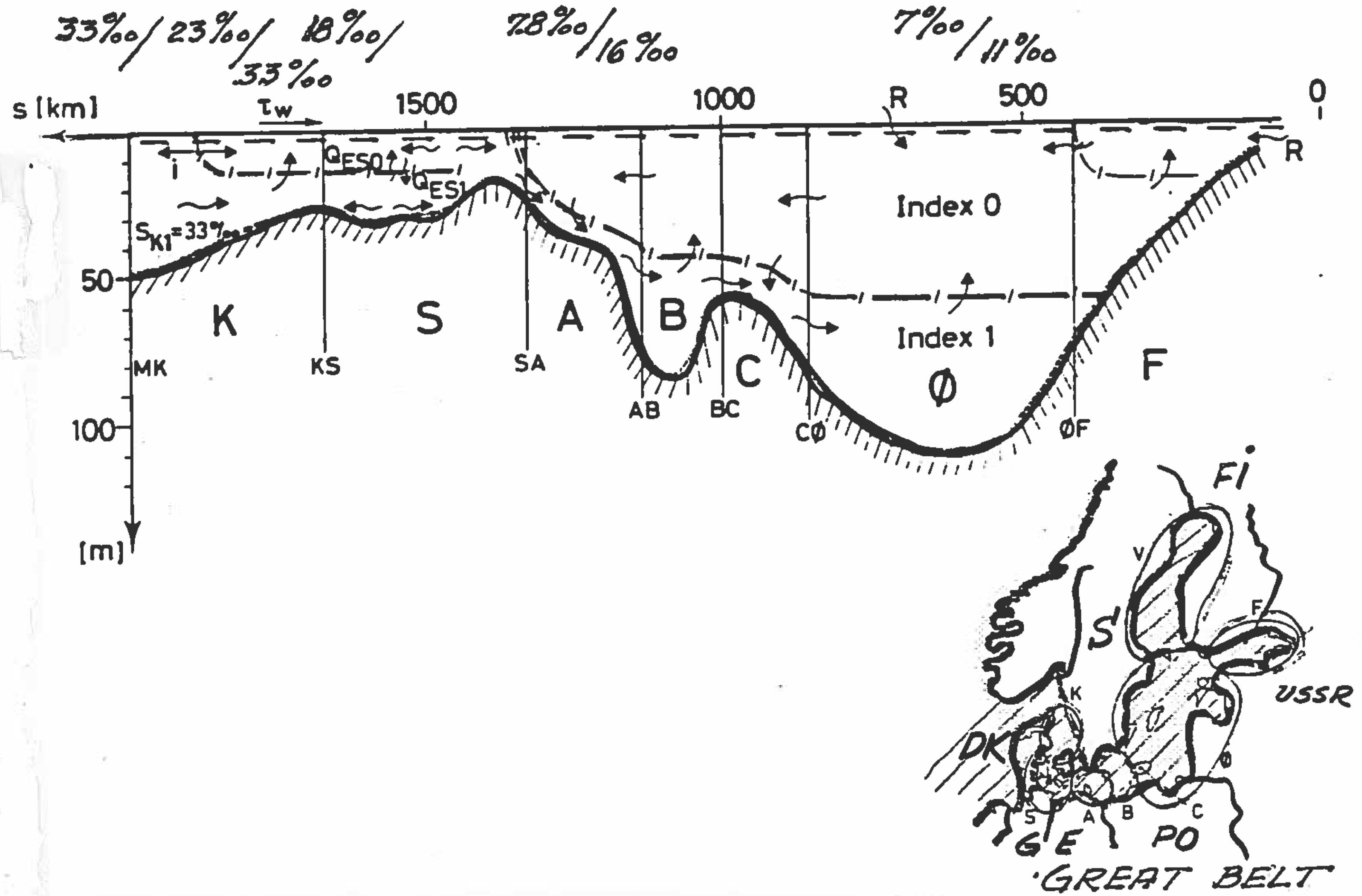
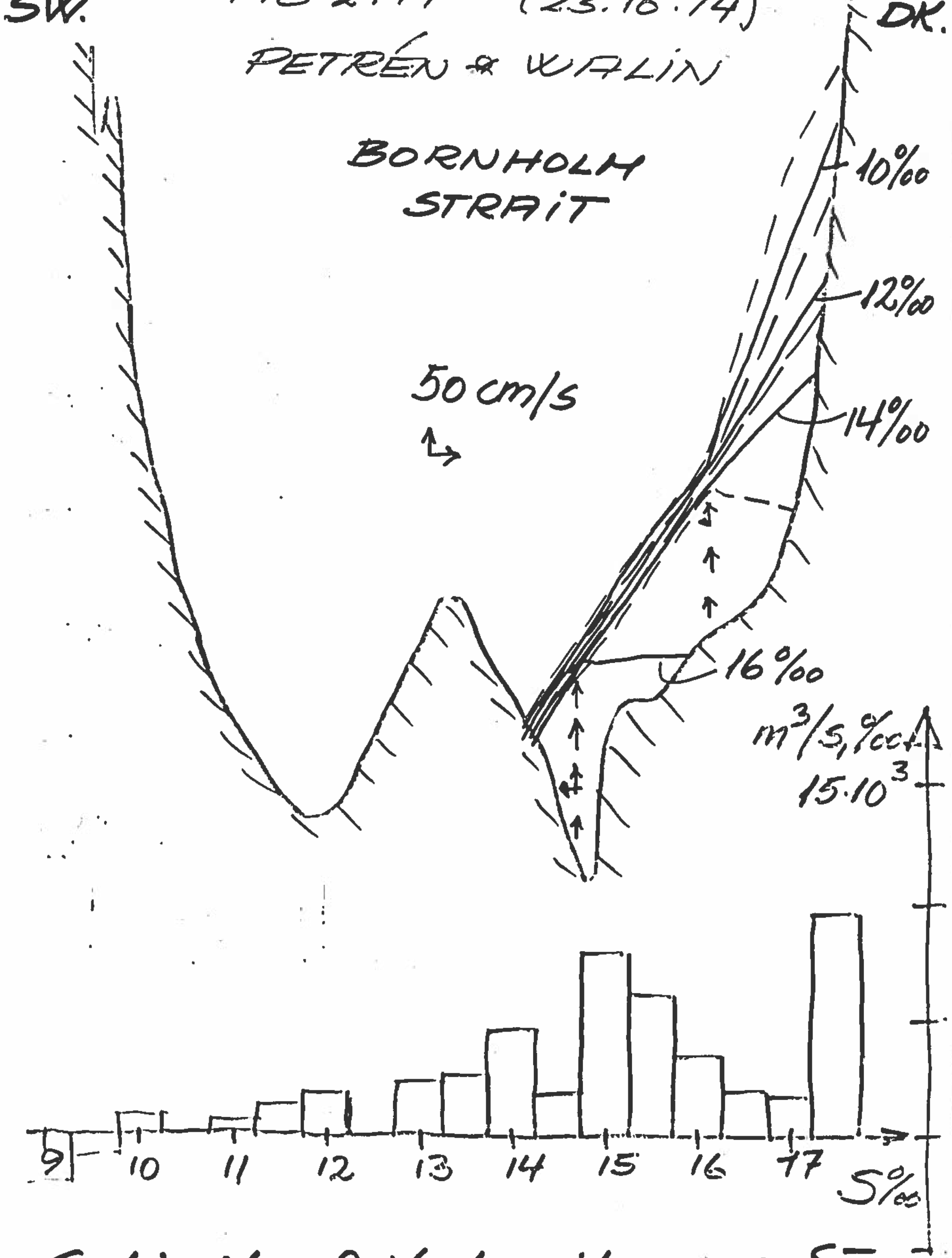


Fig. 1. The Baltic estuary system divided into eight subareas.

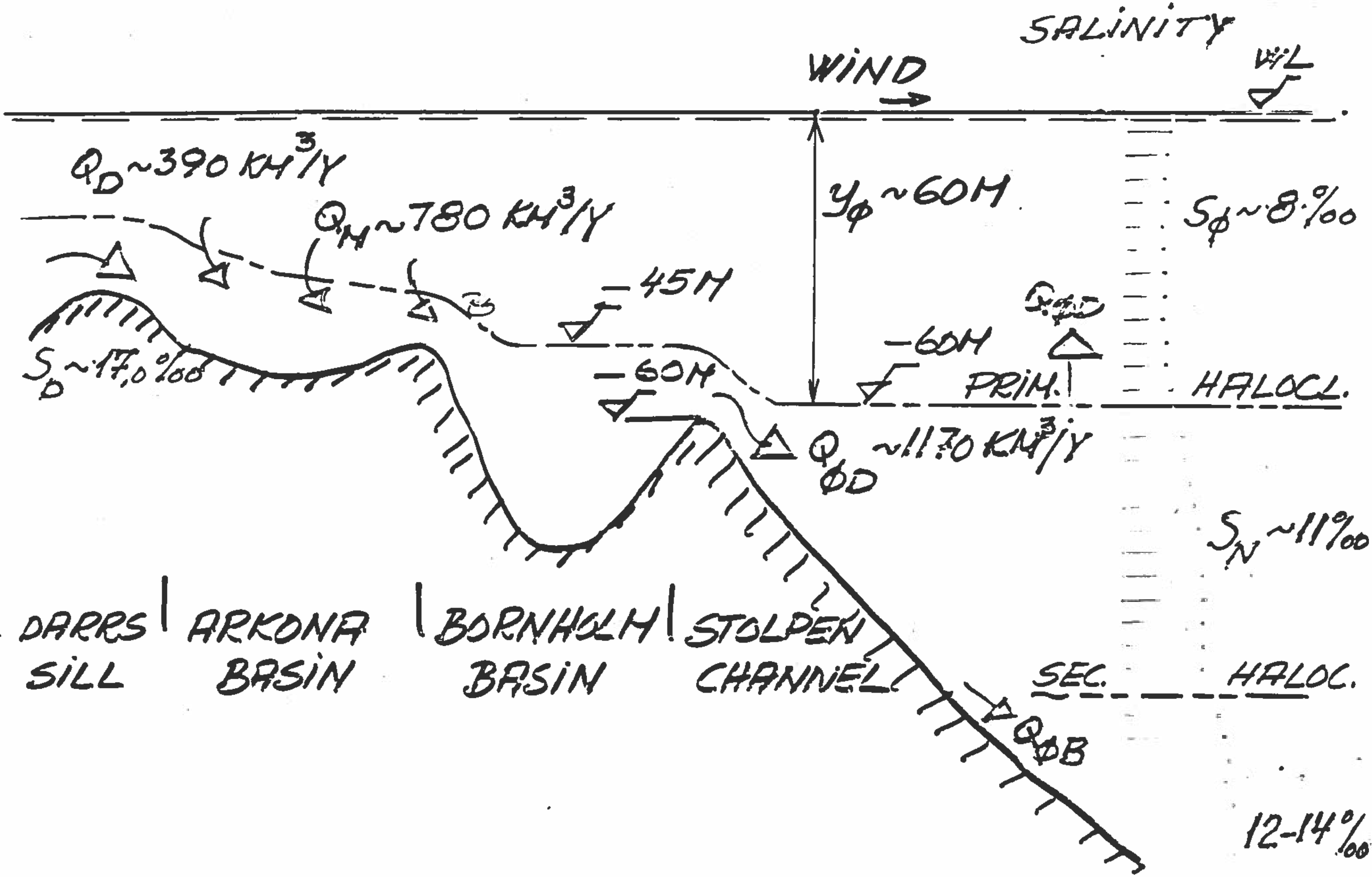
PL5

FIG 2:17 (23.10.74)

PETRÉN & WALIN

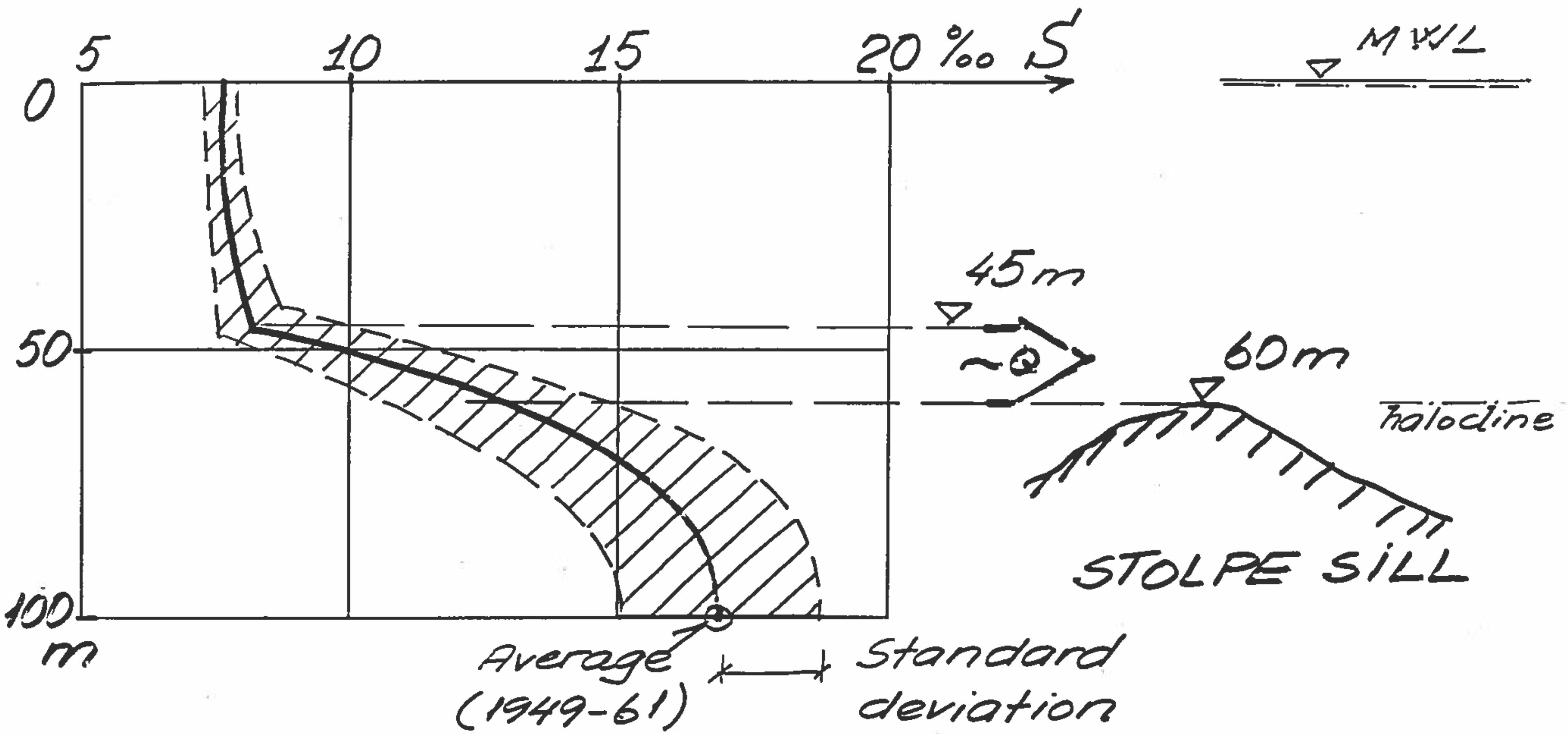


Salinity & Velocity measur.



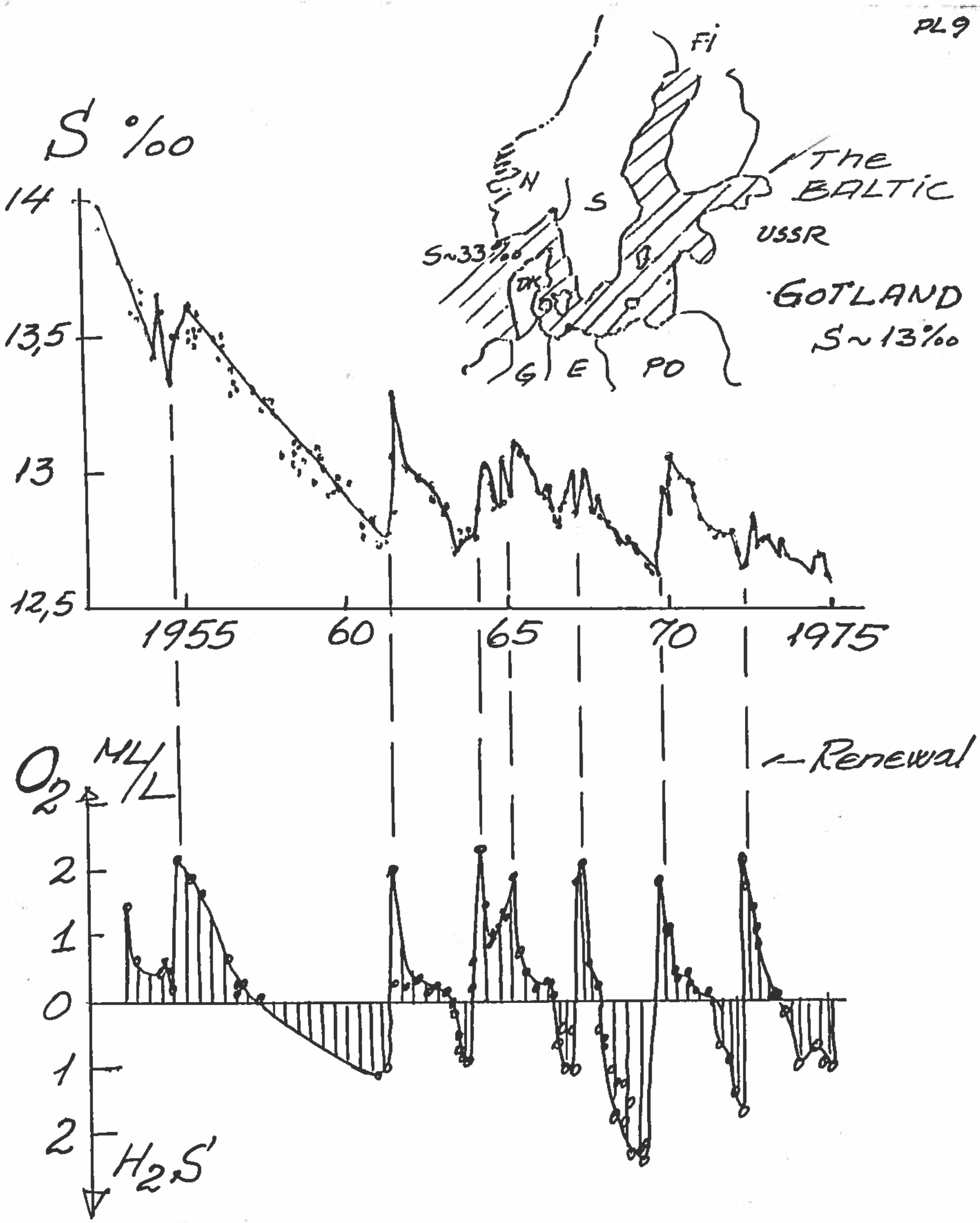
SALINITY
WIND
W.L.
Q_D~390 KM^3/Y
Q_M~780 KM^3/Y
y_ϕ~60M
S_ϕ~8‰
–45M
–60M
–60M
PRIM.
HALOCL.
$Q_{\phi D}$~1170 KM^3/Y
S_D~17,0‰
S_N~11‰
DARRS SILL
ARKONA BASIN
BORNHOLM BASIN
STOLPEN CHANNEL
SEC.
HALOC.
$Q_{\phi B}$
12-14‰

PL7

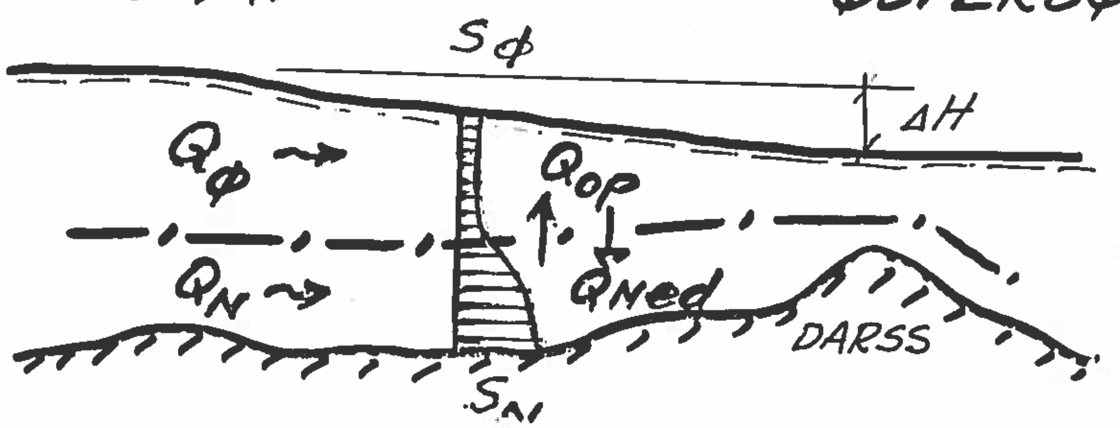


SALINITY distribution in the BORNHOLM BASIN

PLB



The GOTLAND DEEP (below 200m)



NULLØSNING: Q, S Uforandrede

UFORANDRET VANDFØRING

1. Friktionsbestemt (ofte)

Meteorologi — Politisk Diktat

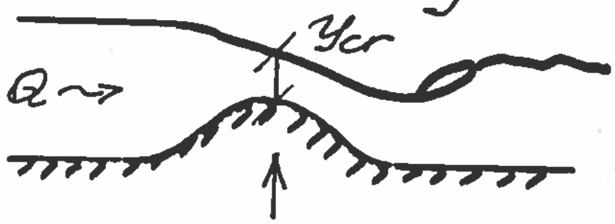
$$\Delta H = K \cdot Q^2$$

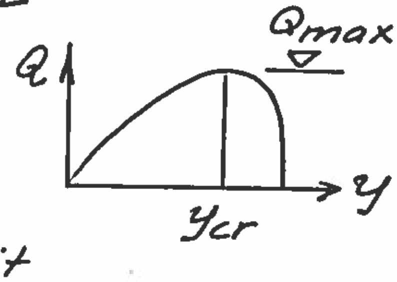
Specifik modstand ɔ: Geometri

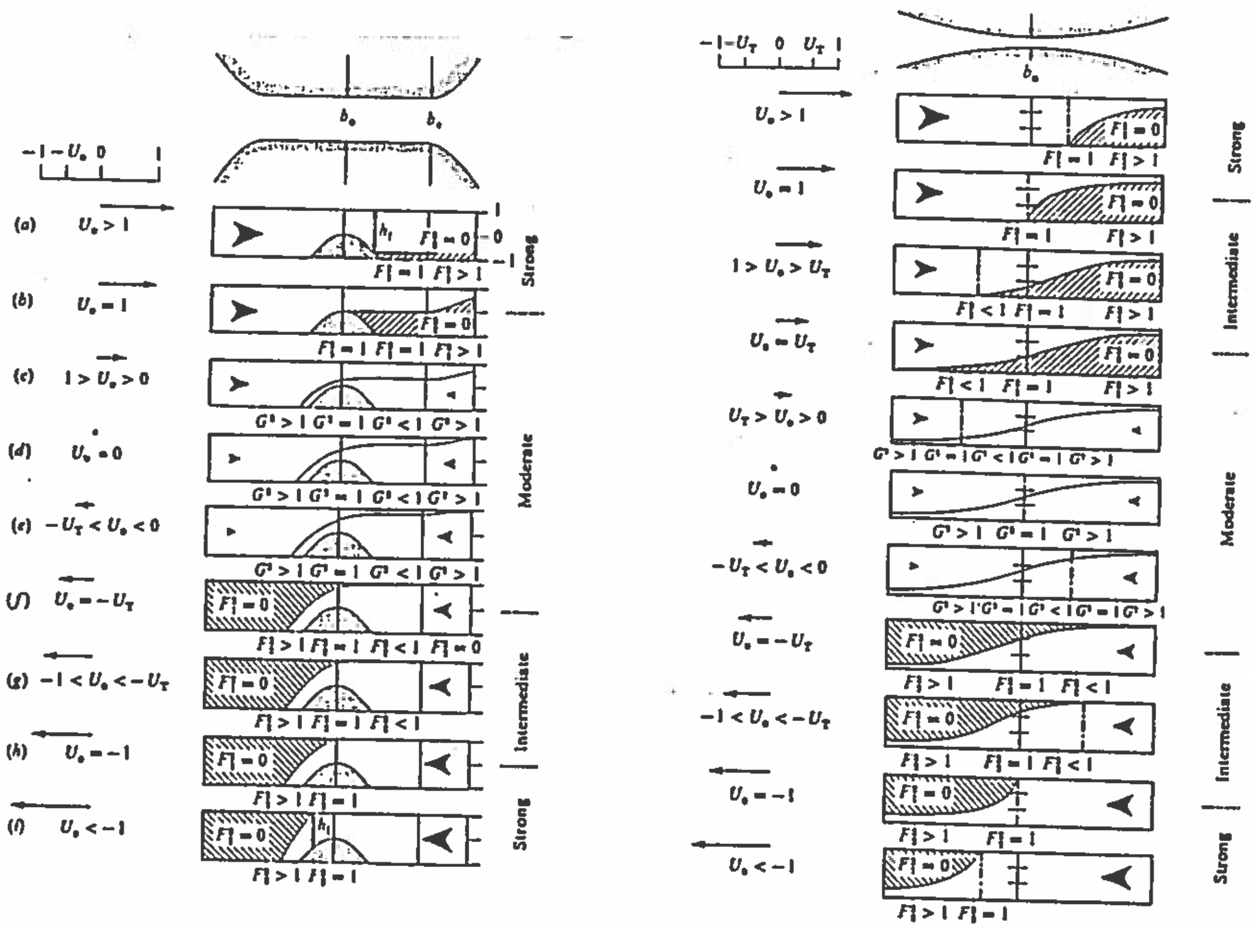
Indsnævringer (energitab) ⇒ Kompensations-afgravninger

2. Kritisk strømning (sjældent).

Eks. almindelig kanal







Kritisk strømning : 2 Lag
Ublandbare
Friktionsløs

STORE BÆLT : + Blandbare
+ Friktionspåvirkede
+ Højde / Bredde indsnævring foregår ikke i samme snit

NULLØSNING:

Q uforandret

1) Samme specifikke modstand K

2) Uændret kritisk strømning y_{cr}

S uforandret

Blanding $\sim \underbrace{\gamma Q \Delta H}_{\text{uændret}} \mathbb{R}_f$

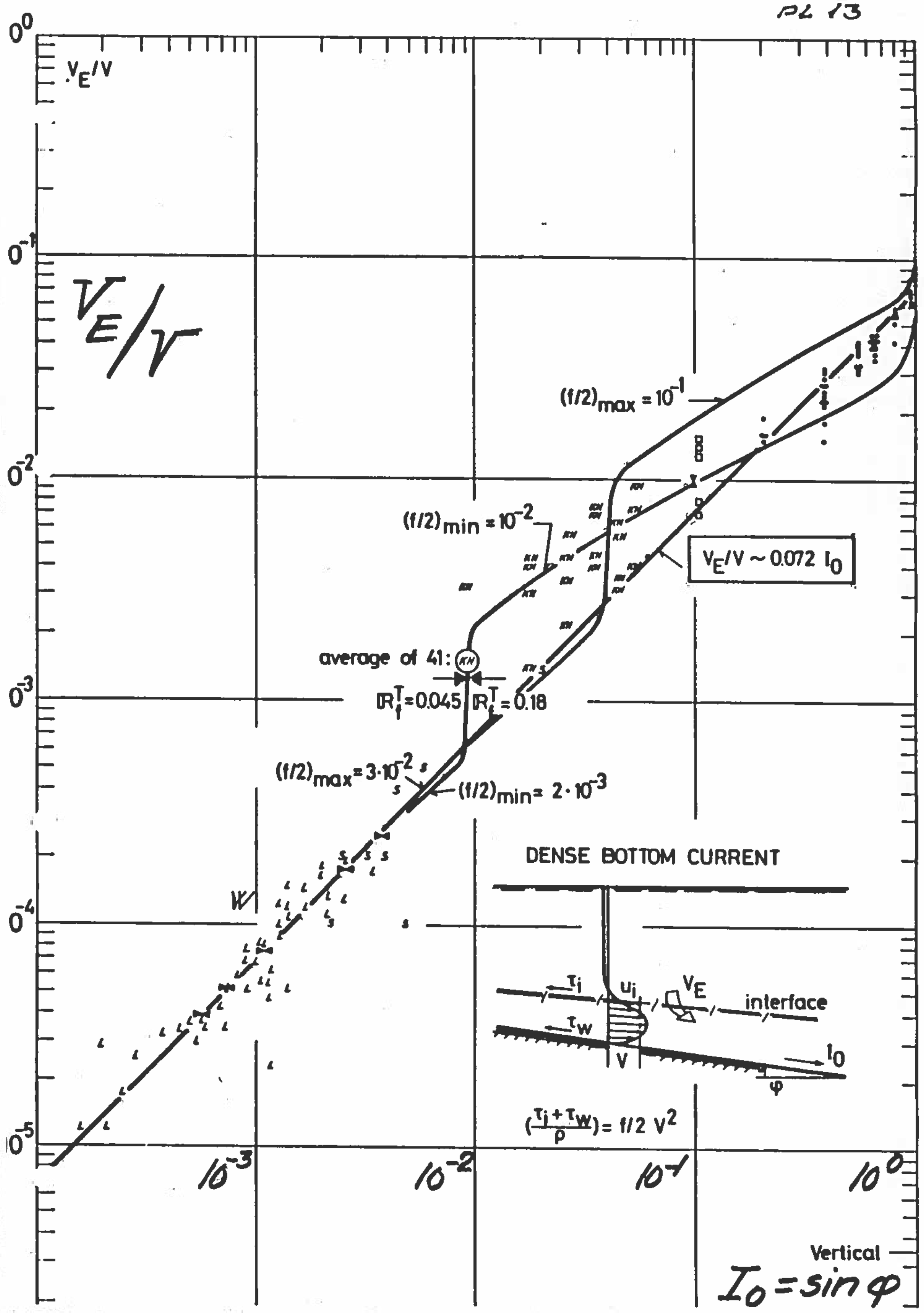
Virkningsgraden $\mathbb{R}_f$ uændret

$\mathbb{R}_f$: empirisk fastlagt ved målinger

Eksempel : Tung bundstrøm
+ mange andre, men
÷ lokale effekter

$\Downarrow$

MODELFORSØG



V_E/V
10^{0}
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
$(f/2)_{max} = 10^{-1}$
$(f/2)_{min} = 10^{-2}$
$V_E/V \sim 0.072\ I_0$
average of 41:
$R_f^T = 0.045$ $R_i^T = 0.18$
$(f/2)_{max} = 3 \cdot 10^{-2}$
$(f/2)_{min} = 2 \cdot 10^{-3}$
DENSE BOTTOM CURRENT
τ_i
u_i
V_E
interface
τ_w
V
I_0
φ
$\left(\frac{\tau_i + \tau_w}{\rho}\right) = f/2\ V^2$
10^{-3}
10^{-2}
10^{-1}
10^{0}
Vertical
$I_0 = \sin\varphi$

DIVERSION OF THE RIVER NEVA

v. Fl. Bo Pedersen & Jacob Steen Møller

Nordic Hydrology, 12, 1981, 1-20

Diversion of the River Neva

How will it Influence the Baltic Sea, the Belts and Cattegat

Flemming Bo Pedersen and Jacob Steen Møller
Technical University of Denmark, Copenhagen

Diverging part of the river Neva discharge to the dry regions in the southern USSR has raised the question, to what extent such a river diversion will influence the hydrographic conditions in the Baltic Sea and the Danish Inland waters. In order to quantify the influence, the system has been divided into eight subareas, each of which is characterized by an equation for the mass, the volume and the dynamic balance (the mixing), respectively. The man-made change in the river runoff has been introduced in the equations, which have then been linearized and solved with respect to changes in the salinities, the discharges and the layer depths in the system.

As a quantitative example the hydrographic consequences of a 25% reduction in the river Neva discharge have been outlined. The most pronounced influence is on the salinities, which are increased by 0.2 to 0.4 ‰ allover in the system. Hence, if the river diversion had become executed in the beginning of this century a 30 to 40% higher salinity-variation would have been encountered in the Baltic Sea – compared to the actual variations during this century.

Introduction

The increasing water demand for irrigational purposes in the dry regions north of the Caspian Sea and the Lake Aral (in the USSR) has actualized the plans of pumping huge amounts of water from the catchments of the river Ob and the river Neva to the river Volga, which are running through the affected dry areas. The USSR's Council of Ministers have, in fact, in their 5-year plan 1976-80 initiated the preliminary planning for diverging up to 2,000 m^3/s from the river Ob, which drains to the Arctic Sea. Although not mentioned directly in the available sparse information on the project, Mikhaylow et al. (1977), Voropaev (1978), Golubev (1978), it is obvious from an engineering point of view, that the river Neva is also attractive as a source to this irrigation project. With a discharge of approximately 3×10^3 m^3/s, the river Neva is the largest single fresh water contributor to the Baltic Sea, to which the total average fresh water input is, in the order of 15×10^3 m^3/s. Therefore, a radical decrease in the runoff from the river Neva has a great bearing on the hydrography of the Baltic. Further, it has also a large effect on the hydrography of the inland Danish waters, which links the brackish Baltic Sea to the ocean.

A man-made regulation of the river Neva is therefore a matter of international concern, as it will influence all the countries boardering the Baltic. On the other hand, there seems to be no international laws or conventions, which makes it possible for the other Baltic countries to change the decisions if possibly unwanted effects of the regulations can be foreseen. The problem has a parallel in the Danish project for building a bridge across the Great Belt, which was estimated to have a measureable influence on the Baltic Sea, Bo Pedersen (1978). Although there was an international reaction against the building of the bridge, it was for economic reasons, that the Danish government finally decided to postpone the bridge project.

The main objective of the present paper is to establish an estimate of the hydrographic changes in the Baltic Sea and the Danish inland waters if part of the river Neva's discharge is diverged from the Baltic Sea. An evaluation or estimation of the possible consequences for the affected countries is beyond the scope, but it is the hope, that the article will act as a trigger for further discussions, and that the findings will serve as a basis for further work.

The Basic Principles and Assumptions for the Model

In the Baltic Sea and its connections with the North Sea (the Cattegat, the Belts and the Sound) all types of estuaries, i.e. semienclosed bodies of water, where a measurable dilution by fresh water are present, can be recognized. Although a throughout hydrographic description of an estuary demands knowledge of the variation in space and time of all relevant physical properties, such as salinity, temperature, oxygen content, phosphate and nitrate concentrations etc. we shall make a common approach and restrict ourselves to a representative *steady-state* situation considering only the *salinity* distribution, which is the property governing the vertical stability and hence the mixing in the actual case. The most simple representation of an estuary in which the basic physical conditions are maintained is a *two-layer* flow. An inspection of the actual conditions in the Baltic Sea and in the Danish inland waters confirms, that this is a fair approximation.

Our approach is then, first to identify the major external forces affecting the system (fresh water discharge, wind, tide, etc.), then to estimate the correct order of magnitude of the strength of these forces, introduce them in our model and then finally confirm with the actual measured conditions in the estuary, that our model is reasonable representative for the dynamics of the estuary. After the verification of the model, we introduce the change in the fresh water discharge from the river Neva – linearize the equations – and solve with respect to the changes in the salinities, the depths and the flows in the idealized estuaries. In these calculations we have focused on the man-made changes in the fresh water discharge. The consequences for the layer depths and salinities in the Baltic Sea for a natural variation in the fresh water discharge are different from our findings, due to the strong correlation between the precipitation (and hence the runoff) and the meteorological conditions, the last being held unchanged in our calculations.

As stated above all types of estuaries are present in the model. The dynamics of an estuary is mainly affected by the following parameters, Bo Pedersen (1980a)

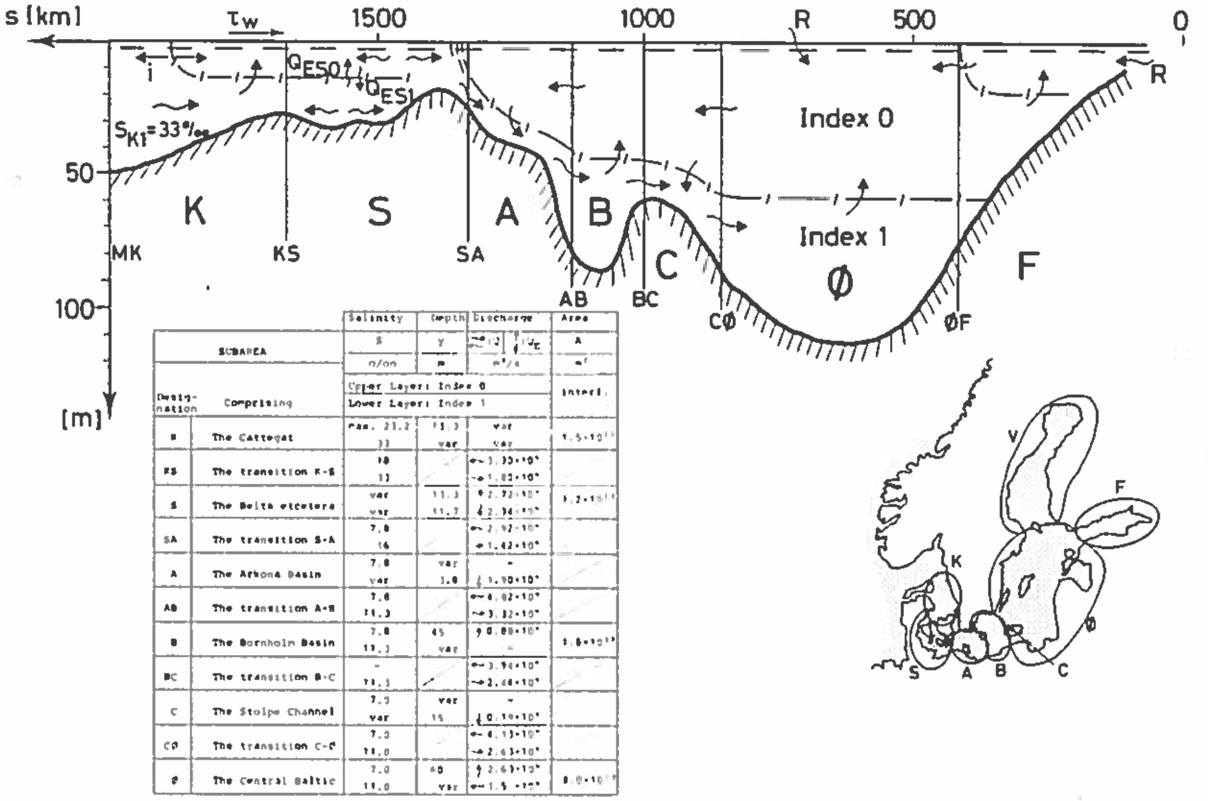
1. The geometry
2. The hydrology of the adjacent watershed
3. The oceanographic conditions outside the estuary
4. The wind field (and the barometric pressure variation due to the large dimensions of the Baltic Sea).

The great variability of these parameters over the actual oceanographic field makes it necessary to divide the total area into eight subareas as indicated in Fig. 1. The subdivision is chosen in such a way that a reasonable simple dynamic description can be given for each region, and hence, the areas do not represent regions of equal importance, merely areas of different dynamic behaviour.

For each subarea steady-state continuity-equations for mass and volume are established. One of the terms of major importance for the continuity equations, is the term representing the mixing accross the interface separating the two layers. This mixing is due to the generation of turbulence by external forces, such as tide, variable meteorological conditions, etc., i.e. all highly non-stationary forces. Therefore, although the basic objective is to establish a steady-state model, it is necessary to incorporate the non-steady dynamic behaviour of the system in the description in order to maintain the correct physics. To transfer a dynamic situation to an artificial steady state demands knowledge of the representative time scale and the representative force scale. With focus on the mixing, a representative averaging time scale is the residence time, T, i.e. a measure of the mean time that a particle of tracer remains inside the actual subarea of the estuary system

$$T = \frac{Vol}{Q} \tag{1}$$

where Vol is the total volume of pure fresh water inside the subarea and Q is the accumulated fresh water discharge at the actual cross section. The residence time for the Baltic estuary system varies from for example, typically a week in the Belt



Designation	SUBAREA Comprising	Salinity S o/oo	Depth Y m	Discharge Q m³/s	Area A m²
		Upper Layer: Index 0 / Lower Layer: Index 1			interf.
K	The Cattegat	max. 21.2 / 33	11.3 / var	var / var	1.5·10¹¹
KS	The transition K-S	18 / 33		← 1.33·10⁴ / → 1.83·10⁴	
S	The Belts etcetera	var / var	11.3 / 11.7	↑ 2.72·10² / ↓ 2.34·10²	1.2·10¹⁰
SA	The transition S-A	7.8 / 16		← 2.47·10⁴ / → 1.42·10⁴	
A	The Arkona Basin	7.8 / var	var / 3.0	– / ↓ 1.90·10⁴	
AB	The transition A-B	7.8 / 11.3		← 4.02·10⁴ / → 3.32·10⁴	
B	The Bornholm Basin	7.8 / 11.3	45 / var	↑ 0.88·10⁴ / –	1.8·10¹⁰
BC	The transition B-C	– / 11.3		← 3.94·10⁴ / → 2.44·10⁴	
C	The Stolpe Channel	7.3 / var	var / 15	– / ↓ 0.10·10⁴	
CØ	The transition C-Ø	7.3 / 11.0		← 4.13·10⁴ / → 2.63·10⁴	
Ø	The Central Baltic	7.0 / 11.0	60 / var	↑ 2.63·10⁴ / ← 1.5·10⁴	8.0·10¹⁰

Fig. 1. The Baltic estuary system divided into eight subareas. The specific hydrodynamic characteristics of the six outermost subareas are summarized in the table.

region, a month in the Cattegat region to 30 years at the central Baltic. As the time scale for the tide (a day) as well as for an average meteorological event (a week) are below the averaging time for the estuary, these two types of external forces can in the time frame be treated as steady, persistent forces, although it may be admitted, that the seasonal variations as for instance in the meteorological activity and in the runoff cannot be incorporated in our theory. On the other hand, the seasonal variations are much weaker than the single events, and can therefore be neglected in the analysis. The other important scale for the mixing is the force scale, i.e. a measure for the energy available for the mixing process. This is the subject of the next chapter.

Mixing in a Two-Layer Stratified Flow

The two-layer stratified flow is characterized by having two nearly homogeneous layers separated by an interface with a sharp density gradient. The mixing between the two layers can be treated as pure (one-way) entrainment if the level of kinetic energy is high in the one layer and negligible in the other layer. If a measurable level of kinetic energy is present in both layers a two-way transport exists, which can be treated either as a combined entrainment/diffusion problem or, as we prefer it, as a double-sided entrainment. A comprehensive analysis of the entrainment functions for a large class of two-layer flows can be found in Bo Pedersen (1980a). The basic assumption for all the flow cases treated there is, that a universal relationship exists between the energy available for the turbulence (i.e. the production with some minor corrections) and the energy gained (potential as well as turbulent kinetic energy) due to the entrained mass. Hence, the characteristic force scale for the mixing, i.e. for the entrainment, can be evaluated by taking a moving average value of the energy input into the system, which by Bo Pedersen (1980a) is shown to be proportional to the mean speed in the layer $|v|$ to the third power. Hence, the proper dynamic transformation from the non-steady to the steady system is done by applying a mean velocity $\tilde{V}$ defined by

$$\tilde{V} = \left(\frac{1}{T} \int_0^T |v|^3 dt \right)^{\frac{1}{3}} \tag{2}$$

The velocity scale in the continuity equation is the simple mean velocity and not the velocity defined by Eq. (2). It is therefore necessary to incorporate a circulation-velocity with no net transport inside some of the regions in order to get dynamic- as well as mass-balance in the simplified systems.

The major external forces producing turbulence in the system are:

1. *The wind,* which generates a flow in the upper layer. A persistent wind acting far from boundaries causes an entrainment velocity V_E which can be evaluated by the following equation, Bo Pedersen (1980a)

$$\frac{V_E}{U_F} = \frac{2.3}{6 + \mathbb{R}i_F} \qquad \mathbb{R}i_F = \frac{\Delta g y}{U_F^2} \tag{3a}$$

where $U_F = (\tau_W/\varrho)^{1/2}$ is the friction velocity in the water due to the windstress τ_W. The bulk Richardson number $\mathbb{R}i_F$ is a measure of the stability of the system as Δ is the non-dimensional density difference between the upper and the lower layer ($\Delta\varrho = \varrho_{lower} - \varrho_{upper}$), g is the acceleration of gravity and y the upper layer depth. All the subareas in the Baltic have rather stable interfaces, i.e. $\mathbb{R}i_F >> 6$, which means that Eq. (3) can be reduced to

$$\frac{V_E}{U_F} \simeq \frac{2.3\, U_F^2}{\Delta g y} \tag{3b}$$

2. *The heating/cooling* process forms during the summertime a stable thermocline. In the winter period it creates an unstable free convection, which erodes the thermo- or halocline.

As shown by Bo Pedersen (1980a) it is only in those parts of the Baltic system, where the halocline is located deep (Bornholm Basin $y \simeq 45$ m, Baltic Proper $y \simeq 60$ m), that a thermocline forms during a pronounced period of the year. The thermocline acts as a lid, which prevents the wind from creating mixing through the halocline – in the actual region during nearly half a year, which has to be taken into account in the dynamical part of the calculations. During the thermocline-free period the free convection plays the minor role in the overall erosion of the halocline. Therefore the only influence from the heating/cooling in our simple model is, that it prevents mixing in the Bornholm Basin and in the Baltic Proper during half a year.

3. *The tide* generates a periodic in and out flow, which can be registrated in the Danish inland waters. On the other hand, the energy input into the system from the tide is sufficiently small to be negligible in the present analysis.

4. *The meteorological activities* over Scandinavia with succeeding low and high pressure acts like a piston on the Baltic Sea. Combined with wind set-up and set-down an oscillating in- and out-flow through the Danish inland waters is generated. In the Cattegat, the Belts and the Sound this means that a large part of the surface and the bottom water is pendling in an out producing turbulent kinetic energy and therefore mixing. The other type of mixing, which shall be considered, occurs in the Arkona region where the pendling only takes place in the surface water. The saline bottom water is trapped in a dense bottom current on the eastern slope of the Darss Sill (16 m depth) in the Great Belt and on the southern slope of the Drogden Sill (8 m depth) in the Sound.

The order of magnitude of the non-steady flow in the Cattegat and the Belts can be evaluated from the discharge measurements performed in the Great Belt, reported by Jacobsen (1980), see Fig. 2. The typical amplitude in the pendling is about $10^5 m^3/s$, which is 10 times the average fresh water outflow through the Great Belt. This ratio between the mass average and the dynamic average velocity demonstrates the presence of a large no net flow circulation.

The circulation induced mixing can be treated as a quasi-steady mixing due to the extreme large ratio between the non-steady period of the circulation (weeks) and the mixing time scale (hours). For a steady-state condition the strength of the circulation induced entrainment to the wind induced entrainment can be shown (Bo Pedersen 1980a) to be equal for a ratio of the dynamic mean velocity $\tilde{V}$ to the wind generated friction velocity U_F of

$$\frac{\tilde{V}}{U_F} \simeq 50 \qquad (4)$$

In the Cattegat a typical high front speed is $V \simeq 0.1$ m/s, while the representative dynamic friction velocity due to the wind is $U_F \simeq 8 \times 10^{-3}$m/s. Hence in Cattegat the circulation contribution to the mixing is only a few per cent of the wind generated mixing and can therefore be neglected.

In the Belts the typical observed velocities are of an order of magnitude which makes them just as important for the mixing process as the wind, i.e. $V \simeq 0.4$ m/s.

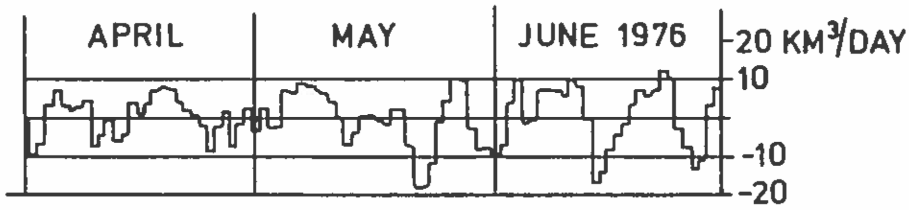


Fig. 2. Typical time series of the measured outwards (positive) and inwards discharge through the Great Belt. From Bo Pedersen (1978).

Fortunately for the present analysis the circulation as well as the wind do both originate from the meteorological activity over Scandinavia, which is kept unchanged in the analysis. The above-mentioned theory considering the ratio between the gain in energy due to entrainment to the production of turbulent kinetic energy simply states for the Belts, that the volume of entrained water Q_{EO} amounts to

$$Q_{E0} = \frac{A \cdot \text{constant}}{(S_{S1} - S_{S0})y_{S0}} \qquad (5)$$

The constant in the numerator stands for the dynamic turbulence production and is estimated below. The denominator represents the gain in potential energy of the entrained mass, namely proportional to the salinity difference (the paranthesis) and the upper layer depth. The high velocities in the non-stationary flow in the Belts creates a downwards as well as an upwards directed entrainment. Again – using the efficiency concept for the mixing – the downwards entrainment is similar to the upwards entrainment discharge

$$Q_{E1} = \frac{A \cdot \text{constant}}{(S_{S1} - S_{S0})y_{S1}} \qquad (6)$$

where the constant stands for the dynamic energy input and y_{S1} is the lower layer depth in the Belts.

The dense bottom current in the Arkona Basin is a highly intermittent flow (Bo Pedersen 1977, Petrén and Walin 1975), which only takes place in connection with an inflow situation to the Baltic. The other dense bottom current in the system – from the Bornholm Basin through the Stolpe Channel into the Baltic proper – is a nearly persistent flow, due to the reservoir effect of the Bornholm Basin (Bo Pedersen 1977, Rydberg 1976). Dense bottom currents in a rotating coordinate system has been treated in Bo Pedersen (1980b). The discharge Q as a function of the distance s along the pathline of the flow is increasing due to entrainment, such that

$$Q(s) = Q(s=0)\ \exp\{0.072 \int_0^s \frac{I\, ds}{y}\} \qquad (7)$$

where I is the local bottom slope. As the depth y is nearly independent of the distance, the integral simply is the drop in elevation non-dimensionalized by the depth. For an unchanged geometry it can furthermore be expected, that the densimetric Froude's number at the sill is unchanged too, see Bo Pedersen (1980a and b), i.e.

$$\left(\frac{Q^2}{\Delta g B^2 y^3}\right)_{\text{before}} = \left(\frac{Q^2}{\Delta g B^2 y^3}\right)_{\text{after}} \tag{8}$$

where the indexes before/after relates to the change of the river Neva discharge into the Baltic Sea.

Characteristics of the Subareas

With the major external forces identified and the associated mixing processes quantified, a description of the individual subareas can be given, including the equations governing the mass and the volume balances and the dynamic behavior. Furthermore, we shall try to a certain extent to verify the simple models presented, against field measurements. To facilitate these assessments a summarizing chart is given, Fig. 1, which contains the symbols used and the values adapted in the present approach to the problem. Finally a summary scheme is given, which contains all the equations needed for the final calculation. The initial values attributed the salinities, the depths and the discharges (see Fig. 1) all satisfy the outlined equations, and are therefore consistent with our interpretation of the dynamics of the system.

Subarea 1: The Cattegat

The western part of the approximately 100 km wide and 240 km long Cattegat has an average depth of approximately 10 m, which is less than the nearly constant upper layer depth in the permanent salt water wedge present in the eastern part, where the total depth is up to 100 m. Hence in the shallow western part a well mixed estuary is normally present, while the eastern part has the characteristics of a two-layer salt water wedge, with constant salinity in the lower layer and varying upper layer salinity.

According to our model, the mixing in the Cattegat is primarily upwards directed entrainment caused by the energy input from the wind. With reference to Fig. 1 the continuity equation for volume states

$$Q_{K0} = Q_{KS0} + \int_0^x V_{EK} B_K \, dx \tag{9}$$

where Q_{K0} is the upper layer net discharge in position x, and Q_{KS0} is the southern boundary value of this discharge (B_K = the interfacial width). The mass deficit flux is constant, because the lower layer salinity is constant and because the mixing is pure upwards directed entrainment. Hence

$$Q_{K0}\Delta_K = Q_{KS0}\Delta_{KS} \tag{10}$$

where Δ_K and Δ_{KS} stands for the non-dimensional mass deficit at position x and at the southern boundary, respectively.

The dynamic conditions in the Cattegat is described by an entrainment function, Eq. (3b) as well as a boundary condition at the northern boundary, where a front is present. In lack of detailed knowledge of the dynamics of this front a common, simple front condition has been used, namely that the densimetric Froude number $\mathbb{F}_\Delta$ at the front is a constant, i.e.

$$\mathbb{F}_\Delta^2 = \frac{Q^2}{\Delta g B^2 y^3} = \text{constant} \tag{11}$$

where the discharge Q stems from the oscillatory flow.

Eqs. (9), (10) and (3b) can be solved to yield the density-deficit distribution in the Cattegat

$$\frac{1}{\Delta_K} = \frac{1}{\Delta_{KS}} \exp\left(\frac{x}{\lambda_K}\right) \tag{12}$$

where the length scale λ_K is determined by

$$\lambda_K = \frac{\Delta_{KS}\, g\, y_{K0} Q_{KS0}}{2.3\, B_K U_F^3} \tag{13}$$

If we introduce the empirical relation between the deficits in the density ϱ and the salinities

$$\rho_1 - \rho_0 \simeq 0.75\,(S_1 - S_0) \tag{14}$$

Eq. (1) can be transformed to

$$S_{K1} - S_{K0} = (S_{K1} - S_{KS0}) \exp\left(\frac{1{,}680 \cdot 10^3 - s}{\lambda_K}\right) \tag{15}$$

where s (in m) is the overall stationing which starts at the head of the Bothnian Bay (at the Neva inlet to the Bay) and end at the Cattegat/Skagerack front where $s = 1{,}920 \times 10^3$ m (s = $1{,}680 \times 10^3$ m corresponds to the Belt/Cattegat transition), and λ_K takes the value 565×10^3 m.

Eq. (15) describes an equivalent steady, yearly averaged salinity-distribution in the upper layer of the Cattegat. This layer is in fact subject to great forwards and backwards movements during the year. Hence Eq. (15) cannot be checked by field measurements before it has been modified slightly. The non-steady movements are reflected in the position of the Cattegat front, which during an inflow situation moves towards the Belts. North to the front the saline Skagerack water is encountered (salinity ≈ 33‰) – south to the front the brackish Cattegat water is present. The front movements have been simulated by Møller (1980) for half a year during 1975 by applying continuity considerations based on the observed in- and outflows through the Belts (Jacobsen 1980), which was estimated to account for 60% of the total flow. From these calculations an intermittency function can be outlined, where the intermittancy *i* is defined as the proportion of the total time in which the front is located north to the actual section. Assuming a Poisson distribution for *i*, we have

$$i = 1 - \exp\left(\frac{s - 1{,}920 \cdot 10^3}{L_K}\right) \quad (16)$$

where $L_K = 40 \times 10^3$ m is the average inwards movement of the front.

In summary: In order to check the outlined salinity distribution with the salinities encountered in the Cattegat, we have to take the front movements into account, which yield an apparent salinity distribution

$$\overline{S_{K0}} = S_{K0}\, i + S_{K1}(1 - i) \quad (17)$$

where $\overline{S_{KO}}$ is the yearly average salinity in the upper layer of the Cattegat. In Fig. 3 is shown the observed and the calculated yearly averaged salinities. The figure confirms the usefulness of the simple description, especially when it is realized that no »curve-fitting« is used in order to obtain agreement.

Subarea 2: The Belts

The hydrodynamics of subarea 2, which comprises the Belts, the Sound and the Kieler Bay are extremely complicated mainly due to the shallowness of the area (depth of 20 m to 30 m) combined with the highly non-steady in- and outflow of huge amounts of stratified brackish water. On an average a two-layer system exists. The existence of a salinity variation in the upper as well as in the lower layer shows that two-way entrainment is present, which agree with the lower layer being dynamical active in this region. A comprehensive description of the hydrography (inclusive some considerations on the dynamic conditions) of subarea 2 can be found in The Belts project (1976), DHI report (1977) Bo Pedersen (1978) and Jacobsen (1980) where further references to the subject are given. Based on these findings (see also Fig. 2) we assume, that a typical flow cyclus representative of the present consequence analysis is as illustrated in Fig. 4, namely a net outwards flow due to the fresh water discharge *R* super-imposed by an oscillatory motion with an amplitude in accordance with the measurements, i.e. approximately a factor of ten times *R*. The amplitude and the frequency of the cyclus is governed by the meteorologic conditions as the forcing and the friction as the damping factors. As demonstrated by Bo Pedersen (1978), the total flow resistance for the upper layer in the Great Belt is at its minimum at the present average

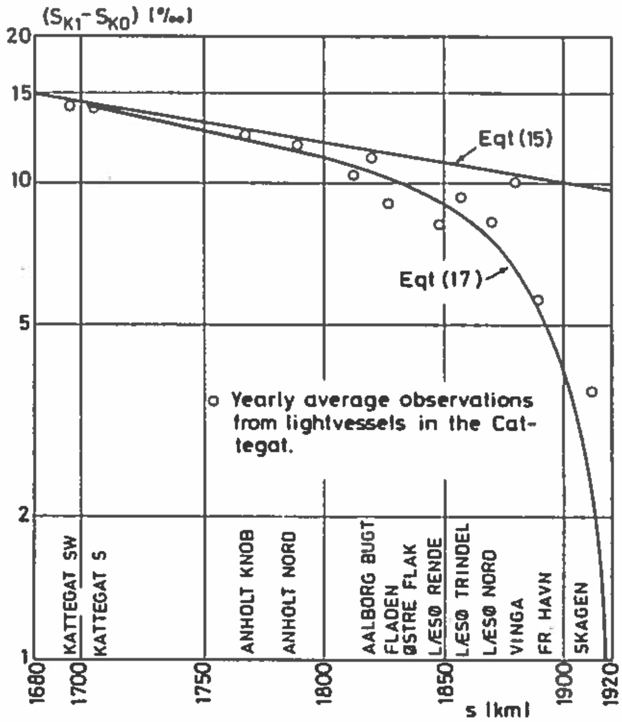


Fig. 3. Calculated and observed yearly average salinity differences between lower(S_{K1}) and upper (S_{K0}) layer in the Cattegat. Eq. (15) illustrates the calculated variation not taking the intermittancy into account (the front movements). Eq. (17) takes the intermittancy into account.

cyclic flow conditions, and hence a minor man-made change in the fresh water discharge will neither create changes in the amplitude nor in the frequency of the pendling discharge, see Fig. 4. Furthermore the minimum condition implies that no change in the production of turbulent kinetic energy occurs.

The sills which separate the Belts and the Sound from the Arkona Basin trap the inwards flowing water which descends as a dense bottom current into the lower layer of the stratified Bornholm Basin, see Fig. 1. If we assume, that the time in which trapping occur is nearly independent of the fresh-water discharge,

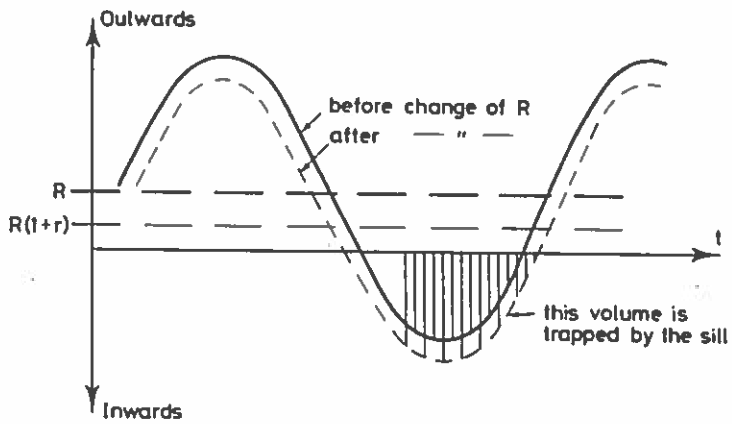


Fig. 4. The inflow/outflow through the Belts schematized by a simple harmonic cyclus superimposed on the fresh-water runoff. The volume trapped by the sills is hatched.

the following simple equation for the sill overflow Q_{SA1} (see Fig. 4) applies

$$Q_{SA1} + R = \mathrm{const}_{SA1} \tag{18}$$

The other equations describing the model outlined are the continuity equations, which are

$$S_{K0}(Q_{ES0} - Q_{ES1} + Q_{SA1} + R) = S_{K1}(Q_{ES0} - Q_{ES1} + Q_{SA1}) \tag{19}$$

stating the no net transport of salt condition at the Belts/Cattegat transition, and

$$S_{SA1}Q_{SA1} = (Q_{ES0} - Q_{ES1} + Q_{SA1})S_{K1} - Q_{ES0}S_{S1} + Q_{ES1}S_{S0} \tag{20}$$

expressing the salt balance for the lower layer in the Belts. The entrainment fluxes (the Q_E's) are as stated in the previous chapter related to the dynamics of the flow, Eqs. (5) and (6). The salinities S_{S1} and S_{S0} in the upper and the lower layer, respectively are for conveniency taken as simple averages of the boundary values, i.e.

$$\begin{aligned} S_{S0} &= 0.5(S_{KS0} + S_{SA0}) \\ S_{S1} &= 0.5(S_{KS1} + S_{SA1}) \end{aligned} \tag{21}$$

The complexity of the flow pattern and of the geometry makes it a rather difficult task to give a convincing verification of the outlined Belt model, but a check can be made by considering some reasonable values for the measured/estimated quantities. Inserting the values shown in Fig. 1 in the governing equations yields

$$\begin{aligned} Q_{ES_0} &= 2.72 \cdot 10^4 \ \mathrm{m^3/s} \\ Q_{ES1} &= 2.18 \cdot 10^4 \ \mathrm{m^3/s} \end{aligned} \tag{22}$$

The upwards directed entrainment Q_{ES0} can be compared with the entrainment one would have had, if it was purely wind generated (see Eq. (3b)). With an interfacial area of approximately 10^{10} m^2 Eq. (3b) gives an upwards entraining flux of 1.2×10^4 m^3/s. The remaining entrainment, 1.5×10^4 m^3/s, can be accounted for by an upper layer dynamic velocity of $\tilde{V} \simeq 0.45$ m/s (see Bo Pedersen 1980a) which is in good agreement with the prevailing flow conditions in the main contributor, the Great Belt. The downward entrainment is created by the turbulence production in the lower layer. The higher resistance the flow experienced at the bottom compared with the interface means, that the dynamic velocity of the lower layer necessary for producing the entrainment is less than that of the upper layer (approximately 2/3 of the upper layer velocity, see Bo Pedersen 1980a), which again agrees well with the observations in the Great Belt. Hence although the model is crude, it reflects the main hydrodynamic behaviour of the complex Belt-area.

Subarea 3: The Arkona Basin

The most characteristic hydrodynamic feature of the Arkona Basin can be observed during an inflow situation, where a dense bottom current is formed along the southern bank of the Basin. Passing the sills in the Great Belt and in the Sound the saline water flows into the Arkona Basin, where it very soon plunge down and descend to the deep water of the Bornholm Basin, i.e. below the interface located at approximately 45 m depth, see Fig. 1. During its course it entrains water from the less saline overlaying water, resulting in a decrease in salinity from 16‰ at the Darss Sill to about 11-12‰, at the merging with the lower layer in the Bornholm Basin. During a subsequent outflow situation the dense bottom current gradually runs dry and finally disappears until the process is repeated by a new inflow. The ambient water in the Arkona Basin is nearly homogeneous (except for the thermocline formation) because the average depth is less than the upper well-mixed layer of the adjacent Bornholm Basin. Hence the flow conditions in the upper layer of the Arkona Basin has no significant influence on the mixing of salt in the Baltic, and the salinity can therefore be taken as equal to the value present at the transition to the Bornholm Basin, namely 7.8‰.

The sparse measurements of the dense bottom current (Petrén et al. 1975), indicates an order of magnitude of typically 5 meter of the thickness of this current when present. If we apply Eq. (7) outlined in the previous chapter on the dense bottom current and takes the drop in elevation equal to the vertical distance from the plunge line to the interface in the Bornholm Basin (y_{B0} = 45 m) we get an average depth of y_{A1} = 3.8 m, in reasonable agreement with the observations. In these calculations we have applied the constant mass deficit flux condition

$$\Delta_{SA} Q_{SA1} = \Delta_B Q_{B1} \tag{23}$$

where indexes $SA1$ and $B1$ stand for the lower layers at the Darss Sill and at the Bornholm Basin respectively. A further condition necessary for the later calculations is Eq. (8) expressing the assumed constancy of the densimetric Froude number at the sill.

Subarea 4: The Bornholm Basin

The Bornholm Basin is a 80 to 100 m deep basin separated from the central Baltic Sea by a shallow sill with a narrow local depression to 60 m depth at the Stolpe Channel. The intermittent flow of dense water from the Arkona Basin is smoothed out and continues through the Stolpe Channel. Hence the lower layer in the Bornholm Basin only acts as a buffer for the discontinous inflow, with a retention time of say 10 weeks (Bo Pedersen 1977), while there is very little downwards directed mixing.

The position of the interface is determined by the rate of outflow through the Stolpe Channel. This flow is assumed to be governed by a constant densimetric Froude number, see Eq. (8). Hence the interfacial depth in the Bornholm Basin is related to the depth y_{C1} of the dense bottom current present in the Stolpe Channel and to the sill depth (= 60 m) in the following way

$$y_{B0} = 60(\mathrm{m}) - y_{C1} \tag{24}$$

The upper layer is exposed to the wind and the heating/cooling sequence, which as stated in the previous chapter gives rise to an upwards directed entrainment through the halocline during half of the year. Hence Eq. (3b) shall be modified to

$$\frac{V_E}{U_F} = 0.5 \frac{2.3 \; U_F^2}{\Delta g \; y_{B0}} \tag{25}$$

which yields a total discharge through the halocline of

$$Q_{EB0} = 1.5 \frac{U_F^3 A_B}{g(S_{B1} - S_{B0}) y_{B0}} \tag{26}$$

Inserting $U_F \simeq 8 \times 10^{-3}$ m/s yields $Q_{EB} = 8.9 \times 10^3$ m^3/s in yearly average. The upper layer salinity S_{B0} can be found by applying the combined equation for transport of salt and volume through the Bornholm/Arkona section

$$S_{B0}(Q_{B1} + R) = S_{B1} Q_{B1} \tag{27}$$

A check on the calculated entrainment Q_{EB0} is that it is compatible with a salinity of $S_{\emptyset 0} = 7.0‰$ in the upper layer of the Central Baltic, in accordance with observations.

Subarea 5: The Stolpe Channel

Surbarea 5 has the same hydrodynamic status as subarea 3, the Arkona Basin, the only significant difference being, that the dense bottom current here is nearly steady. The continuity of volume yields an upstream discharge of

$$Q_{BC1} = Q_{B1} - Q_{EB0} \tag{28}$$

i.e. the bottom water inflow to the Bornholm Basin Q_{B1} reduced by the upwards directed entrainment Q_{EB0}.

The salinity has accordingly decreased slightly, confer with continuity equation for mass or salt deficit

$$Q_{BC1}(S_{B1} - S_{\emptyset 0}) = Q_{\emptyset 1}(S_{\emptyset 1} - S_{\emptyset 0}) \tag{29}$$

which yields the salinity of inflowing bottom water to the Baltic Proper $S_{\emptyset 1} = 11‰$ in agreement with observations.

Similar to the Arkona bottom current the depth of the Stolpe Channel current is assumed to be governed by a constant densimetric Froude number at the head of the current.

Subarea 6: The Central Baltic

The Central Baltic Basin is the largest of the subareas and furthermore the deepest (up to about 400 m). The main inflows of fresh water to the system takes place here, comprising direct river runoff (31%), contribution from the Finnish Bay (27%) and from the Bothnian Bay (42%).

The nearly persistent brackish water flow from the Stolpe Channel descends to below the primary halocline, located at approximately 60 m depth, where it spreads out at the density-matching level. The continuity of the volume demands an upwards directed entrainment $Q_{E\emptyset}$, caused by the energy input from the wind and the cooling, both of which are kept unchanged in the present analysis. The combined effects of all the external forces can be integrated to a single representative dynamic friction velocity $U_{F\emptyset}$, yielding an entrainment of (see Eq. (26))

$$Q_{E\emptyset} = Q_{\emptyset 1} = \frac{1.5\, A_{\emptyset}\, U_{F\emptyset}^{3}}{(S_{\emptyset 1} - S_{\emptyset 0})\, g\, y_{\emptyset 0}} \tag{30}$$

Inserting the observed values gives a dynamic friction velocity of $U_{F\emptyset} \simeq 8 \times 10^{-3}$ m/s. Although this velocity compares well with the observed wind velocity, one has to remember that $U_{F\emptyset}$ contains the integrated effects of all the external forces and hence may be difficult to evaluate exactly, but as demonstrated, the order of magnitude is correct.

On an average, the upper layer of the Central Baltic is nearly homogeneous due to the multi-directed wind-driven circulations and the long retention time. The salinity is determined by the continuity equation for salt which states

$$S_{\emptyset 0}(R + Q_{\emptyset 1}) = S_{\emptyset 1} Q_{\emptyset 1} \tag{31}$$

Subareas 7 and 8: The Finnish Bay and The Bothnian Bay

The salinity distribution and the position of the interface in the two subareas are governed by the boundary conditions, which, besides the external forces, are the fresh water input and the conditions in the Central Baltic, respectively. A change in these boundary conditions will have a measurable effect on the hydrography of both Bays' – but it will not influence the conditions in the Central Baltic – besides the changes caused by the change in the fresh-water input. Hence, as we are mainly concerned with the Central Baltic and the Danish Inland waters, a detailed analysis or modelling of the Finnish Bay and the Bothnian Bay is omitted. Another reason for not taking the two Bays into account is, that major changes may take place in the Finnish Bay which make the linearized approach doubtful.

The Consequence Analysis

The equations outlined in the previous chapters give an overall quantitative description of the Baltic estuary system, as it behaves under the present average meteorologic and hydrographic conditions. A man-made change in the runoff from the river Neva has of course a direct influence on the total fresh-water inflow to the Baltic, while it is unlikely to have any significant influence on the meteorological conditions and hence on the external forces. Therefore, the conditions before and after the river diversion both satisfies the outlined governing equations. Furthermore, the changes are relatively small, which suggests a linearization of the equations. The procedure is to introduce the new parameters as the old ones plus a minor correction, as for instance

$$\begin{aligned} R_{new} &= R(1+r) \\ Q_{new} &= Q(1+q) \\ S_{new} &= S(1+s) \\ y_{new} &= y(1+\eta) \\ \lambda_{new} &= \lambda(1+\ell) \end{aligned} \qquad (32)$$

where r, q, s, η and λ all are small dimensionless quantities. By use of a Taylor expansion in which only the first order terms are retained, a set of linear equations in the correction terms is obtained, which can be solved directly. The original as well as the linearized equations are for conveniency summarized in Table 1, where the solution to the set of equations is shown as well. In the linearized equations due respect to the changes in the interfacial widths or areas with depth have been taken.

All the corrections have been related to the fresh-water discharge reduction r in the data output. Hence – as an example – if the fresh-water diversion from the river Neva amounts to 5 per cent of the total fresh-water inflow to the Baltic (i.e. approximately 25 per cent reduction of the river Neva's discharge), then $r = -0.05$. By use of Table 1 the associated changes in the salinities, depths and discharges can be evaluated. Some of the calculated changes can be compared with the natural variations encountered in the Baltic estuary system during this century. We have chosen to illustrate this variability by plotting the 10-year sliding mean values of the runoff R from river Vuoksi, the salinity $S_{\emptyset 0}$ and the depth $y_{\emptyset 0}$ in the upper layer in the Central Baltic and finally the upper layer salinity in the Cattegat region, see Fig. 5. Two comments to this illustration are appropriate. First, the natural variations are caused by the combined effects of a variability in the runoff and in the climate, the last one being held unchanged in our calculations. Second, the natural variations are highly non-steady, while our analysis deals with the steady state. In the non-steady case, the reservoir effect damps the amplitudes of a cyclic variation (as for instance in the salinity) compared to the long-term response of a step function and furthermore the output is delayed compared to the time

TABLE 1

EQT. No	THE GOVERNING EQUATIONS	REGION	THE LINEARIZED EQUATIONS	SOLUTION
14	$B^2 y_{K0}^3(33\cdot10^{-3}-S_{MK0}) = 9.44\cdot10^{10}$		$1.80\eta_{K0} -2.37s_{MK0} = 0$	$\eta_{K0} = -0.36\,r$
18	$(33\cdot10^{-3}-S_{MK0}) = (33\cdot10^{-3}-S_{KS0})\exp(-\frac{240\cdot10^3 m}{\lambda_K})$		$-2.37s_{MK0}+1.20s_{KS0}-0.43\ell_K = 0$	$s_{MK0} = -0.27\,r$
16	$\lambda_K = 1.81\cdot10^5\,\frac{R\,y_{K0}}{B_K}$	K	$-\ell_K + 1.60\eta_K = -r$	$s_{KS0} = -0.39\,r$
	$B_K = 64\cdot10^3(1-0.6\eta_{K0})$			$\ell_K = 0.42\,r$
21	$Q_{SA1}+R = 2.92\cdot10^4[m^3/s]$		$-1.42q_{SA1} = 1.50\,r$	$q_{SA1} = -1.06\,r$
22	$S_{KS0}(Q_{ES0}-Q_{ES1}+Q_{SA1}+R) =$ $33\cdot10^{-3}(Q_{ES0}-Q_{ES1}+Q_{SA1})$		$s_{KS0}-0.69q_{ES0}+0.59q_{ES1}-0.36q_{SA1} =$ $-0.46\,r$	$q_{ES0} = -0.03\,r$ $q_{ES1} = -0.81\,r$
23	$S_{SA1}Q_{SA1} = (Q_{ES0}-Q_{ES1}+Q_{SA1})33\cdot10^{-3}$ $-Q_{ES0}S_{S1}+Q_{ES1}S_{S0}$	S	$s_{SA1}-1.06q_{SA1}-1.07q_{ES0}+2.07q_{ES1}$ $+2.93s_{S1}-1.33s_{S0} = 0$	$s_{SA1} = 0.04\,r$ $s_{S1} = 0.01\,r$
24	$S_{S0} = 0.5(S_{KS0}+S_{B0})$ $S_{S1} = 0.5(33\cdot10^{-3}+S_{SA1})$		$s_{S0}-0.70s_{KS0}-0.30s_{B0} = 0$ $s_{S1}+0.33s_{SA1} = 0$	$s_{S0} = +0.33\,r$
8	$Q_{ES0} = \frac{4.20\cdot10^3}{(S_{S1}-S_{S0})y_{K0}}$		$q_{ES0}+1.0\eta_{K0}+2.11s_{S1}-1.11s_{S0} = 0$	
9	$Q_{ES1} = \frac{3.22\cdot10^3}{(S_{S1}-S_{S0})(25-y_{K0})}$		$q_{ES1}-1.14\eta_{K0}+2.11s_{S1}-1.11s_{S0} = 0$	
26	$(S_{SA1}-S_{B0})Q_{SA1} = (S_{B1}-S_{B0})Q_{B1}$		$1.95s_{SA1}+1.28s_{B0}+q_{SA1}-q_{B1}-3.23s_{B1} = 0$	
10	$Q_{B1} = Q_{SA1}\exp\{0.072\,\frac{y_{B0}}{y_{A1}}\}$	A	$q_{B1}-q_{SA1}+0.85\eta_{A1}-0.85\eta_{B0} = 0$	$\eta_{A1} = -0.79\,r$
11	$\frac{Q_{SA1}^2}{(S_{SA1}-S_{B0})y_{A1}^3} = 4.48\cdot10^8$		$2q_{SA1}-1.95_{SA1}+0.95s_{B0}-3\eta_{A1} = 0$	
27	$y_{B0} = 60[m]-y_{C1}$		$3\eta_{B0}+\eta_{C1} = 0$	$s_{B0} = 0.16\,r$
29	$Q_{EB0} = \frac{7.79\cdot10^{-8}A_B}{(S_{B1}-S_{B0})y_{B0}}$	B	$q_{EB0}+1.60\eta_{B0}+3.23s_{B1}-2.23s_{B0} = 0$	$s_{B1} = -0.31\,r$ $\eta_{B0} = 0.16\,r$
30	$S_{B0}(Q_{B1}+R) = S_{B1}Q_{B1}$ $A_B = 1.8\cdot10^{10}(1-0{,}6\eta_{B0})$		$s_{B0}-s_{B1}+1.69q_{B1} = -0.31\,r$	$q_{B1} = -0.25\,r$ $q_{EB0} = 0.29\,r$
31	$Q_{BC1} = Q_{B1}-Q_{EB0}$		$q_{BC1}-1.36q_{B1}+0.36q_{EB0} = 0$	$\eta_{C1} = -0.48\,r$
32	$Q_{BC1}(S_{B1}-S_{\emptyset 0}) = Q_{\emptyset 1}(S_{\emptyset 1}-S_{\emptyset 0})$		$q_{BC1}-q_{\emptyset 1}+2.63s_{B1}+0.12s_{\emptyset 0}-2.75s_{\emptyset 1} = 0$	$q_{BC1} = -0.44\,r$
11	$\frac{Q_{BC1}^2}{(S_{B1}-S_{\emptyset 0})y_{C1}^3} = 4.10\cdot10^7$	C	$2q_{BC1}-2.63s_{B1}+1.63s_{\emptyset 0}-3\eta_{C1} = 0$	
10	$Q_{\emptyset 1} = Q_{BC1}\exp\{0.072\,\frac{y_{\emptyset 0}-y_{B0}}{y_{C1}}\}$		$q_{\emptyset 1}-q_{BC1}-0.29\eta_{\emptyset 0}+0.22\eta_{B0}+0.07\eta_{C1} = 0$	
33	$Q_{\emptyset 1} = \frac{7.89\cdot10^{-8}A_{\emptyset}}{(S_{\emptyset 1}-S_{\emptyset 0})y_{\emptyset 0}}$		$q_{\emptyset 1}+2.1\eta_{\emptyset 0}+2.75s_{\emptyset 1}-1.75s_{\emptyset 0} = 0$	$q_{\emptyset 1} = -0.46\,r$
34	$S_{\emptyset 0}(R+Q_{\emptyset 1}) = S_{\emptyset 1}Q_{\emptyset 1}$		$s_{\emptyset 0}-0.36q_{\emptyset 1}-s_{\emptyset 1} = -0.36\,r$	$s_{\emptyset 0} = -0.85\,r$
	$R = 1.5\cdot10^4 m^3/s$	Ø		$s_{\emptyset 1} = -0.33\,r$
	$A_{\emptyset} = 8\cdot10^{10}(1-1.1\eta_{\emptyset 0})$			$\eta_{\emptyset 0} = -0.06\,r$

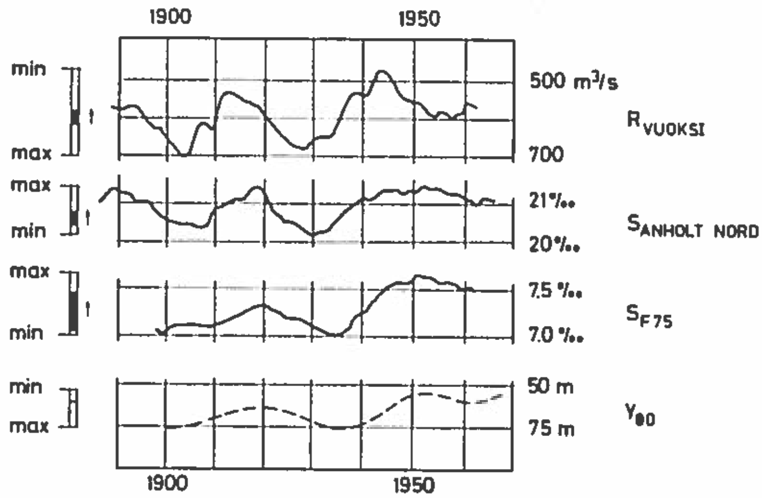


Fig. 5. The secular changes in the Baltic Sea estuary system illustrated by the 10-year sliding mean of the runoff from river Vuoksi (Nilsson and Svansson 1974), the surface salinity at Anholt Nord (Nilsson and Svansson 1974), the upper layer salinity at station F75 (the Central Baltic, Hela 1966), the upper layer depth in the Central Baltic (Fonselius 1969).
In the column diagram to the left is shown the min/max values observed and the changes calculated for a 25 per cent reduction of the river Neva's runoff.

for the input. The time delaying effect is demonstrated in the salinity observed at station F75, which shows a response time of about 10 years, compatible with the retention time in the upper layer of the Central Baltic. A throughout discussion of the observed data shall not be given, mostly due to lack of knowledge of the variability in the meteorologic forcing. Instead we shall compare and comment upon the natural and man-made variations as observed and calculated, respectively.

The man-made reduction in our example is 5% of the total runoff to the Baltic. This reduction is an order of magnitude less than the natural variations encountered during this century as represented by the runoff of river Vuoksi, see Fig. 5.

The most pronounced influence of a man-made reduction in the runoff is to be found in the upper layer salinity in the Central Baltic, see Fig. 5, S_{F75}, where the calculated salinity change is approximately half the variation observed during this century. The observed salinity variations reflects the non-steady input from fresh water (R) and wind (U_F), and are therefore highly damped by the reservoir effect, as compared to the values one would have had if a permanent change in R and U_F had occurred. Moreover the observed salinity is highly influenced by the change in the meteorological conditions, which are not taken into account in our calculations. Therefore, the relatively strong influence of a man-made change in the runoff is understandable.

The change in salinity of the upper layer of the Cattegat light-vessel Anholt Nord is only approximately 20% of the natural maximum variation observed. In fact the 10-year sliding mean is a bad representation of the salinities in the Cattegat region because the retention time is orders of magnitude less. Compared to the more representative month sliding mean the influence is reduced to less than say 5% of the natural variations.

Finally we found a negligible change in the upper layer depth of the Central Baltic. The actual maximum change has been about 20%. Hence the position of the interface in the Central Baltic is primarily determined by the changes in the meteorological conditions, which have an influence directly in determining the rate of entrainment and indirectly in determining the sill overflow and hence the discharge into the lower layer.

Conclusion

The possible influence on the hydrography of the Baltic Sea estuary system subject to a man-made change in the river runoff, has been investigated. The density stratification, the geometry and the variability of the external forces makes the estuary a rather complex hydrodynamic system and hence it was necessary to divide it into eight subareas, each of which described by its own balance in mass, volume and dynamic (mixing). It has been demonstrated that the prevailing meteorological conditions over the area play the dominating role in the formation of fronts, salt water wedges, dense bottom currents and all the other types of density currents encountered in the system and in the mixing processes. These meteorological conditions are kept unchanged in the present consequence analysis, which has been performed by linearizing the governing equations with respect to the minor changes caused by the man-made change in the river Neva's runoff. The changes in the depths, discharges and salinities in the system have been presented as functions of the change in the fresh-water discharge. Although it is not the intention of this paper to reach any conclusions concerning the possible positive or negative consequences of the river diversion it shall be emphasized, that the dynamic stability of the estuary system will be reduced, which makes the system more sensitive to the inevitable changes in the external forces, i.e. in the natural variations in the meteorology. This reduced stability is caused by the combined effects of an increased salinity and a decreased upper layer depth.

The salinity increase in the upper layer of the Central Baltic is remarkably large compared to the natural variations encountered during this century. The reason for this is the reservoir effect, which highly damps a cyclic variation (the natural) but not a step-variation (the man-made).

References

Bo Pedersen, Fl. (1977) On dense bottom currents in the Baltic Deep Water. *Nordic Hydrology,* 8, 297-316.

Bo Pedersen, Fl. (1978) On the influence of a bridge across the Great Belt on the hydrography of the Baltic Sea, 11th conference of the Baltic Oceanographers, Rostock, DDR.

Bo Pedersen, Fl. (1980a) A monograph on turbulent entrainment and friction in two-layer stratified flow. Series Paper No. 25, Inst. of Hydrodynamics and Hydraulic Eng., Tech. Univ. of Denmark.

Bo Pedersen, Fl. (1980b) Dense bottom currents in Rotating Ocean. American Society of Civil Engineers, *Proc. Vol. 106, Hy 8,* 1291-1308.

DHI-Report (1977) Bæltprojektet. Matematiske modeller af Store Bælt og Øresund – Slutrapport. Dansk Hydraulisk Inst. DK-2970 Hørsholm.

Fonselius, S. H. (1969) Hydrography of the Baltic Deep Basins III, Fishery Board of Sweden, Series Hydrography, Report No. 23.

Golubev, G. (1978) Environmental Issue of Large Interregional Water Transfer Projects. *Water Supply and Management, Vol. 2,* 177-185.

Hela, I. (1966) Secular changes in the salinity of the upper waters of the Northern Baltic Sea. *Commentationes Physico-Mathematicae, Vol. 31,* Nr. 14, 1966.

Jacobsen, T. (1980) Sea Water exchange of the Baltic. Measurements and methods. Preprint of Dr. thesis. Inst. of Physical Oceanography. Univ. of Copenhagen.

Mikhaylov, N. I., Nikolayev, V. A., and Timashev, I. Ye. (1977) Environmental Protection Issue and Southward Diversion of Siberian Rivers. English Translation from: Vestnik Moskovskog Universiteta, Geografiya (1977), No. 5, pp. 50-56.

Møller, Jacob Steen (1980) Østersøens Hydrografi. Internal Report in Danish. Inst. of Hydrodynamics and Hydraulic Eng. Tech. Univ. of Denmark.

Nilsson, H., and Svansson, A. (1974) Long term variations of Oceanographic Parameters in the Baltic and Adjacent Waters. Meddelande från Havsfiske-laboratoriet, Lysekil, nr. 174.

Petrén, O., and Walin, G. (1975) Some observations of the deep flow in the Bornholm strait during the period June 73 – December 74, Rep. No. 12, Inst. of Oceanography, Univ. of Gothenburg, Sweden.

Rydberg, L. (1976) Observations of the deep water flow through the Stolpe Channel during August 1976. Rep. No. 15. Inst. of Oceanography, Univ. of Gothenburg, Sweden.

The Beltproject (1976) Interim report on the Danish Belt project (In Danish). Publ. by Miljøstyrelsen, Kampmannsgade 1, DK-1604 Copenhagen.

Voropaev, G. V. (1978) The Scientific Principles of Large-Scale Areal Redistribution of Water Resources in the USSR. *Water Supply and Management, Vol. 2,* pp. 91-101.

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Address:
Institute of hydrodynamics and hydraulic engineering,
ISVA,
Technical University of Denmark,
Building 115,
DK-2800 Lyngby, Denmark.

FLERLAGSMODELLER I HYDROGRAFIEN.
STYRKER OG SVAGHEDER VED EKSISTERENDE MODELLER

v. Jacob Steen Møller, DHI

20. marts, 1987
JSM/be/4001/N6

<u>Noter til:</u>

Møde i Dansk Vandbygningsteknisk Selskab,
mandag den 16. marts, 1987:

Flerlagsmodeller i hydrografien.
Styrker og svagheder ved eksisterende modeller.

Jacob Steen Møller, civ.ing. lic.tech.
Dansk Hydraulisk Institut

FYSISKE MODELFORSØG

Dansk Hydraulisk Institut (ATV) og Lic-engineering gennemfører en forsøgsserie for Ministeriet for Offentlige Arbejder. Forsøgene gennemføres på Instituttet for Strømningsmekanik og Vandbygning, Danmarks Tekniske Højskole.

Undersøgelsen tjener to formål, - for det første, at dokumentere et beregningsgrundlag, som kan forudsige effekterne af en Storebæltsforbindelse på hyppigheden af interne hydrauliske spring - og for det andet, at undersøge virkningen af lokale konstruktionselementer, som tunnel og bropiller, på den lokale blandingseffektivitet.

I fig. 1 er gengivet, fra præsentationen den 16. marts, principskitser for forsøgene med kritisk strømning.

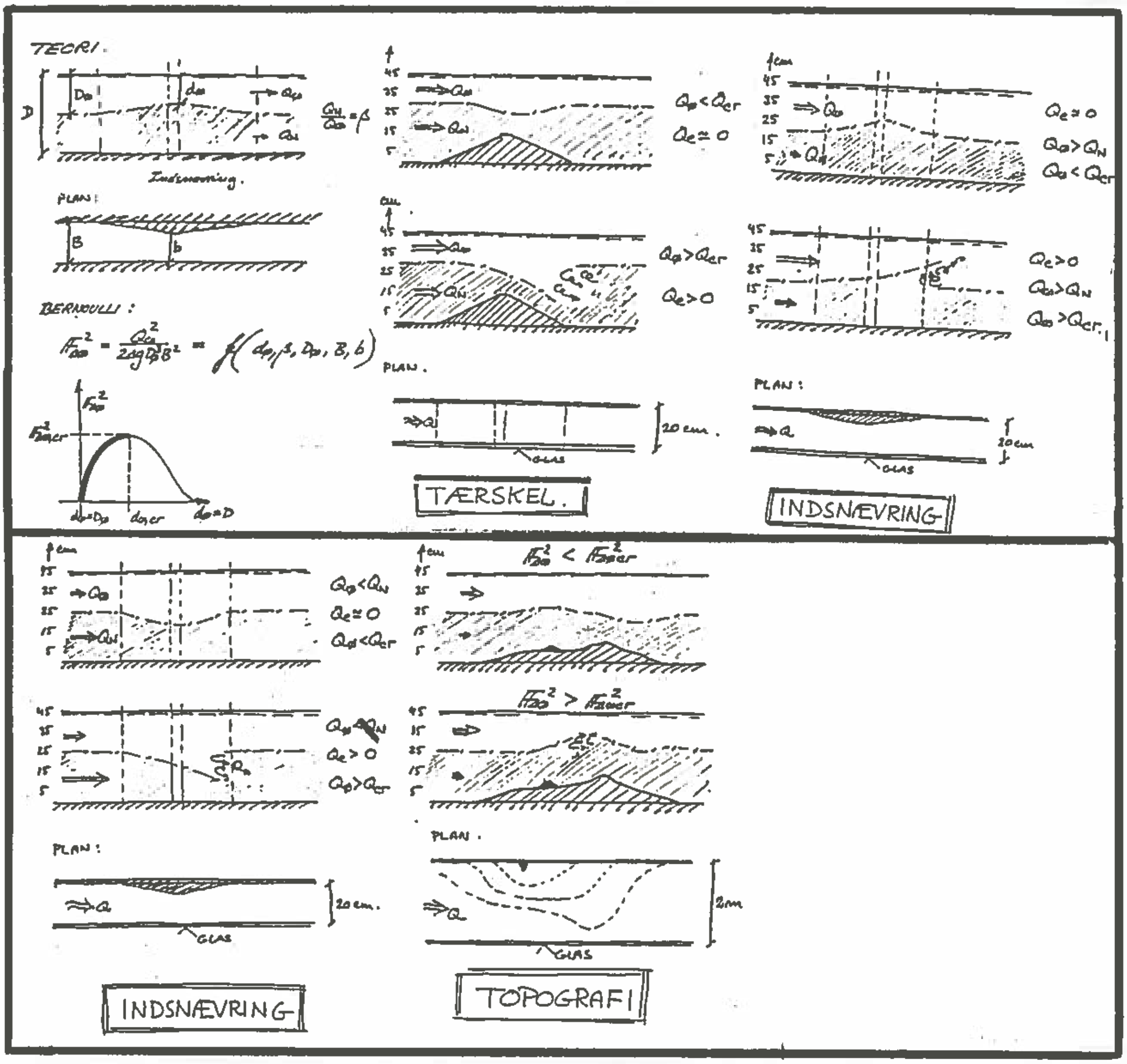


Fig. 1 Overhead-plancher fra præsentation af modelforsøg.

Der gennemføres en række modelforsøg med indsnævring og tærskel efterfulgt af en række forsøg med en principiel topografi. Den principielle topografi udviser samme arealvariation langs strømretningen, som findes i Østerrenden, se fig. 2.

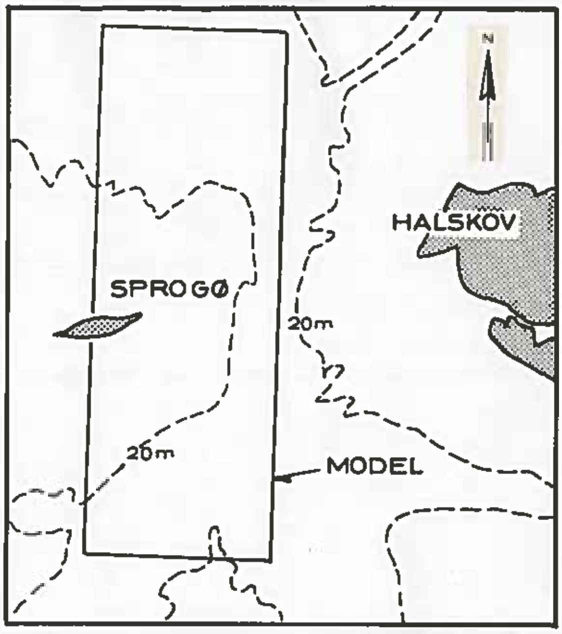


Fig. 2 Modeludsnit af Østerrenden til 'topografiske' modelforsøg.

MATEMATISKE MODELLER

INDLEDNING

I de tidligere foredrag har interessen koncentreret sig om Bælthavet. I fig, 3 vises et eksempel på en udpræget 2-lagsstrømning i Bosporusstrædet. I Sortehavet har sovjettiske kunstvandingsprojekter betydet en markant forøgelse af saltholdigheden af Sortehavets overfladevand.

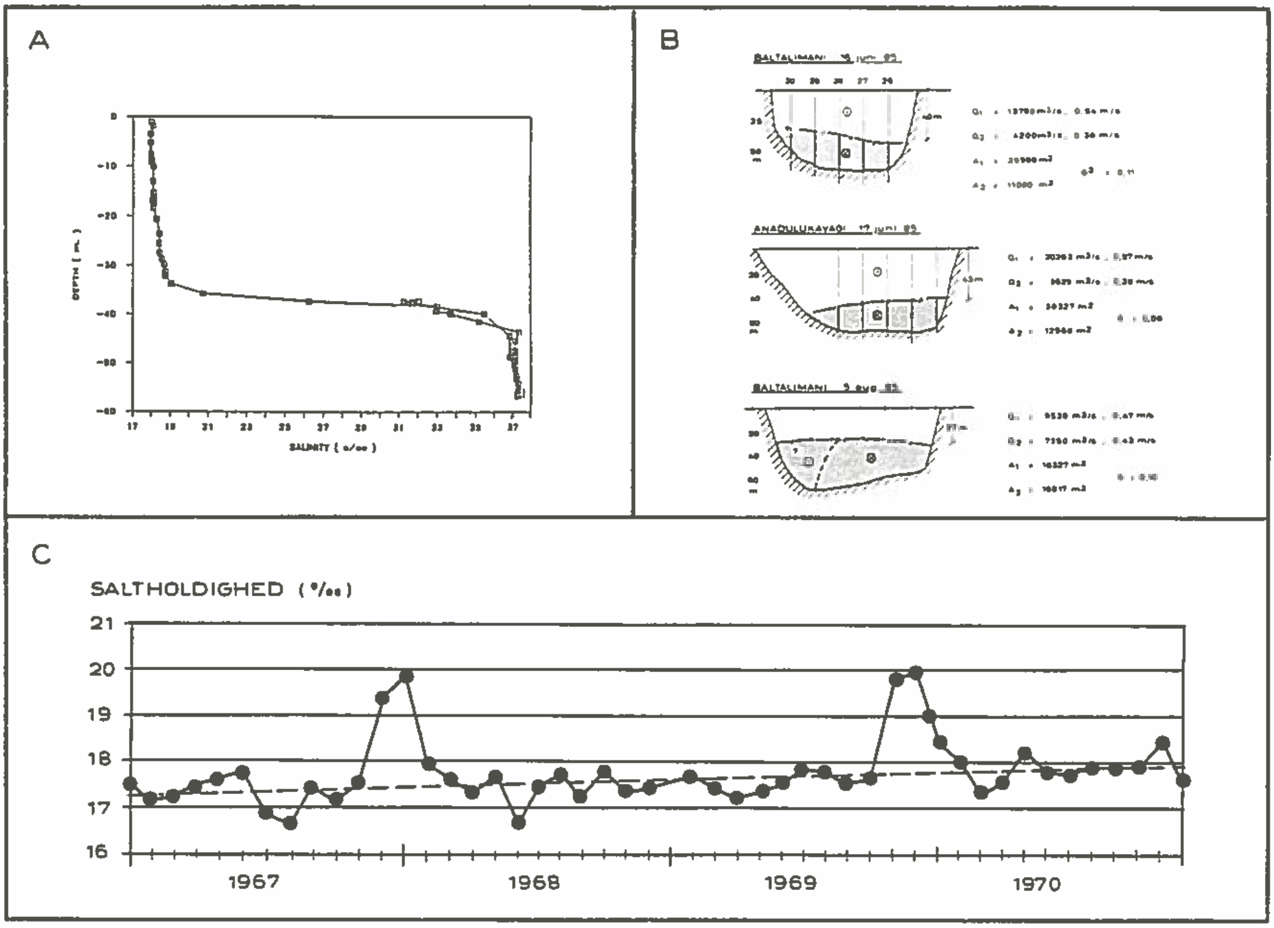
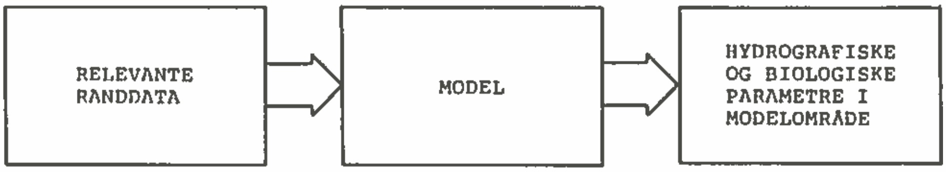


Fig. 3 A og B: Saltholdighedsprofil. Vandføringsmålinger i Bosporusstrædet. C: Saltholdighedens udvikling i overfladevandet.

DEN IDEELLE MODEL

Det er af ressourcemæssige og faglige grunde umuligt at etablere den ideelle model, fig. 4. Derfor må man i hvert enkelt tilfælde analysere, hvilke fysiske og biologiske processer, der er dominerende og vælge en passende forenklet model.

"DEN IDEELLE MODEL"



FX: LUFTTRYK	STRØMNINGSLIG-NINGERNE (HD-LIGN)	VANDSTAND
VINDFELT	TRANSPORT-DISPERSIONSLIGNINGEN (TD-LIGN)	STRØM
VANDSTAND	BIOLOGISKE PROCESUDTRYK (WQ-LIGN)	LAGDELING
SALTHOLDIGHED	ETC.	SALTFORDELING
TEMPERATUR		TEMPERATUR
UDLEDNINGER		BIOLOGI
		FISK
		ETC.

Fig. 4 "Den idelle model".

EKSEMPLER PÅ MODELANVENDELSER

Igennem de seneste 10 år er der gennemført en række matematiske modelundersøgelser i Danmark. I fig. 5 er vist eksempler på disse modeller. Modellerne baserer sig på løsning af de hydrodynamiske ligninger (HD-ligningerne), d.v.s., impulsligningen, transport dispersionsligningen (TD-ligningerne) og en matematisk beskrivelse af de biologiske processer (WQ-ligninger). Til disse ligningssæt knytter der sig oftest størst usikkerhed til formuleringen af WQ-ligningerne.

FYSISK/BIOLOGISK PROBLEM	MODEL	METODE	ANVENDELSER
VANDSTAND (STRØM)	S21 (DHI)	NUMERISK LØS-NING AF HD-LIGNINGER	KYSTINSPEKTORAT NORDSØ/ØSTERSØ 1973-87
MIDDELSALT-HOLDIGHED I ØSTERSØ	ØSTERSØ-MODELLEN (LIC)	STATIONÆR, LINEÆR MODEL	EFFEKT AF KUNST-VANDING/STORE-BÆLTSBRO 1980-87
EUTROFIERING LOKALOMRÅDE	S22 (DHI) WQ-MODEL (VKI)	NUMERISK LØSNING AF HD/TD og WQ LIGNINGERNE	KØLEVANDSUNDER-SØGELSER NOVO-KALUNDBORG 1976-87
EUTROFIERING KATTEGAT	KATTEGATMODEL (LIC) WQ-MODEL (VKI)	SEMIDYNAMISK LØSNING AF HD/TD LIGNINGERNE DYNAMISK LØSNING AF WQ-LIGNING	MILJØSTYRELSENS NPO-RAPPORT EFFEKT AF STORE-BÆLTSPROJEKT 1983 og 1986
EUTROFIERING BÆLTHAVET	ANALYTISK LØSNING (DHI)	SIMPEL MASSE-BALANCE OG ILT-FORBRUGSMODEL	KATTEGAT OG BÆLTHAVET EUTROFIERINGS-DEBAT 1987
"HYDROGRAFISK MODEL"	3-D MODEL (DHI)	NUMERISK LØSNING AF HD/TD/WQ LIGNINGERNE	M.M. 1990 --

Fig. 5 Eksempler på modelanvendelser.

<u>EKSEMPEL 1</u>

EN SIMPEL ILT MODEL.

I en artikel af Schrøder og Møller, som vedlægges, er gennemført en modellering af iltkoncentrationsudviklingen i Kattegats bundvand.

<u>EKSEMPEL 2</u>

S-22 JAMMERLAND BUGT.

For Jammerland bugt har DHI og Vandkvalitetsinstituttet i 1985 gennemført modelberegninger af effekten af udledning af store mængder kvælstof.

Modelarbejdet blev baseret på DHI's System 22, som er en dynamisk to-dimensional model, som løser strømnings- og transportligningerne for lagdelte vandområder. De styrende ligninger er integreret over lagtykkelsen af hvert enkelt lag og løses ved den endelige differensers metode. I fig. 6 er vist en principskitse for modellen.

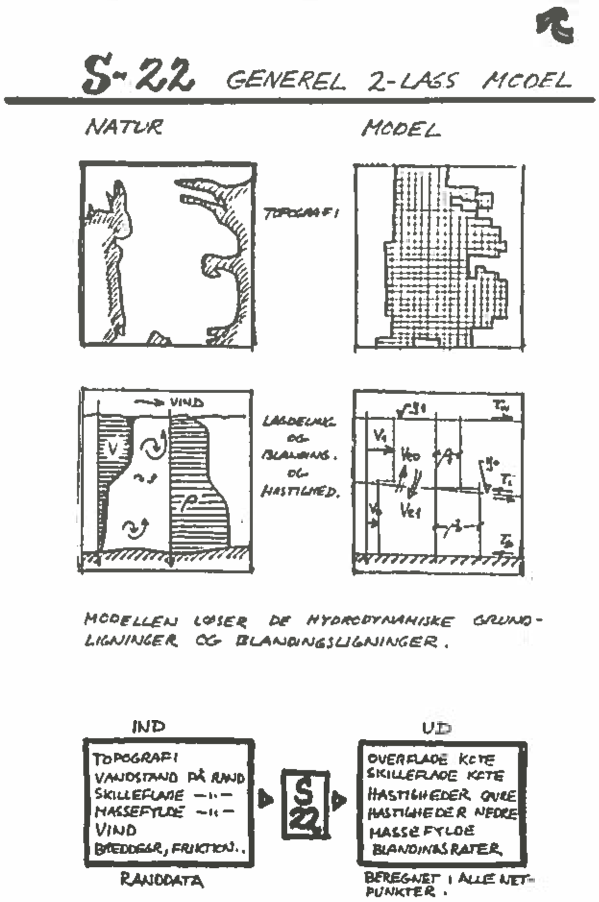
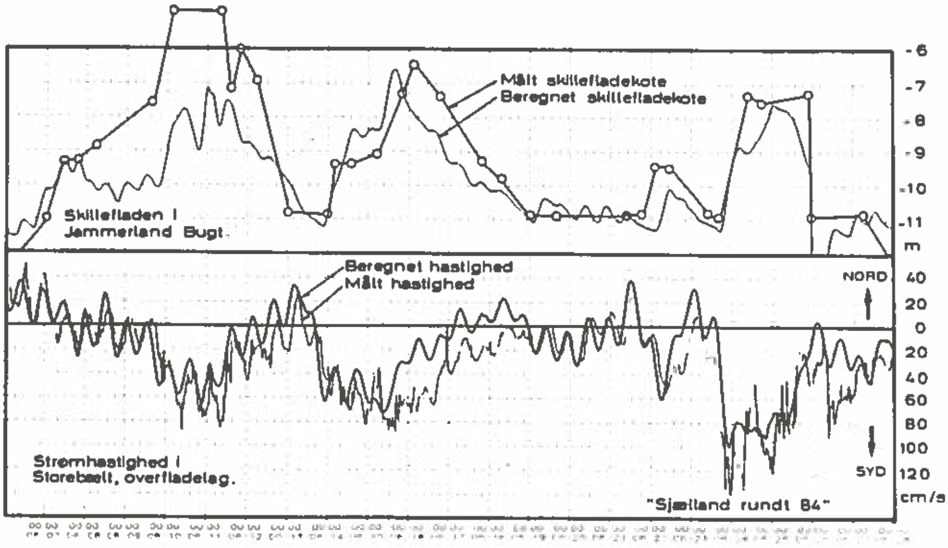


Fig. 6 Principskitse for S-22.

I fig.7 er vist eksempler på beregnede og målte værdier.



S22 modellen beregner såvel skillefladebeliggenhed som strømhastigheder med god overensstemmelse med målte værdier. I dette udsnit af den beregnede periode ses simuleringen af stormen under Sjælland Rundt sejladsen.

Fig. 7

På fig. 8 er vist eksempler på beregnede strømfelter og koncentrationer.

Overfladelag

Bundlag

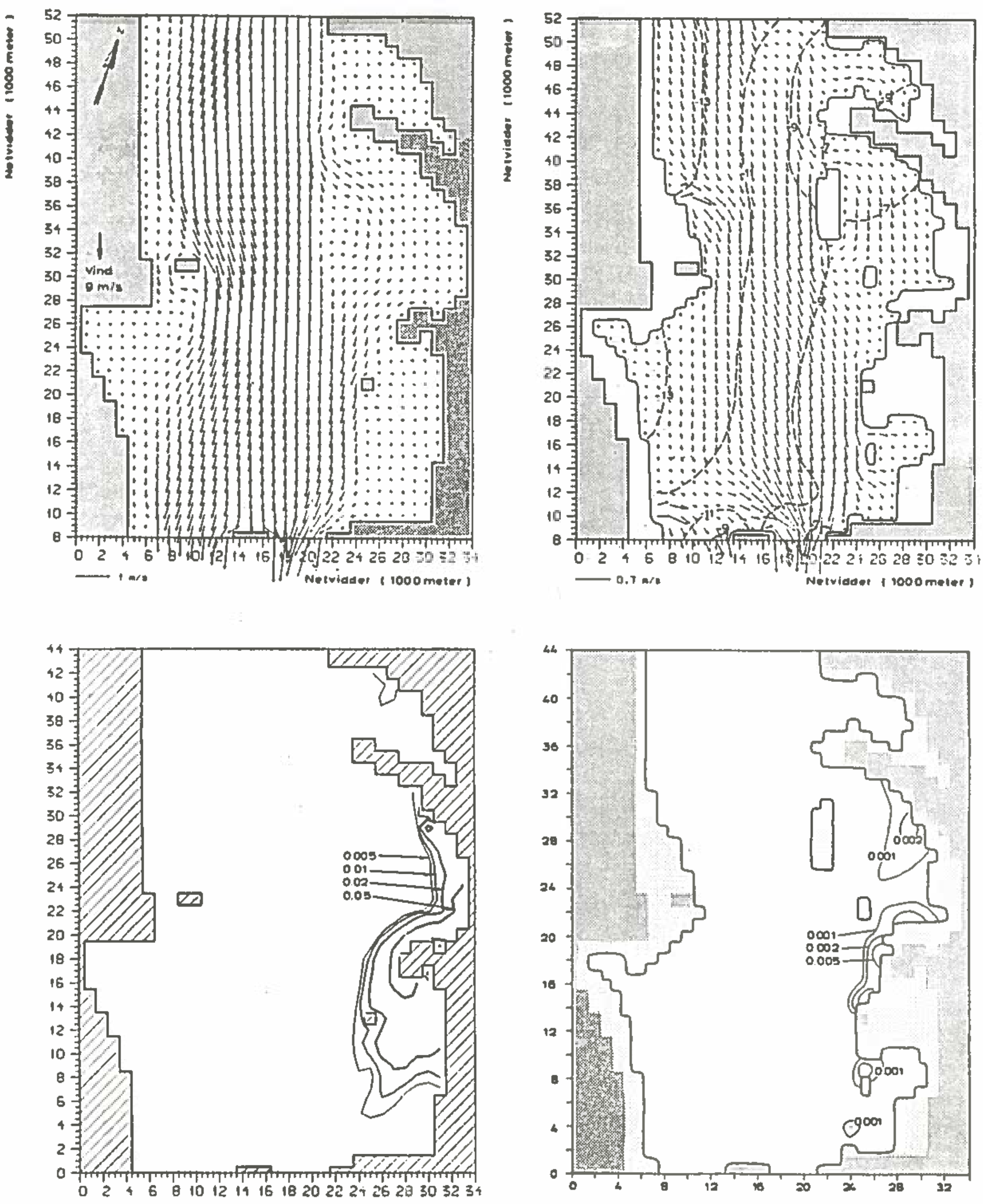


Fig. 8 Eksempel på S-22 resultater.

EKSEMPEL 3

STATIONÆR ØSTERSØMODEL.

I forbindelse med vurderingen af effekterne af Storebæltsforbindelsen er gennemført en række modelberegninger med en stationær model. Modellen er af samme type, som den der beskrives i vedlagte artikel fra Nordic Hydrology af Pedersen og Møller.

FREMTIDEN

Udviklingen indenfor modellering favoriserer numeriske modeller, som baserer sig på direkte løsning af de styrende differentialligninger. I fig. 9 er vist et eksempel på en sådan model anvendt på et vertikalsnit i Åbenrå Fjord.

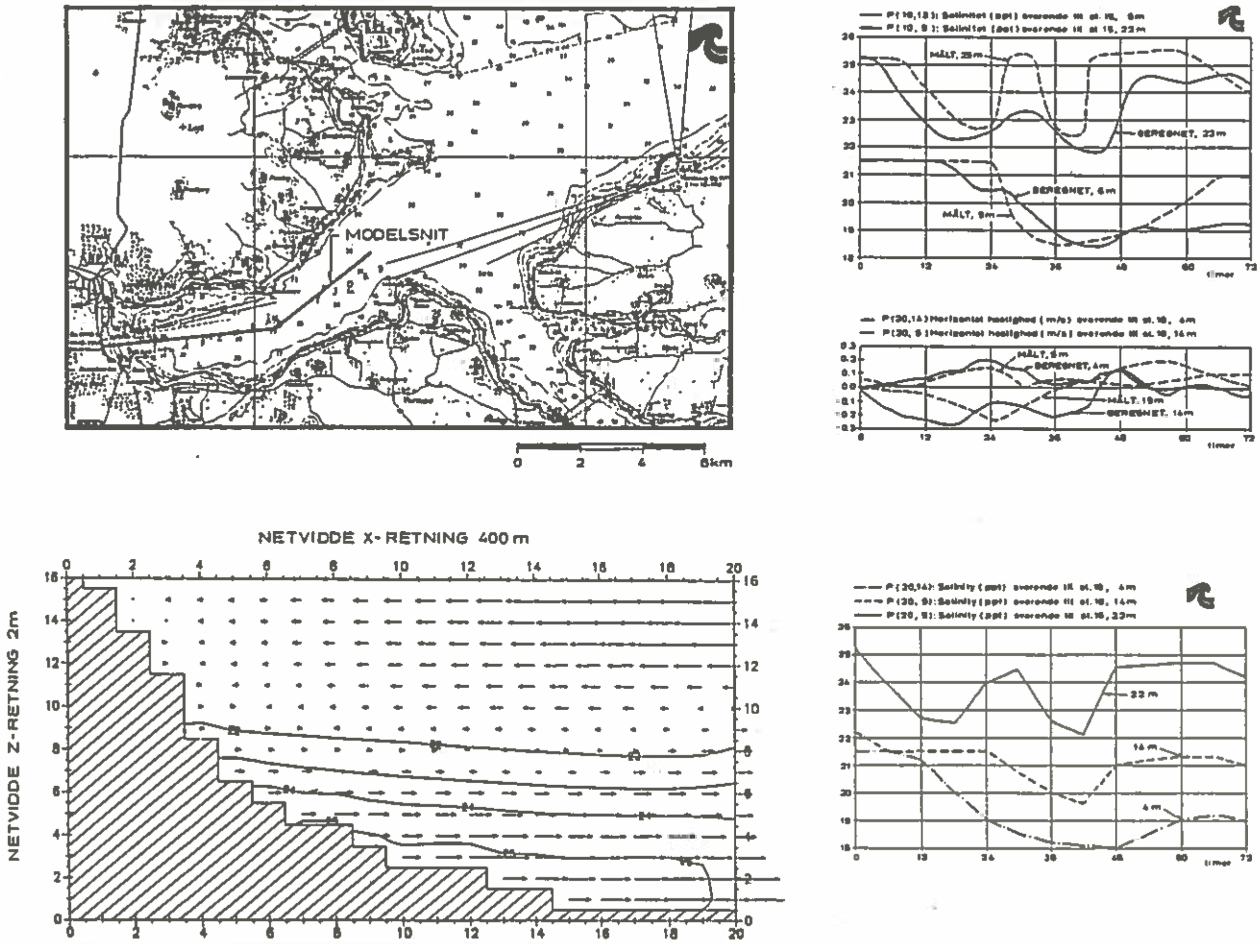


Fig. 9 Modelberegning med vertikalstrømningsmodel.

Udviklingen af fuldt 3-dimensionale modeller for oceanografiske strømninger finder sted flere steder i verden. Etablering af en 1. generationsmodel forventes på DHI at ske inden 1990. På det biologiske modelområde synes der stadig at være afgørende usikkerheder ved fastlæggelsen af de styrende processer og disses sammenhæng med de hydrografiske forhold.

Etablering af egnetlige konsekvensmodeller for biologiske indgreb, som f.eks. næringssalttilførslen, kræver derfor et yderligere forskningsarbejde specielt på det biologiske mmråde.

MECHANISMS RESPONSIBLE FOR OXYGEN CONDITIONS IN THE DEEP WATER OF THE KATTEGAT AND THE BELT SEA

v. Hans Schrøder & Jacob S. Møller, DHI

15th Conference of Baltic Oceanographers, 18-21 November 1986, Copenhagen.

MECHANISMS RESPONSIBLE FOR OXYGEN CONDITIONS IN THE DEEP WATER OF THE KATTEGAT AND THE BELT SEA

by Hans Schrøder and Jacob S. Møller
Danish Hydraulic Institute
Agern Allé 5
DK-2970 Hørsholm
Denmark

ABSTRACT

The equation for conservation of oxygen contained in a control volume with a 1 m^2 base and a height equal to the thickness of the deep water layer is set up and discussed. Using equations for conservation of salt and heat, the downward oxygen supply is determined for the critical autumn period. Each term in the oxygen balance is evaluated, except the horizontal supply term, which is determined from the balance. It appears that the downward entrainment of surface water and oxygen is a very important mechanism.

In the southern part of Kattegat, including the bordering coastal areas, a local minimum oxygen content exists. In these and similar regions the oxygen conservation equation is easily integrated over a complete annual cycle. The result obtained for S. Kattegat, which is in good agreement with observations, has yielded useful insight into the sensitivity of the annual minimum oxygen content towards changes in the oxygen consumption rate.

INTRODUCTION

The Great Belt Linkage project gives rise to some concern regarding its impact on the marine environments of Danish Waters and the Baltic Sea. At the same time, oxygen depletions in Danish and Swedish waters have apparently become returning events since 1981.

Both aspects require predictive tools in addition to careful consideration.

Since it is not possible to model the problem physically, a mathematical model is the only alternative. Below the main features of the oxygen conditions in Danish Waters are sketched in an attempt to provide a starting point for the development of such a model.

The mechanisms responsible for the supply and consumption of oxygen in the deep water of the Kattegat and the Belt sea are not well understood.

Below the conclusions are drawn:

1) The downward supply of oxygen due to entrainment of oxygen rich surface water into the deep water is a very important mechanism in the southern part of Kattegat and the dominating mechanism in the Belt Sea, and

2) The low oxygen content in the critical autumn period is extremely sensitive to (changes in) the oxygen consumption rate which in turn is closely linked to the production of organic matter in the upper photic zone of the sea.

The first conclusion is interesting when considering the Linkage project. The second conclusion supports earlier conclusions concerning the cause behind the deep water oxygen depletions, Ref. /10/ and /11/.

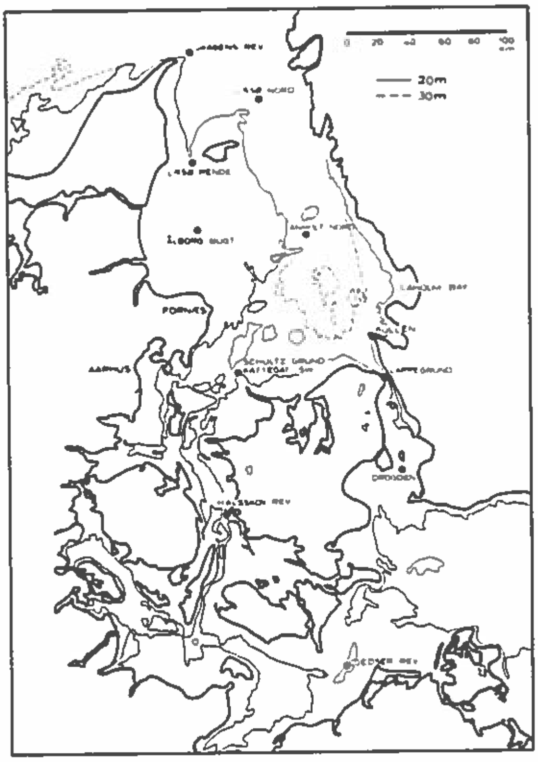


Figure 1

THE CONSERVATION EQUATIONS

Consider a control volume extending from the bottom to the pycnocline (H m) with a 1 m x 1 m base and placed perpendicular to the direction of the net current which flows in the positive direction of the x-axis.

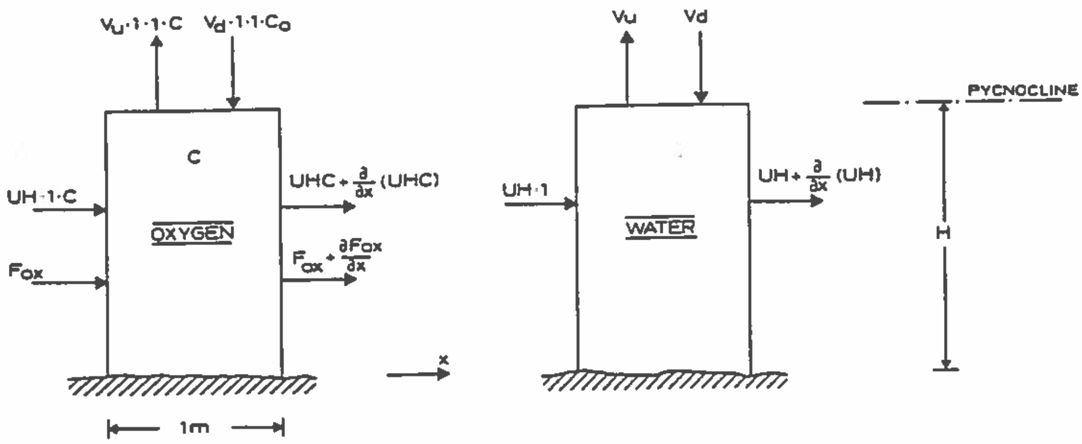


Figure 2 Fluxes of oxygen and water across the boundaries of the control volume.

The conservation of oxygen equation is:

$$\frac{\partial}{\partial t} (HC) = - \frac{\partial}{\partial x} (UHC) - \frac{\partial}{\partial x} (F_{ox}) + V_d C_o - V_u C - K \tag{1}$$

in which t = time
C = the oxygen content of the bottom layer
C_o = the oxygen content of the surface layer
F_{ox} = -1 H D $\partial C/\partial x$ (the dispersive flux of oxygen)
D = the longitudinal dispersion coefficient in the bottom layer.
V_d = the downward entrainment velocity
V_u = the upward entrainment velocity

The last term in Eq. (1) is the oxygen consumption per unit area and time. Oxygen is consumed by oxidation of part of the organic matter formed by photosynthesis in the productive top part of the surface layer (the primary production). There is no oxygen production in the bottom layer.

The conservation of water (volume) reads:

$$\frac{\partial H}{\partial t} = - \frac{\partial}{\partial x} (UH) + V_d - V_u \tag{2}$$

Using this equation and the expression for the dispersive flux, Eq. (1) can be rewritten as follows:

$$H \frac{\partial C}{\partial t} = - UH \frac{\partial C}{\partial x} + HD \frac{\partial^2 C}{\partial x^2} + V_d (C_o - C) - K \tag{3}$$

in which the first two terms on the right hand side are the horizontal supply by advection and dispersion respectively, and the third term is the vertical supply from the surface layer through the "lid" of the control volume due to entrainment.

In the same manner the equation of conservation of salt becomes:

$$H \frac{\partial S}{\partial t} = - UH \frac{\partial S}{\partial x} + HD \frac{\partial^2 S}{\partial x^2} + V_d (S_o - S) \tag{4}$$

where S_o is the salinity of the surface layer and S the salinity of the bottom layer.

An order of magnitude of the ratio between the dispersive term and the advective term can be evaluated from:

$$\frac{\text{Disp. term}}{\text{Adv. term}} = \frac{1}{U} \frac{D}{l_H}$$

where l_H is a characteristic horizontal mixing length.

With a net velocity U = 0.03 m/s, Ref. /4/, an estimated dispersion coefficient of 200 m^2/s and l_H in the order of 150 km (the distance from Skagen to central part of the Southern Kattegat) the ratio is in the order of 0.04. Hence, the transport by dispersion is at least an order of magnitude smaller than by advection.

With a satisfactory degree of approximation the conservation of oxygen equation then becomes:

$$H \frac{\partial C}{\partial t} = - UH \frac{\partial C}{\partial x} + V_d (C_o - C) - K \tag{5}$$

In analogy, the equation for conservation of salt is:

$$H \frac{\partial S}{\partial t} = - UH \frac{\partial S}{\partial x} + V_d (S_o - S) \tag{6}$$

and the conservation of heat:

$$H \frac{\partial T}{\partial t} = - UH \frac{\partial T}{\partial x} + V_d (T_o - T) \tag{7}$$

where T is the temperature in the bottom layer and T_o is the temperature of the surface layer.

Hence the average salinity and temperature in the deep water are determined by a balance between the supply by horizontal advection and the supply by vertical entrainment.

Note that C, S, and T are functions of position (x) and time (t). For instance:

$$C = C (x,t) \tag{8}$$

The total derivative of the function is:

$$\frac{dC}{dt} = \frac{\partial C}{\partial t} + U \frac{\partial C}{\partial x} \tag{9}$$

Hence, Eq. (5) can be written as:

$$H \frac{dC}{dt} = V_d (C_o - C) - K \tag{10}$$

which is the equation for conservation of oxygen in a control volume travelling at velocity U in the positive direction of the x-axis.

Eq. (6) and (7) can be written in the same manner.

AN ESTIMATE OF THE DOWNWARD ENTRAINMENT VELOCITY

Water from the top layer is entrained into the bottom layer since a production of turbulent kinetic energy takes place also in this layer. The entrainment mechanism is similar to the mechanism responsible for the entrainment of ambient water into a turbulent jet.

The upward rate of entrainment is larger than the downward rate, since the production of turbulent kinetic energy is larger in the surface layer, but the upward entrainment velocity needs not to be evaluated since it does not appear in Eq. (5).

Consequently it is to be expected that the salinity of the surface layer increases significantly from the Baltic to the Skagerak while the gradient in the bottom layer is smaller. However, it is clear that downward entrainment takes place since the salinity of the bottom layer decreases en route through the Kattegat and the Belt Sea. If, for instance, the salinity in the bottom layer is 30 o/oo (central part of Kattegat), it means that the bottom water is composed of 15% Baltic water (8 o/oo) and 85% North Sea water (34 o/oo).

The production of turbulent energy in the bottom layer is ultimately caused by the wind (and air pressure differentials), while the tide contributes with an insignificant amount.

Average monthly salinities extracted from Ref. /2/ are listed in Table 1.

Depth m	Skagens Rev Aug.	Skagens Rev Sept.	Anholt Nord/ Anholt Knob Aug.	Anholt Nord/ Anholt Knob Sept.	Kattegat SW/ Schultz's Grund Aug.	Kattegat SW/ Schultz's Grund Sept.	Halsskov Rev Aug.	Halsskov Rev Sept.
0	29.1	29.9	19.9	20.5	17.3	18.4	14.0	15.4
5	31.0	31.4	20.5	21.0	18.0	18.8	14.8	16.2
10	32.0	32.2	22.2	22.5	20.5	21.0	17.8	18.6
15	32.4	32.6	28.2	27.8	28.3	27.2	24.0	22.3
20	32.8	33.0	31.2	31.1	31.2	30.3	28.1	26.2
30	33.2	33.3	32.6	32.6	31.9	31.3	(28.8)	(27.2)
38	33.6	33.6	-	-	32.2	31.5	-	-
Average Surface (0-10 m)	30.4	31.0	20.6	21.1	18.2	19.1	15.1	16.4
Bottom (>10 m)	32.7	32.9	29.5	29.4	29.0	28.4	25.7	24.3
Distance km	130		80		100			

Table 1 Average monthly salinities (o/oo) for the period 1931-60. Values in brackets are estimated.

From Table 1 the terms $\frac{\partial S}{\partial t}$ and $\frac{\partial S}{\partial x}$ can be determined which allows a calculation of the downward entrainment velocity from Eq. (6):

$$V_d = \frac{H}{S_o - S} \left(\frac{\partial S}{\partial t} + U \frac{\partial S}{\partial x} \right) \qquad (11)$$

Results of the calculation are shown in Table 2.

Region	$\frac{\partial S}{\partial t}$ o/oo/s	$\frac{\partial S}{\partial t}$ o/oo/day	$\frac{\partial S}{\partial x}$ o/oo/m	$S_o - S$ o/oo	V_d (Eq. (11)) m/s	V_d (Eq. (11)) m/day
Between Anholt Nord and Kattegat SW (Southern Kattegat)	$-1.4 \cdot 10^{-7}$	-0.01	-10^{-5}	-9.3	$9.6 \cdot 10^{-7}$	0.08
Between Kattegat at SW and Halsskov Rev (Northern part of Gt. Belt)	$-3.9 \cdot 10^{-7}$	-0.03	$-3.7 \cdot 10^{-5}$	-9.6	$3.1 \cdot 10^{-6}$	0.27

Table 2 Values estimated from Table 1 and Equation (11), assuming H = 20 m and U = 0.03 m/s.

A CHECK ON THE ESTIMATE OF V_d

In order to enjoy confidence, the magnitude of V_d calculated above should be consistent with the magnitude which can be derived using temperature as a "tracer". In the bottom layer, temperature can be regarded as a conservative (non-decaying) "tracer".

Average monthly temperatures are extracted from Ref. /2/ and listed in Table 3.

Depth	Anholt Nord/ Anholt Knob		Kattegat SW/ Schultz's Grund		Halsskov Rev	
m	Aug.	Sept.	Aug.	Sept.	Aug.	Sept.
0	17.8	15.4	17.5	15.4	17.5	15.3
5	17.8	15.5	17.5	15.5	17.2	15.3
10	17.4	15.6	16.7	15.2	16.3	14.9
15	16.1	15.4	13.0	13.7	13.6	14.1
20	14.9	15.0	10.9	12.6	11.6	13.2
30	12.1	13.4	10.2	12.9	(10.9)	(12.7)
38	-	-	10.0	11.9	-	-
Average surface (0-10 m)	17.72	15.50	17.34	15.40	17.06	15.22
Bottom (10-30 m)	14.80	14.74	12.12	13.14	12.64	13.56

Table 3 Average monthly temperatures (^{o}C) for the period 1931-60. Values in brackets are estimated.

From Table 3 the terms $\frac{\partial T}{\partial t}$ and $\frac{\partial T}{\partial x}$ can be determined, and the downward entrainment velocity can be calculated from:

$$V_d = \frac{H}{T_o - T} \left(\frac{\partial T}{\partial t} + U \frac{\partial T}{\partial x} \right) \tag{12}$$

Results of the calculation are shown in Table 4.

Region	$\frac{\partial T}{\partial t}$	$\frac{\partial T}{\partial x}$	$T_o - T$	V_d (Equation (12))	
	$^{o}C/s$	$^{o}C/m$	^{o}C	m/s	m/day
Southern Kattegat	$1.8 \cdot 10^{-7}$	?	2.8	?	
Northern Gt. Belt	$3.7 \cdot 10^{-7}$	$0.5 \cdot 10^{-5}$	3.4	$3.1 \cdot 10^{-6}$	0.26

Table 4 Values estimated from Table 3 and Equation (8) assuming H = 20 m and U = 0.03 m/s.

In the Northern part of the Great Belt the magnitude of V_d calculated in this way is very close to the magnitude calculated using salinity as a "tracer".

In Southern Kattegat (between Anholt and Kattegat SW) $\frac{\partial T}{\partial x}$ cannot be determined since it becomes zero somewhere in this region. In other words there is a temperature minimum (in August-September) in the Southern part of Kattegat, probably close to the position of the Kattegat SW light vessel.

If we assume that the temperature minimum is close to Kattegat SW we have:

$$V_d \cong \frac{H}{(T_o - T)} \frac{\partial T}{\partial t} = \frac{20}{2.8} (1.8 \cdot 10^{-7}) = 1.3 \cdot 10^{-6} \text{ m/s}$$

which is 34% higher than the value found in Table 2.

A further check on the magnitude of V_d can be obtained in the following way:

In the region at or close to the temperature minimum the horizontal heat supply term vanishes and consequently we have (from Eq. (7) and (9)):

$$\frac{dT}{dt} + \frac{V_d}{H} T = \frac{V_d}{H} T_o \tag{13}$$

This is equivalent to considering a stationary control volume which exchanges heat with the surface layer only, see Fig. 3.

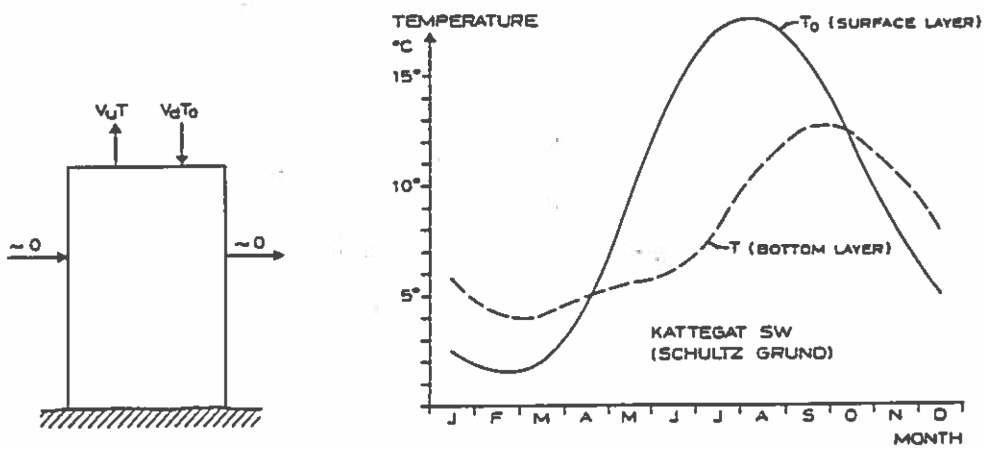


Fig. 3 The temperature control volume at or close to the temperature minimum and the average annual temperature variation in the surface and bottom layer.

Long term temperature measurements, Ref. /2/, carried out at Schultz Grund/ Kattegat SW show that the temperature variation in the bottom layer can be regarded as a damped and phase-lagged reflection of the variation in the surface layer.

The surface temperature, T_o, in Eq. (3) can be assumed to vary sinusoidally over the year:

$$T_o = \bar{T} + a \cos (\omega t) \quad ; \quad \omega = \frac{2\pi}{T}$$

With this, the solution to Eq. (13) is:

$$T = \bar{T} + \frac{\frac{V_d}{H}}{\sqrt{\omega^2 + \left(\frac{V_d}{H}\right)^2}} \cdot a \cdot \cos(\omega t - \phi) \tag{14}$$

in which

$$tg\phi = H \frac{\omega}{V_d} \tag{15}$$

The downward entrainment can then be calculated either by Eq. (14) using the damping ratio or by Eq. (15) using the phase lag.

From Ref. /2/ the damping ratio is calculated at 0.73 and the phase lag at 1.7 months, corresponding to $\phi = 0.89$ radians. V_d is calculated at $2.5 \cdot 10^{-6}$ m/s and $3.2 \cdot 10^{-6}$ m/s using Eq. (14) and Eq. (15) respectively. Hence, as an annual average for the Southern part of Kattegat we have:

$$V_{d\ \text{annual average close to Kattegat SW}} \cong 3 \cdot 10^{-6}\ \text{m/s} = 0.26\ \text{m/day}.$$

The value for the critical period in September is probably somewhat smaller, say $2 \cdot 10^{-6}$ m/s.

The value estimated in this way agrees well with the value calculated for the region between Kattegat SW and Halsskov Rev using salt and temperature as a tracer (see Table 2 and 4).

ENTRAINMENT VELOCITIES AND DIFFUSION COEFFICIENTS

The results above indicate that the order of magnitude of the entrainment velocity in the period August-September is:

10^{-6} m/s in the southern part of Kattegat
$2 \cdot 10^{-6}$ m/s at the entrance to the Great Belt
$3 \cdot 10^{-6}$ m/s in the northern part of the Great Belt

Corresponding diffusion coefficients can be calculated as the product of the entrainment velocity and a characteristic vertical mixing length. If we assume the mixing length to be the distance from the middle of the surface layer to the middle of the bottom layer, we have $l_z \cong 15$ m and hence:

$$D_z = V_d\, l_z \tag{16}$$

in the range between 0.15 cm^2/s in the southern part of Kattegat and 0.5 cm^2/s in the northern part of the Great Belt. These values agree well with results of measurements reported in Ref. /9/.

THE OXYGEN BALANCE

Returning to the conservation of oxygen equation:

$$H \frac{\partial C}{\partial t} = - UH \frac{\partial C}{\partial x} + V_d (C_o - C) - K$$

the terms are evaluated in the following (from left to right):

Rate of change term: $H \frac{\partial C}{\partial t}$

Judging from Fig. 4 the oxygen content at Anholt and Kullen decreases steadily from a value close to saturation in March (9 mg/l) to about 40% of saturation in September. At this critical time of the year the oxygen content is at a minimum and hence $\partial C/\partial t = 0$.

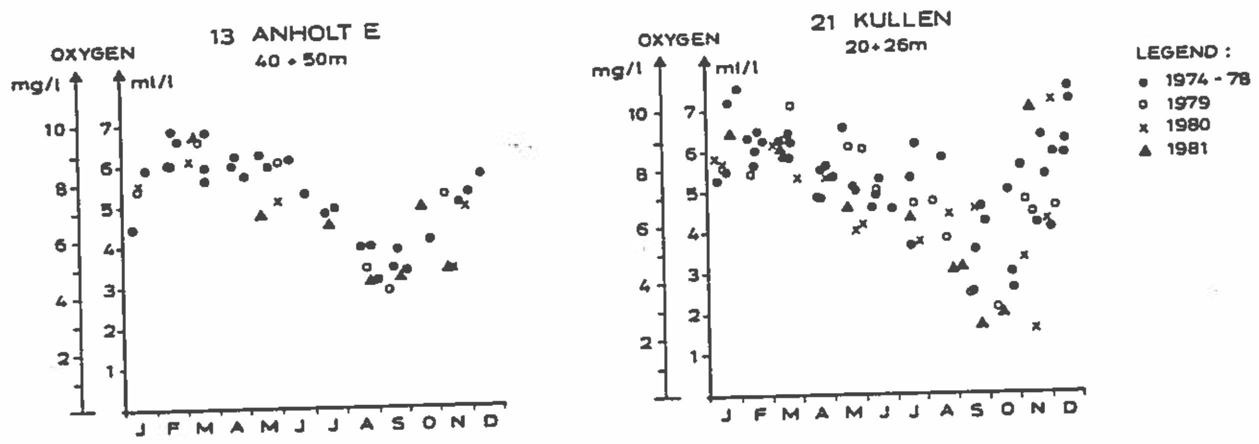


Fig. 4 Observed oxygen contents in the deep water at Anholt E (water depth 55 m) and Kullen (water depth 26 m), Ref. /4/.

The horizontal supply term: $- UH \frac{\partial C}{\partial x}$

This term is difficult to evaluate since U is not very accurately determined and especially since the horizontal gradient is not well defined. It appears that minimum contents are somewhat lower at Kullen than at Anholt. However, a comparison of the individual measurements yields no clear picture.

The term is no doubt the one which lends itself poorest to determination by measurements and shall therefore be calculated from the balance, see below.

The vertical supply term: $V_d \ (C_o - C)$

The term is easily evaluated on the basis of the findings above. For instance for S. Kattegat we find:

$$V_d \ (C_o - C) = 10^{-6} \text{ m/s} \cdot 24 \cdot 3600 \text{ s/day} \cdot (9-4) \text{ g/m}^3 = 0.43 \text{ g/day}$$

per m^2. In the northern part of the Great Belt the vertical supply is about 3 times this amount.

The oxygen consumption term: K

Part of the organic matter formed in the photic upper zone sinks down through the pycnocline to the deep water where it is mineralized at the expense of oxygen. If we assume that 50% of the organic matter sinks through the pycnocline and is oxidized there, the annual oxygen consumption is about 220 g O_2 per m^2 at a primary production rate of 140 gC per m^2 per year. The oxygen consumption in the deep water in August-September has been estimated at 0.77 g O_2/m^2/day, see below.

Measurements carried out in Kattegat by B.B. Jørgensen, Ref. /5/, indicate that the rate of consumption in the sediments in the waters between the Baltic and the North Sea is in the range 0.3-0.8 g O_2/m^2/day with an average of about 0.5 g O_2/m^2 day, see Fig. 5. However, to this value shall be added the consumption of oxygen in the deep water column.

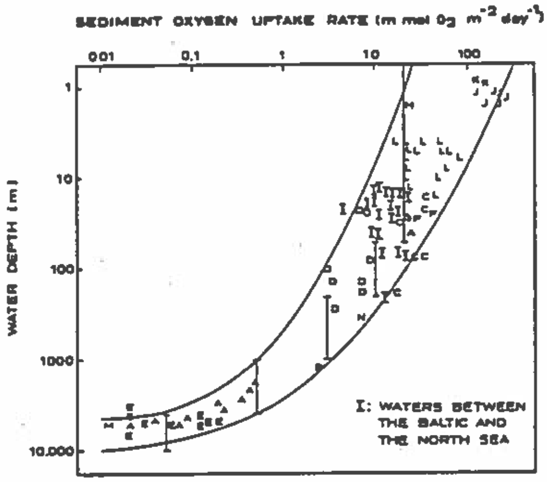


Fig. 5 Oxygen consumed at the bottom at different ocean depths, from Ref. /5/.

The results of the measurements are in accordance with the percentage of the mean annual primary production collected in sediment traps at levels corresponding to the depth of the Kattegat, Ref. /5/.

The oxygen balance

The horizontal oxygen supply can now be calculated from the balance equation. Assuming that the rate of change term and the consumption rate are approximately the same for the 3 regions in question, the balances can now be determined. The result is shown in Table 5.

		Region		
		(a) Kattegat	(b) Entrance to the Belt Sea	(c) Northern part of the Great Belt
Rate of change	$H\ \partial C/\partial t$	0	0	0
Horizontal supply*)	$-UH\ \partial C/\partial x$	0.34	-0.13	-0.58
Vertical supply	$V_d\ (C_o-C)$	0.43	0.90	1.35
Oxygen consumption	K	0.77	0.77	0.77

*) calculated from the balance.

Table 5 Calculated terms in the conservation of oxygen equation in g O_2/day (per m^2).

The results in Table 5 indicate the presence of a (local) oxygen minimum in the vicinity of the entrance to the Belt Sea ($\partial C/\partial x$ passes through zero). Data extracted from Ref. /9/ and presented in Fig. 6 support this indication.

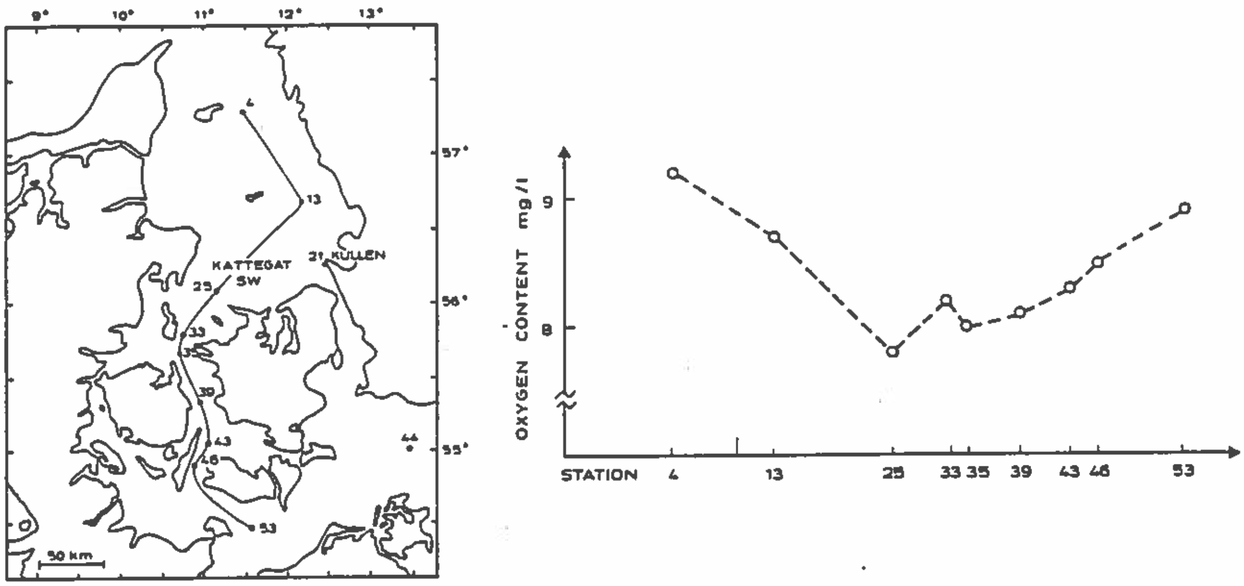


Fig. 6 Annual mean oxygen contents at depths between 20 and 30 m measured over the period 1974-79 from Ref. /9/.

The calculation further shows that the vertical supply constitutes a significant part of the total supply in the Southern part of Kattegat and that the vertical supply totally dominates in the Great Belt.

THE ANNUAL VARIATION OF THE DEEP WATER OXYGEN CONTENT

The oxygen balance considerations above offers some insight into the question concerning the impact of increased oxygen consumption rates on oxygen depletion events which have been observed in the Kattegat since 1981.

As indicated above the oxygen content reaches a minimum at the entrance to the Belt Sea. The region in which the minimum occurs is characterized by $\partial C/\partial x = 0$ and hence no horizontal supply of oxygen.

It is reasonable to believe that the region characterized by no horizontal supply of oxygen is not limited to the entrance to the Belt Sea but encompasses the coastal regions bordering the Kattegat to the south, southwest, and southeast, i.e. the coastal waters off North Zealand, the Bay of Århus and the Swedish coast including the Bay of Laholm.

With no horizontal oxygen supply we have:

$$\frac{\partial C}{\partial t} = \frac{dC}{dt}$$

and the conservation of oxygen equation becomes the ordinary differential equation:

$$H \frac{dC}{dt} = V_d \, (C_o - C) - K \tag{18}$$

It is virtually impossible to integrate this equation in an analytical form since the coefficient to the dependent variable, as well as the right hand side of the equation are far from being constant over the annual cycle. Fortunately it is relatively easy to carry out a numerical integration.

In the following the annual variation of all independent variables in Eq. (18) are evaluated, from left to right:

H: the thickness of the deep water layer

The distance from the surface to the level of the pycnocline is a slowly varying function over the annual cycle. During the summer the pycnocline reaches its highest level in Kattegat and the Belt Sea, but this level is only 2-3 m higher than the level in autumn and winter. In the present context it therefore yields a satisfactory approximation if we take the level to be constant over the year.

In the central and southern part of the Kattegat this (constant) level has been estimated to be located about 13 m below the surface. Hence the thickness of the deep water layer is taken to be

$$H = D - 13 \text{ m}$$

where D is the total water depth.

V_d : the downward entrainment velocity

Using long-term average monthly salinities at Schultz Grund and Halsskov Rev light vessels, in combination with modified Knudsen-relations, the annual variation of the vertical diffusion coefficient (D_z) in the northern part of the Great Belt has been determined. The result is shown in Fig. 7.

The magnitude of V_d is determined by the wind and the stabilizing buoyancy forces. It can be shown that V_d, and hence also D_z, is proportional to the wind speed to the power 3 and inversely proportional to the relative density difference between the layers and to the thickness of the deep water layer:

$$V_d = \frac{D_z}{l_z} = \text{constant} \cdot \frac{w^3}{\Delta \, g \, H} \tag{19}$$

The annual variation of D_z calculated using this relationship is shown in Fig. 7 (dashed curve). Monthly average wind speeds from a weather station situated at Fornæs have been utilized.

The annual variation shown in Fig. 7 can be assumed to apply in the Belt Sea as well as the Kattegat, but the actual magnitude of V_d of course varies from location to location.

In order to produce the annual variation of V_d in the southern part of the Kattegat, the annual variation shown in Fig. 7 is normalized with respect to the value at the end of August - beginning of September and multiplied by the value of V_d found valid for the southern Kattegat at that time of the year ($V_d = 10^{-6}$ m/s = 2.63 m/month). The values in m/month centered at the time of transition from one month to the next are listed below:

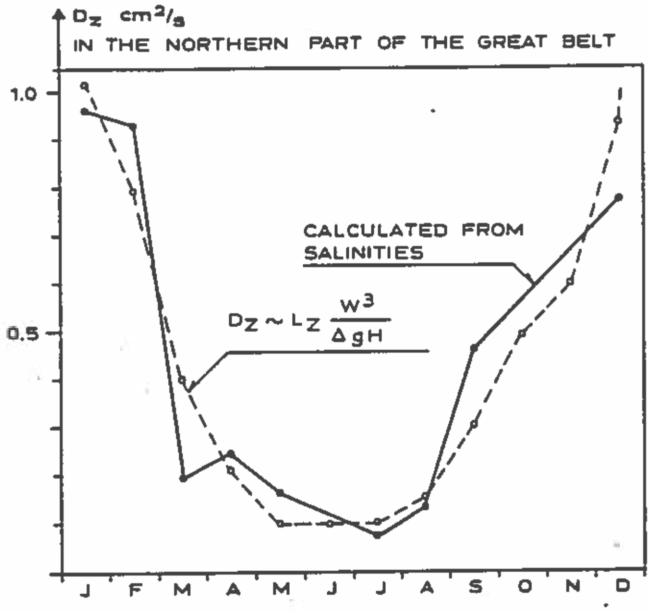


Fig. 7 The annual variation of D_z (cm^2/s) in the northern part of the Great Belt.

	m/month		m/month
Jan.		July	
	10.41		1.50
Febr.		Aug.	
	6.86		2.63
Mar.		Sept.	
	3.42		4.47
Apr.		Oct.	
	1.71		6.05
May		Nov.	
	1.13		8.68
June		Dec.	
	1.13		11.20
July		Jan.	

C_o : the oxygen content of the surface layer

The oxygen content in the surface layer is always close to saturation. During the productive period, oxygen produced as a result of the photosynthesis gives rise to a slight supersaturation (+ 10% approx.), while the saturation percentage is lower (about 90%) during autumn and winter, due to intense mixing with the deep water layer.

However, the oxygen content of the surface layer is dominated by the supply of oxygen directly from the atmosphere, and C_o can therefore be assumed to be an independent variable. Average oxygen contents (in mg/l) from observations in the Kattegat and the Belt Sea are taken from Ref. /9/ and listed below:

Jan.	11.2 mg/l	July	9.5 mg/l
Febr.	12.4	Aug.	8.3
Mar.	12.4	Sept.	8.8
Apr.	12.2	Oct.	9.5
May	11.4	Nov.	9.7
June	10.7	Dec.	10.9

The oxygen saturation content depend on salinity as well as temperature. The lowest content (8.3 mg/l) is found in August, mainly due to the high water temperature at that time of the year.

K : the oxygen consumption rate

The relationship between the rate of production of organic carbon in the surface layer and the oxygen consumption rate in the deep water layer still remains to be established. But it is fair to assume that K is proportional to the rate of primary production.

Since it takes about 3.2 g of oxygen to oxidize 1 g carbon, the annual oxygen consumption in the deep water can be expressed as:

$$\bar{K} = \beta \, P \; 3.2 \text{ g } O_2/m^2/\text{year} \tag{20}$$

where β is the fraction of the organic matter which sinks below the pycnocline and is oxidized there, and P is the total primary production in the surface layer in g C/m^2/year.

The oxygen consumption rate varies over the year as a consequence of the water temperature variation. It is generally assumed that this temperature variation can be taken as:

$$2^{\left(\frac{T}{10} - 1\right)}$$

where T is the temperature. The temperature dependence indicates that the oxidation rate is doubled when the temperature is raised 10^{o}C.

Using long-term monthly average temperatures of the deep water at Schultz Grund light vessel, the values of the function have been determined and normalized by the annual average value. These normalized values have then been multiplied by $\bar{K}/12$ with $\bar{K}$ calculated from Eq. (20) using $\beta = 0.5$. The magnitude of K/P thus calculated are as follows (centered at the time of transition from one month to the next):

Jan.		July	
	0.112		0.141
Febr.		Aug.	
	0.107		0.167
Mar.		Sept.	
	0.107		0.179
Apr.		Oct.	
	0.111		0.171
May		Nov.	
	0.113		0.149
June		Dec.	
	0.120		0.128
July		Jan.	

It appears that the oxygen consumption rate reaches a maximum value close to the transition from September to October when it is about 1.34 times the annual average value.

If, as an example, we take the annual primary production to be 140 g of carbon per m^2, the annual maximum oxygen consumption rate becomes:

$$K_{max} = 0.179 \cdot 140 = 25.1 \frac{g\ O_2}{m^2\ \text{month}} \cong 0.8 \frac{g\ O_2}{m^2\ \text{day}}$$

In August-September K is a little smaller

$$K_{aug-sept} = 0.167 \cdot 140 \frac{1}{30.5} = 0.77 \frac{g\ O_2}{m^2\ \text{day}}$$

which is a probable value, see Ref. /5/.

Integration of Equation (18)

The numerical integration is carried out using a time step of 1 month taking care that the value of the coefficient to the dependent variable and the right hand side of the equation are centered in the time step. Starting with a value close to saturation during winter, the solution converge quickly to a periodical function (after only 2 annual cycles).

In a worked example the following magnitudes of the dependent variables have been chosen:

$H = 25$ m
$V_d = 10^{-6}$ m/s
$P = 140$ g C/m^2/year ($\beta = 0.5$)

These values are chosen to fit conditions valid for the southern part of Kattegat and the coastal regions bordering S. Kattegat. Among observations from this region are observations from Kullen (water depth 26 m).

The result of the numerical integration is plotted along with oxygen observations from Kullen in Fig. 8.

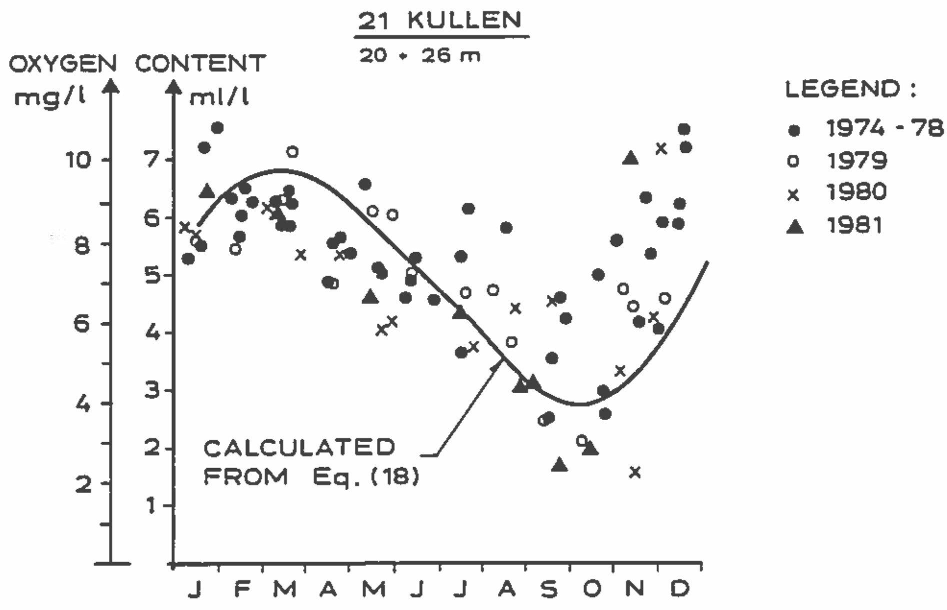


Fig. 8 Observed and calculated oxygen content over the year in the deep water at Kullen (S. Kattegat, see Fig. 1). Observations are taken from Ref. /4/.

The "model" is a long-term average model producing monthly normal values and full accordance with measurements carried out over an actual year is therefore not to be expected. However, the calculation appears to reproduce the general trend in a satisfactory manner, and it correctly predicts the time of occurrence of the minimum oxygen content (C_{min}) to September-October.

The time location of C_{min} is determined by 3 factors in combination::

1) the annual variation of C_o
2) the high deep water temperature in autumn, and
3) the annual variation of V_d

THE ANNUAL MINIMUM OXYGEN CONTENT

The annual minimum oxygen content, C_{min}, occurs for dC/dt = 0 and can from Eq. (18) therefore be written as:

$$C_{min} = C_o - \frac{K}{V_d}\Bigg|_{\frac{dC}{dt} = 0} \tag{21}$$

For a given variation of C_o, C_{min} is therefore a function of two independent variables

$$C_{min} = C_{min}\ (K, V_d) \tag{22}$$

The first independent variable (K) is influenced by man since the primary production (to which K is assumed proportional) has increased over recent years as a result of increased loads of nutrients, especially nitrogen nutrients, Ref. /10/. In Ref. /12/ it is reasoned that nitrogen loads from Danish agriculture has increased by a factor of 3 since the second world war and that this is a probable cause for the increase in primary production rates reported in Ref. /10/.

The second independent variable (V_d) is a function of the wind energy (proportional to w^3) and it is therefore, of course, beyond the influence of man.

It is of considerable interest to investigate C_{min} with respect to its sensitivity to changes in the man-influenced independent variable (K) and the natural independent variable (V_d). To this end we differentiate Eq. (22) and obtain:

$$d\,C_{min} = \frac{\partial\, C_{min}}{\partial\, K}\, d\,K + \frac{\partial\, C_{min}}{\partial\, V_d}\, d\,V_d$$

From Eq. (21) we find the partial derivates as:

$$\frac{\partial\, C_{min}}{\partial\, K} = - \frac{C_o - C_{min}}{K} \tag{23}$$

and

$$\frac{\partial\, C_{min}}{\partial\, V_d} = \frac{C_o - C_{min}}{V_d} \tag{24}$$

Unfortunately, it is not possible to determine these partial derivatives analytically, since the exact time of occurrence of dC/dt = 0 cannot be determined.

However, the numerical integration can be carried out for a range of values of K (and constant V_d) and a range of V_d (and constant K) to produce the desired result.

Fig. 9 shows the annual variation of C for a range of oxygen consumption rates. The sensitivity of C to changes in P and consequently K is evident.

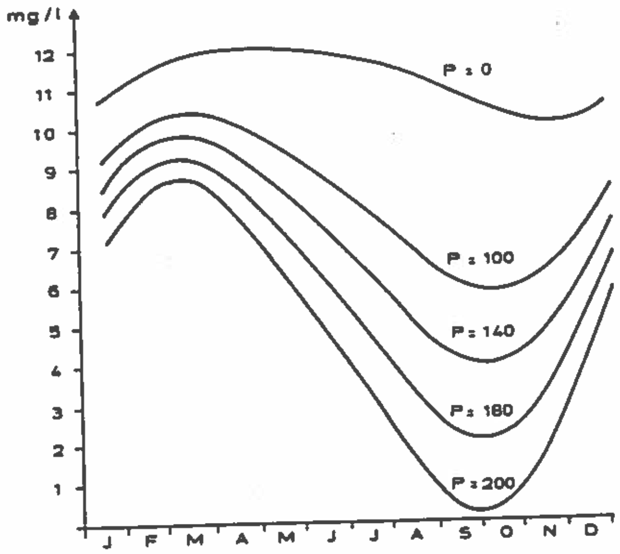


Fig. 9 The annual variation of the oxygen content in the deep water of S. Kattegat for a range of values of the rate of primary production (P in g C/m^2/year).

A similar calculation has been carried out for the range of values of V_d.

The variation of C_{min} is plotted in Fig. 10.

From Fig. 10 a) and b) the partial derivatives are determined as follows:

$$\frac{\partial\ C_{min}}{\partial\ K}\ dk = -6.1\ \frac{dK}{K}\ mg/l$$

$$\frac{\partial\ C_{min}}{\partial\ V_d}\ d\ V_d = 6.1\ \frac{d\ V_d}{V_d}\ mg/l$$

The magnitude of the first derivative shows that the minimum oxygen content decreases 2 mg/l when the oxygen consumption rate (or the primary production rate) is increased only 32% and everything else is equal. Considering that C_{min} in the southern part of the Kattegat is in the order of 4 mg/l, the calculation hence indicates that the marine life in the deep water of Kattegat is at stake for a modest increase in the oxygen consumption rate, or equivalent: in the primary production rate. As already indicated, it is most likely that the primary production has increased more than the above mentioned amount over recent years.

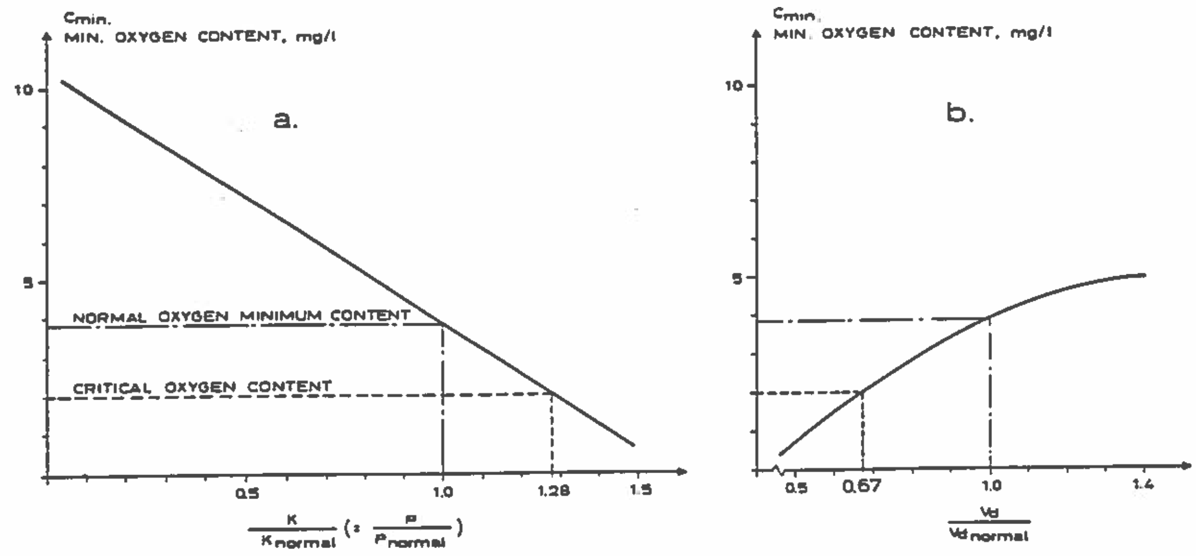


Fig. 10 The annual minimum oxygen content C_{min} as a function of K (constant V_d) and as a function of V_d (constant K).

An investigation carried out on data from the Arkona Basin (st. 44, se Fig. 6) reported in Ref. /10/ shows that the minimum oxygen content in this region has decreased about 2.3 mg/l from the turn of the century to the time of the Belt Project (1974-79). Assuming that V_d is of the same order to magnitude in Kattegat and the Western part of the Baltic, this finding indicates an increase in the primary production of about 37%. However, it is noted that the primary production in the middle - late 70'ies were relatively low due to a low precipitation during these years, Ref. /12/. (Leaching of nitrogen nutrients from farmlands is low when precipitation is low).

The actual increase in primary production from the turn of the century to the 1980'ies is more likely to be in the order of 100% and this increase is thus large enough to explain why the deep water of Kattegat and the Western Baltic is suffering from oxygen shortage today.

The magnitude of the second derivative shows that C_{min} decreases 2 mg/l (from 4 to 2 mg/l) in case V_d at the critical time of the year drops 33% below the normal level. This is probably within the normal variation range of the wind energy. It corresponds to a decrease of about 10% in the average wind speed. Consequently, the calculation indicates that oxygen depletions can occur also as a natural phenomenon. This is in accordance with statements made by Danish fishermen, reported in Ref. /10/.

Needless to say, the negative effect of an increase in K can be amplified by a coincident decrease or weakened by a coincident increase in the wind energy input.

Hence both the man-influenced and the meteorologically determined independent variable vary within a range large enough to create critical oxygen conditions in themselves. The combined effect of increasing K is, however, that the frequency of occurrence of critical oxygen conditions is significantly increased. This is in principle illustrated in Fig. 11.

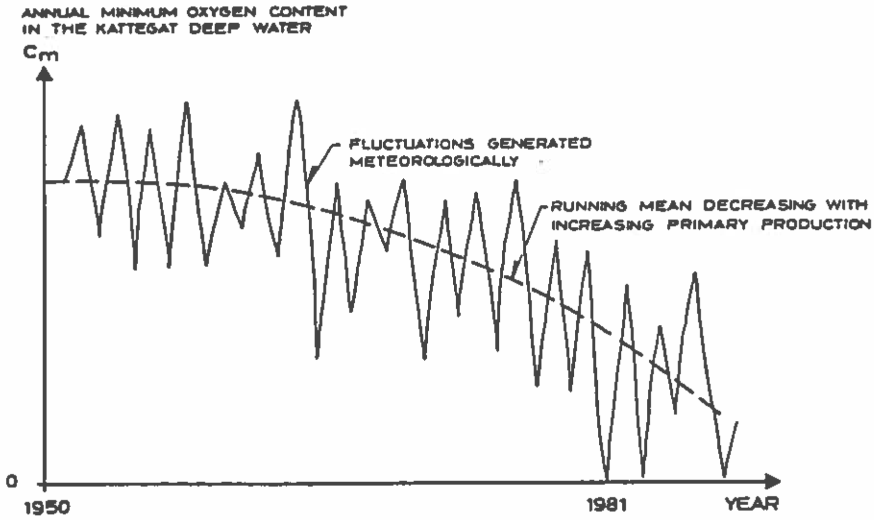


Fig. 11 Long term variation of the annual minimum oxygen content in the Kattegat deep water, in principle.

MODEL CONSIDERATIONS

Oxygen conditions in the Baltic have deteriorated since the turn of the century and conditions in the transition zone between the Baltic and the North Sea have deteriorated over the past 5-6 years.

The Danish Government is in the process of undertaking to demonstrate that the planned Great Belt Linkage will have no negative effects on the marine environment. In order to do so it appears necessary, not only to predict the effects of the linkage on oxygen and salinity conditions in the waters between the North Sea and the Baltic Sea, in addition it appears mandatory to be able to account for changes which have already taken place. Otherwise one will not be able to distinguish between the possible effects of the linkage and the effect of other causes, if conditions come to worse.

When considering the possible impact of the linkage, a fully dynamic two-layer model with a time step small enough to resolve dynamic events is required. Such a model should be used to study the details of the mixing and transport processes in the vicinity of the linkage, in the Kattegat and the entire Belt Sea in well defined oceanographic and meteorological events.

When considering the oxygen depletions, a long-term simulation model using much larger time steps (in the order of days) could be employed to study already observed oxygen conditions as a function of nutrient loadings and to predict the effect of alternative remedial measures.

The analysis above has clearly indicated that such models need to be carefully calibrated and verified with respect to deep water salinities in order to enjoy confidence. To this end the long-term observations of salinities carried out at the Danish light vessels are well suited. Some of these observations were initiated as early as 1881 (but have now unfortunately ceased because the light vessels have been withdrawn).

With respect to the Baltic, the causes for the deterioration of oxygen conditions since 1900 still remains to be clarified. It also remains to be elucidated to what extent the deep water is supplied with oxygen through the Belt Sea and to what extent it is supplied by downward entrainment. Until this is done it is not obvious what the requirements to the linkage project should be.

CONCLUSION

In the southern part of Kattegat downward entrainment is responsible for a considerable part (50%) of the supply of oxygen to the deep water.

At the entrance to the Belt Sea, and in the Belt Sea itself, downward entrainment is the only mechanism which supplies oxygen to the deep water.

The magnitude of the minimum oxygen content is extremely sensitive to changes in the magnitude of the oxygen consumption rate. This rate, in turn, is closely linked to the magnitude of the primary production.

REFERENCES

In English if not otherwise stated.

/1/ Danish Meteorological Institute (1983): Nautical - Meteorological Annual 1932 (in Danish)

/2/ Danish Meteorological Institute (1984): The Climate of Denmark Summaries of Observations from Light Vessels. Climatological Papers No. 10 and 11.

/3/ Gargas, E. et. al (1978): "Dynamics of Baltic Ecosystems and Causes of Their Variability". Kieler Meeresforschungen.

/4/ Jacobsen, T.S. (1982): "The Oxygen Balance in the Kattegat Deep Water". 13th Conference of Baltic Oceanographers, 1982, Helsinki, Finland.

/5/ Jørgensen, B.B. (1983): "Processes at the Sediment - Water Interface. In: The Major Biogeochemical Cycles and Their Interactions. Scope".

/6/ Jørgensen, B.B. and Sørensen, J (1985): "Seasonal Cycles of O_2, NO_3^- and SO_4^{--}. Reduction in Estuarian Sediments: The Significance of an NO_3^- Reduction Maximum in Spring". Marine Biology - Progress Series, Vol. 24, pp. 65-74.

/7/ Krogh, O. et. al (1978): "A Numerical Model of Oxygen Conditions in a Two-Layered Marine Environment". International Conference on Water Resources Engineering. Bangkok, Thailand, 10-13 January 1978.

/8/ Malmgren-Hansen, A. (1984): "Kattegat Model Calculations", Miljøprojekter 59. The National Agency for Environmental Protection (in Danish).

/9/ Miljøstyrelsen (The Danish National Agency of Environmental Protection, Denmark) (1981): "The Belt Project: Evaluation of the Physical, Chemical and Biological Measurements".

/10/ Miljøstyrelsen (The Danish National Agency of Environmental Protection, Denmark) (1984): "Oxygen Depletion and Fish Kill in 1981 - Extent and Causes" (in Danish).

/11/ Schrøder, H. (1985): "Nitrogen Losses from Danish Agriculture - Trends and Consequences". Agriculture, Ecosystems and Environment, 14 (1985), pp. 279-289.

/12/ Schrøder, H. (1984): "The Development over Time of Nitrogen Losses from Danish Agriculture and the Consequences for the Environment, Water Quality Institute (in Danish).

ÅRSAGER TIL OG EFFEKTER AF EUTROFIERING I KATTEGAT OG BÆLTHAVET

v. Gunni Ærtebjerg

Miljøstyrelsens Havforureningslaboratorium.

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ÅRSAGER TIL OG EFFEKTER AF EUTROFIERING I KATTEGAT OG BÆLTHAVET.

Marinbiolog Gunni Ærtebjerg
Miljøstyrelsens Havforureningslaboratorium
Jægersborg Alle 1 B
2920 Charlottenlund

1. Introduktion

I løbet af 1980'erne er der for Kattegat og Bælthavet beskrevet en række forandringer, der alle synes at være forårsaget af en stigende eutrofiering. Fytoplanktonets produktion er steget, og u-sædvanlige masseforekomster samt forekomst af nye eventuelt toxiske arter er observeret. Gentagne gange er oxygen-mangel i bundvandet observeret med død af fisk og bunddyr tilfølge. Bundfaunaens artssammensætning og biomasse er ændret, og fiskebestandene udviser store ændringer i bestandsstørrelserne.

Tilsammen tegner de fremkomne oplysninger et billede af et havområde, hvor den stigende eutrofiering har nået et niveau, som kan have alvorlige konsekvenser for erhvervsfiskeriet med væsentlig nedgang i en række fiskerier på ikke-pelagiske bestande.

I det følgende søges årsagerne til og effekterne af den stigende eutrofiering belyst.

2. Belastning med næringssalte

Belastningen til Kattegat, Bælthavet og Øresund (areal 38.000 km^2) med total-P og total-N fra Danmark, Sverige og atmosfæren for peridoen 1975-81 er opgjort af Miljøstyrelsen (1983, 1984). Belastningen med total-P var i denne periode stort set konstant på ca. 14.000 t P pr. år, og jævnt fordelt over året. Dertil skal lægges ca. 3.600 t P pr. år fra Øst- og Vesttyskland til det sydlige Bælthav (Edler 1984). Belastningen med total-P kan antages at være forblevet på samme niveau også i 1982-85.

Belastningen med total-N fra land var i perioden 1975-81 først og fremmest afhængig af størrelsen af vandafstrømningen fra land. Belastningen i årene 1982-85 er derfor her estimeret ud fra vandafstrømningen fra Danmark

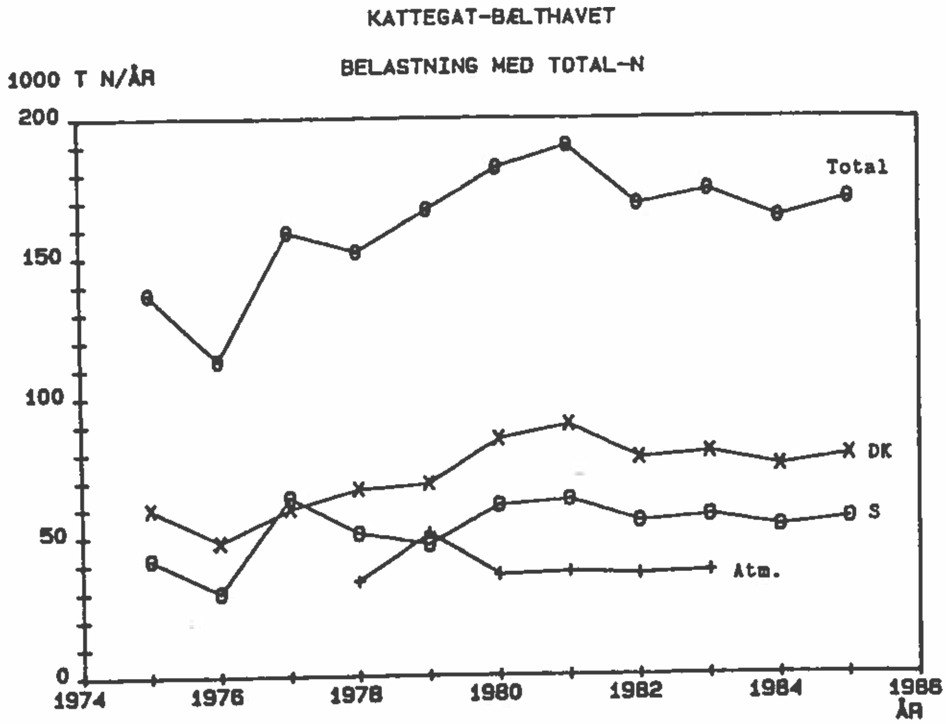

Fig. 1. Belastning af Kattegat, Øresund og Bælthavet med total-N fra Danmark, Sverige og atmosfæren 1975-85.

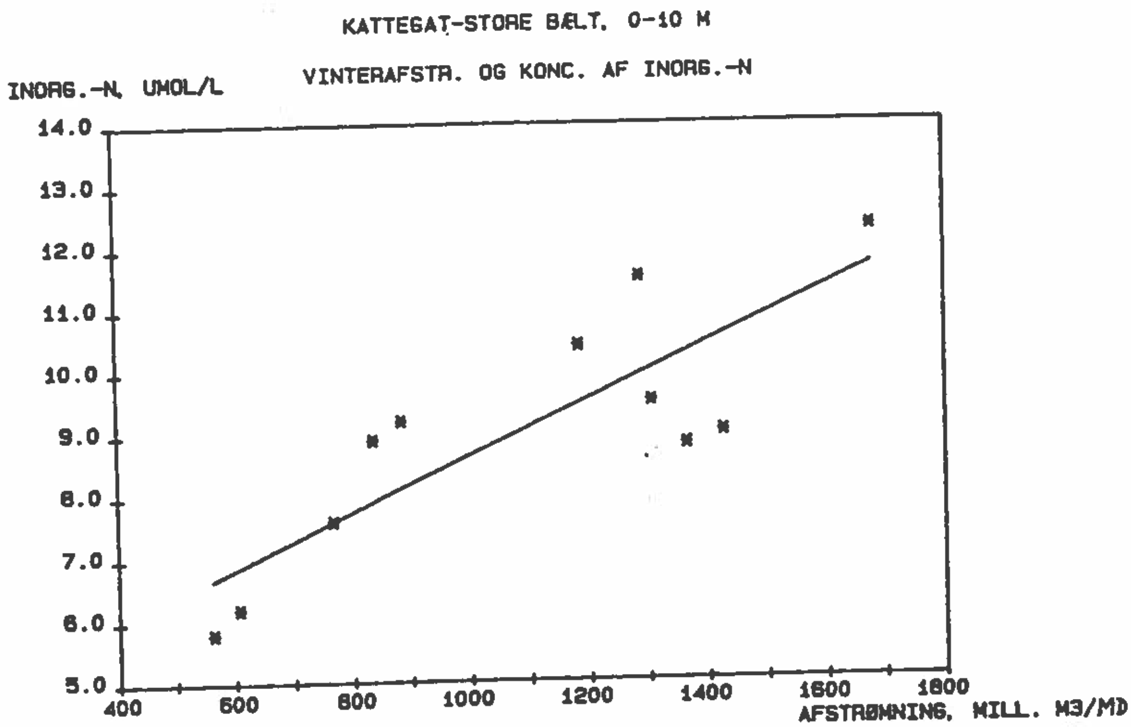

Fig. 3. Gennemsnitlig vinterkonc. (jan.-feb.) af nitrogen næringssalte (NO_3+ NO_2+ NH_4) i Kattegat-Storebælt som funktion af vinterafstrømningen (dec.-feb.) fra Danmark 1975-86, r=0,835.

opgjort af Hedeselskabet (1982 og upubl.). Ved beregningen af den samlede belastning med total-N (Fig. 1) er belastningen fra atmosfæren sat til 35.000 t N pr. år for de år, hvor denne ikke er opgjort.

Den samlede belastning med total-N i perioden 1975-85 udviser en generel stigning fra midten af 1970'erne til begyndelsen af 1980'erne og forblev i 1982-85 på et relativt højt niveau. Den gennemsnitlige belastning med total-N for perioden 1975-85 var ca. 160.000 t N pr. år. Hertil skal lægges ca. 25.000 t N pr. år fra Øst- og Vesttyskland til det sydlige Bælthav (Rydberg 1983, Edler 1984).

3. Næringssalte i Kattegat og Bælthavet

I overensstemmelse med den næsten konstante belastning med total-P findes ingen generel udvikling i koncentrationen af fosfat i Kattegat og Bælthavet i perioden 1975-85 (Miljøstyrelsen 1984).

Vinter-koncentrationen af nitogen næringssalte er derimod i samme periode steget svarende til udviklingen i afstrømningen og dermed belastningen fra land (Fig. 2a). Hovedparten af belastningen med total-N finder sted om vinteren, hvor den biologiske omsætning i vandområderne er ringe. De tilførte nitrogen næringssalte føres derfor også ud og opblandes i de åbne dele af Kattegat og Bælthavet (Fig. 3). Den mindre belastning med total-N i sommerhalvåret omsættes derimod sandsynligvis hovedsageligt i kystområderne, undtagen ved ekstremt store afstrømninger som i foråret 1983 (jvf. Fig. 5b). Imidlertid viser sommerkoncentrationen af nitrogen næringssalte i de dybere vandlag i de åbne farvande samme udvikling som vinterkoncentrationerne med en generel stigning fra midten af 1970'erne til begyndelsen af 1980'erne (Fig. 2b). Denne stigning er både vinter og sommer mest udpræget i Bælthavet og i perioden 1976-81.

4. Fytoplankton produktionen

Da nitrogen er det mest begrænsende næringssalt for fytoplanktonets primær produktion i Kattegat og Bælthavet, må produktionen være steget i takt med stigningen i vinter-koncentrationerne af nitrogen næringssalte, idet disse opbruges helt under fytoplanktonets forårsopblomstring. Stigningen i sommer-produktionen afhænger i de åbne farvande af, hvor effektivt næringssaltene i bundvandet på grund af saltholdigheds- og temperaturspringlag er afskåret fra overfladelaget. I det østlige Kattegat er denne afskæring

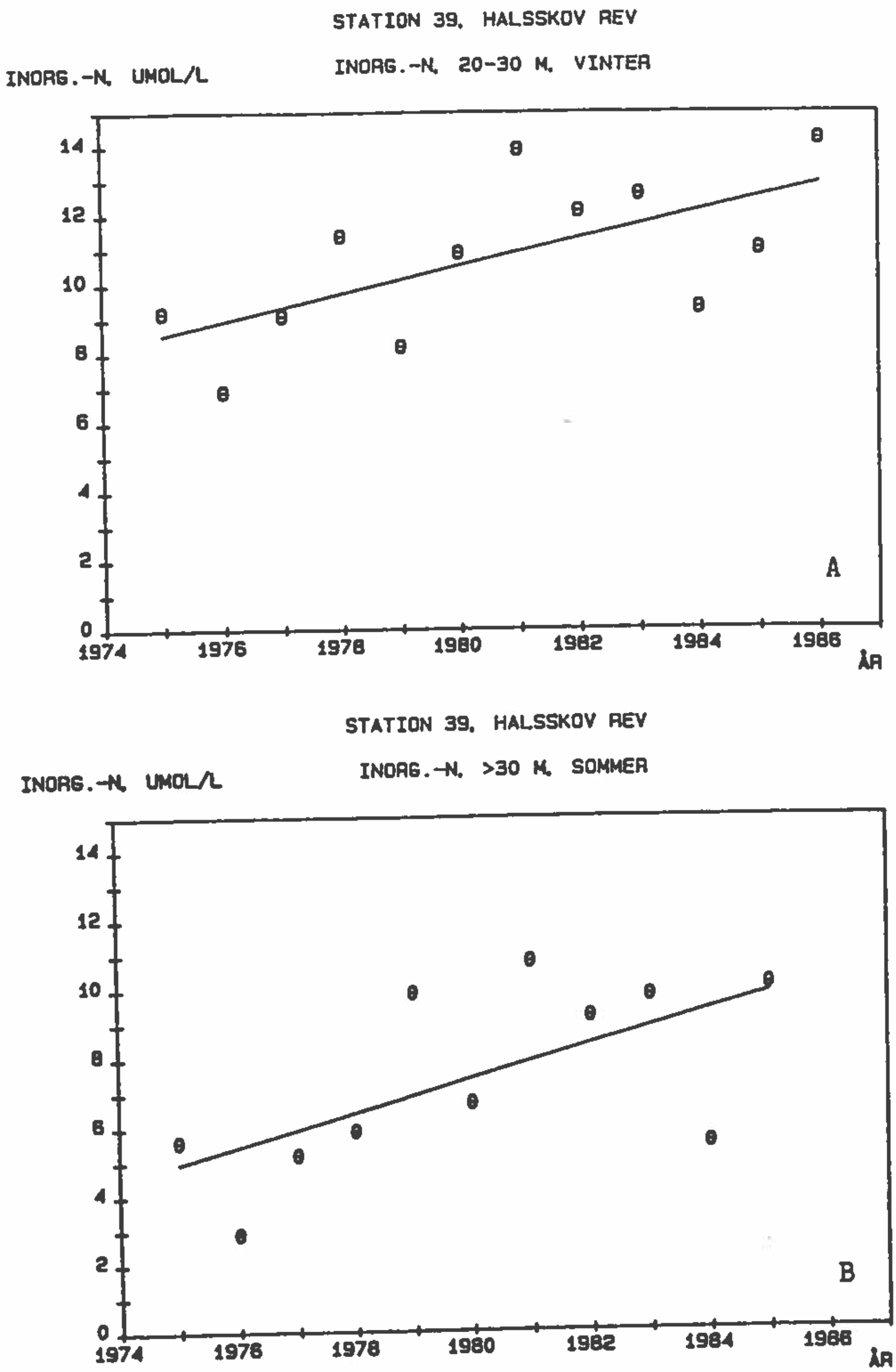

Fig. 2. Udviklingen i koncentrationen af nitrogen næringssalte (NO_3+ NO_2+ NH_4) ved station 39, Halsskov Rev, i Storebælt 1975-86. a) Gennemsnitlig vinterkonc., jan.-marts, r=0,636. b) Gennemsnitlig sommerkonc. i bundvandet, juli-sept., r=0,623.

ret så effektiv, og ingen generel stigning i sommerproduktionen er dokumenteret (Fig. 4a) (muligvis på grund af for få målinger i 1970'erne og 1980'erne). I de mere kystnære dele af Kattegat som Ålborg Bugt og Laholms Bugten synes en væsentlig stigning at have fundet sted (Ærtebjerg 1985, Edler 1984).

I Bælthavet foregår en betydelig blanding mellem overflade og bundvand, og her er der sket en væsentlig stigning i fytoplanktonets sommerproduktion fra midten af 1960'erne til begyndelsen af 1980'erne, når der ses bort fra 1976 med den lave N-belastning (Fig. 4b og Fig. 6). Sommerproduktionerne i 1980'erne er dårligt bestemt (få målinger), og kan ikke tages som udtryk for et fald i de seneste år.

5. Langtidsudvikling i belastningen med total-N

I perioden 1975-85 var belastningen med total-N og dermed koncentrationen af nitrogen næringssalte i Kattegat og Bælthavet altså først og fremmest afhængig af vandafstrømningen fra land. Imidlertid er koncentrationen af f.eks. nitrat i Kattegat generelt steget, ihvertfald siden slutningen af 1960'erne (Fig. 5), og fytoplankton produktionen i Storebælt er steget siden midten af 1960'erne (Fig. 4b og Fig. 6). Dette indikerer, at belastningen med total-N også generelt er steget uafhængigt af vandafstrømningen, når en længere tidsperiode betragtes. Dette støttes af forskellige andre opgørelser:

Edler (1984) estimerede, at belastningen med total-N til Kattegat, Øresund og Bælthavet er steget med ca. en faktor 4 fra 1930 til 1980. Schrøder (1984) estimerede, at tabet af nitrogen fra det danske landbrug til overflade- og grundvand er mere end fordoblet fra 1950 til 1980, samt at den største stigning er sket fra midten af 1960'erne og frem. Brugen af nitrogen gødning i det danske landbrug, hvorfra idag ca. 65 % af belastningen med total-N fra Danmark kommer, er steget kraftigt siden begyndelsen af 1960'erne (jvf. Fig. 6). Det er for nogle danske vandløb påvist, at transport af nitrat generelt er steget med omkring 4 % pr. år fra slutningen af 1960'erne til begyndelsen af 1980'erne (Hagebro et al. 1983, Schrøder 1984). Nitrat indholdet i grundvandet i Danmark er mere end fordoblet fra midten af 1950'erne til begyndelsen af 1980'erne (Miljøstyrelsen 1983b, 1984b). Atmosfærisk nedfald af nitrat og ammonium ved en målestation i Jylland er mere end fordoblet fra midten af 1950'erne til slutningen af 1970'erne (Oden 1976, Jørgensen 1979).

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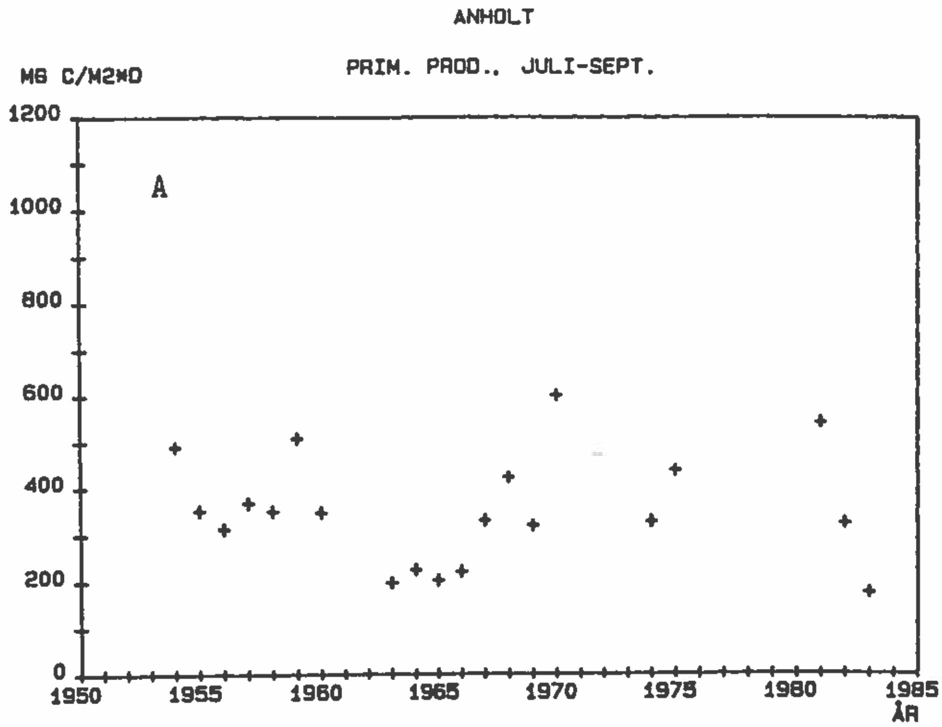
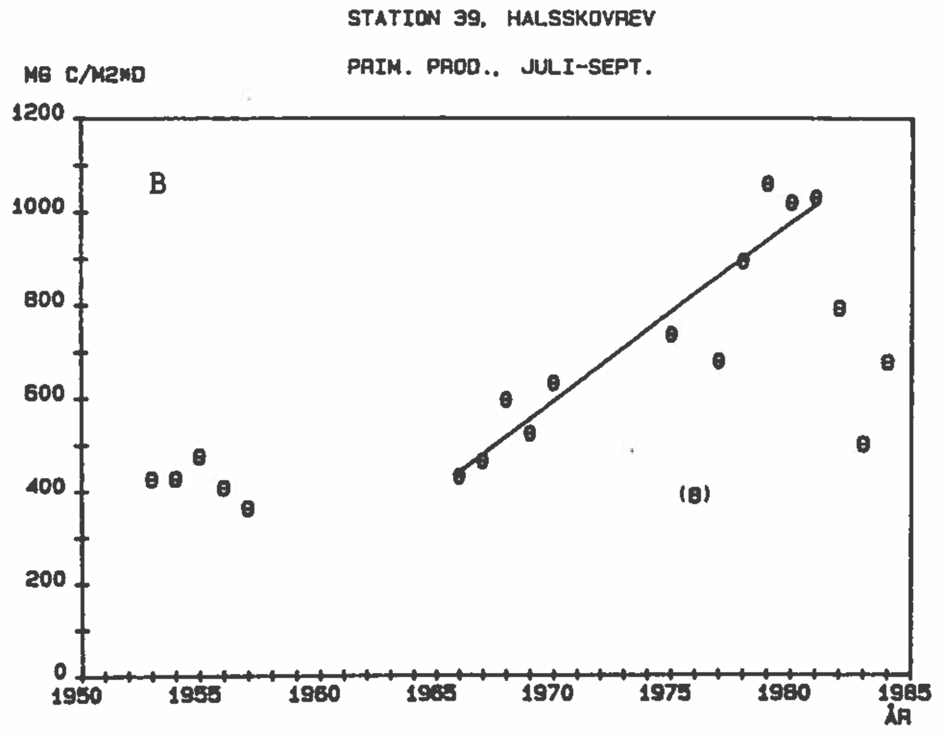




<u>Fig. 4.</u> **Fytoplanktonets gennemsnitlige sommerproduktion (juli-sept.) 1953-1984. a) Ved Anholt Nord Fyrskib og station 413, Anholt E, i det østlige Kattegat. b) Ved Halsskov Rev Fyrskib og station 39, Halsskov Rev, i Storebælt, r=0,941.**

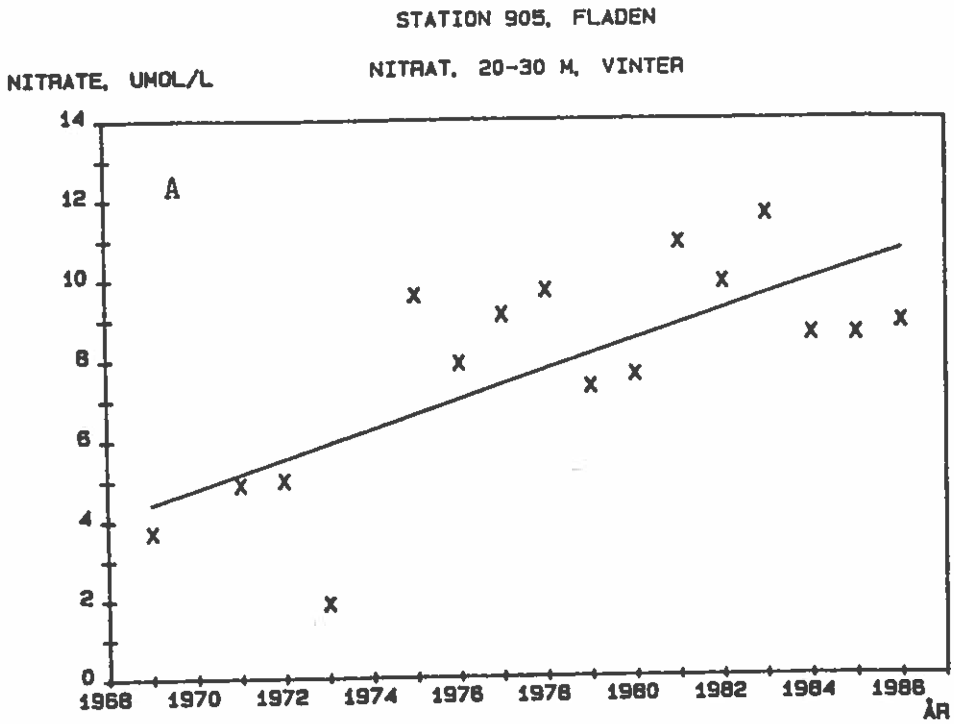
Alt dette tyder på, at belastningen af Kattegat og Bælthavet er steget betydeligt, og at stigningstakten har været størst siden 1960'erne.

Belastningen til de tilstødende havområder, Østersøen og Skagerrak/Nordsøen, har sandsynligvis udviklet sig tilsvarende. Det er derfor nærliggende at undersøge, i hvilken grad stigningen i næringssalt koncentrationerne i Kattegat og Bælthavet skyldes Østersøvand, der som en brak overfladestrøm passerer området med netto ca. 450 km^3 vand pr. år, eller Skagerrakvand, der strømmer ind i Kattegat og Bælthavet som en salt bundstrøm.

Det er påvist (HELCOM 1986), at vinterkoncentrationen af nitrat og fosfat er steget i overfladevandet i Østersøen siden midten af 1960'erne, men koncentrationerne er stadig betydeligt lavere end i Kattegat og især Bælthavet (Fig. 7). Det vil sige, at med hensyn til koncentrationerne i Kattegat og Bælthavet har det gennemstrømmende Østersøvand en fortyndende effekt, som dog muligvis er blevet mindre efterhånden som næringssalt koncentrationerne er steget i Østersøen.

Koncentrationerne af næringssalte i Skagerrak er også generelt lavere end i Kattegat og især Bælthavet (Fig. 7), og der er ikke påvist nogen generel stigning. Det kan dog ikke udelukkes, at der undertiden tilføres Kattegat vand fra Skagerrak med forhøjet indhold af næringssalte, bragt hertil fra den Tyske Bugt af Jyllandsstrømmen (f.eks. i foråret 1983, jvf. Fig. 5b og Jensen et al. 1984). Det er dog ikke sandsynligt, at dette kan forklare den generelle stigning i nitrogen næringssalte i Kattegat og Bælthavet, men at det er de lokale forhold og belastninger, der betinger koncentrationerne af næringssalte. Dette bekræftes af, at de højeste koncentrationer findes i Bælthavet, der ligger tættest ved Østersøen med dens lave koncentrationer, og hvor vandvolumenet er mindst i forhold til den modtagne direkte belastning. For fosfats vedkommende ses desuden tydeligt forhøjede koncentrationer i bundvandet i Øresund og det sydøstlige Kattegat (Fig. 7b), som sandsynligvis skyldes den store direkte belastning af Øresund med total-P.

Rydberg (1983, 1984) estimerede transporterne af total-N til og fra Kattegat, Øresund og Bælthavet og fandt, at netto indtransporten til Øresund og Bælthavet fra Østersøen og Øst- og Vesttyskland var af samme størrelse som netto udtransporten fra Kattegat til Skagerrak (ca. 130.000 t N pr.



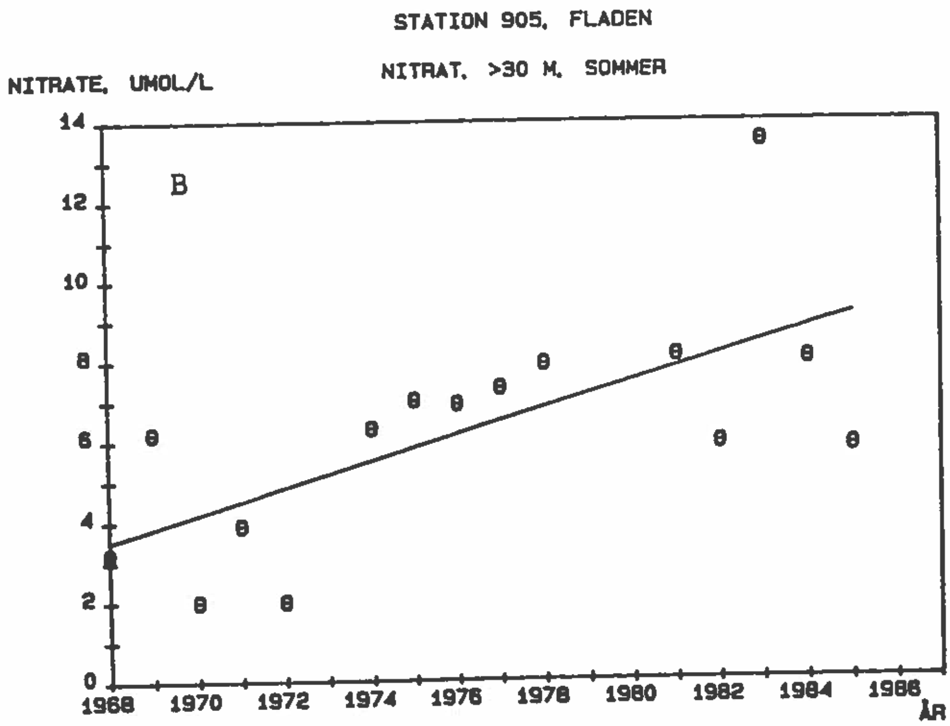


Fig. 5. Udviklingen i koncentrationen af nitrat ved station 905, Fladen, i det nordøstlige Kattegat 1968-1985. a) Gennemsnitlig vinterkonc., jan.-marts, r=0,723. b) Gennemsnitlig sommerkonc. i bundvandet, juli-sept., r=0658.

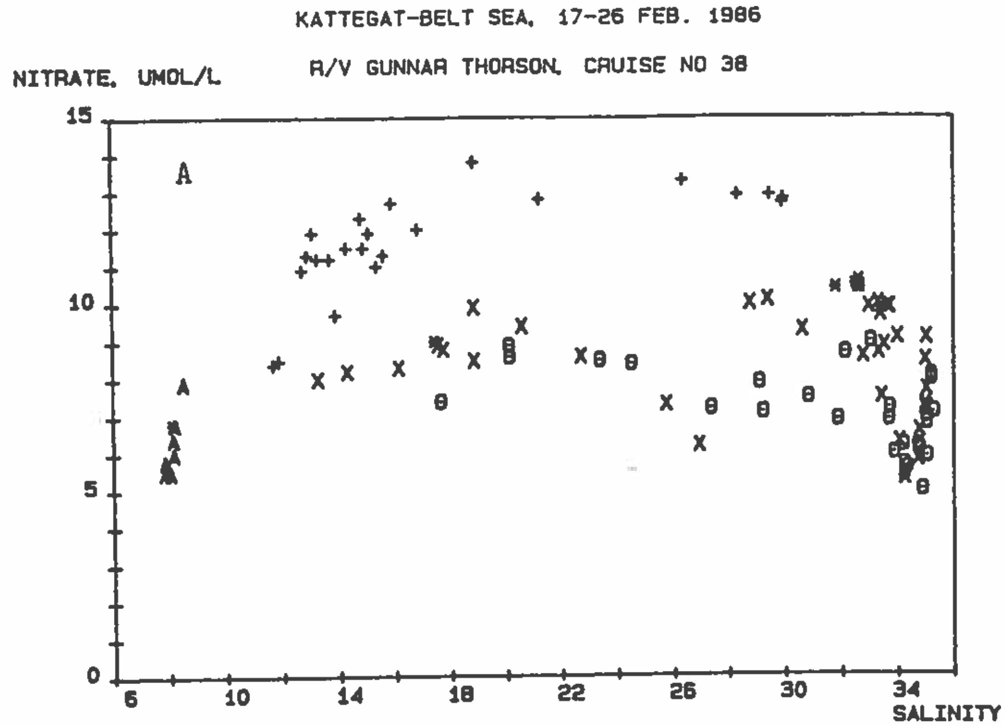
år). Det vil sige, at en mængde total-N svarende til belastningen fra Danmark, Sverige og atmosfæren må elimineres inden for Kattegat, Øresund og Bælthavet ved kraftig sedimentation og/eller denitrifikation. En stigende lokal belastning kan da medføre en stigende akkumulation i området med stigende næringssalt koncentrationer og eutrofiering til følge.

6. Eutrofieringseffekter

Som følge af den stigende eutrofiering er der i 1980'erne ved specielle meteorologiske og hydrografiske forhold flere gange observeret oxygenmangel i udbredte områder af Kattegat og Bælthavet. I september 1981 fandtes oxygenmangel i områder på tilsammen mindst 10.000 km^2, og flere tilfælde af døde bunddyr og fisk observeredes. I slutnigen af august 1983 observeredes meget lave oxygen indhold i hele Kattegat og Bælthavet i en kortere periode. I oktober-november 1985 er der observeret massedød af jomfruhummere (_Nephrops norvegicus_) og andre bunddyr i det sydøstlige Kattegat, sandsynligvis på grund af oxygen-mangel.

Foruden den generelt stigende fytoplankton produktion er usædvanlige masseforekomster samt forekomster af nye eventuelt toxiske fytoplanktonarter observeret i 1980'erne (Edler et al. 1982, Ærtebjerg & Borum 1984). I efteråret 1980 og 1981 sås masseforekomster af _Ceratium spp_. i det sydøstlige Kattegat fulgt af oxygen-mangel. _Gyrodinium aureolum_ og _Prorocentrum minimum_ observeredes for første gang i Kattegat og Bælthavet i efteråret 1981. Sidstnævnte har i de senere år gentagne gange dannet masseforekomster i fjordområderne. Gulalgen _Distephanus speculum_ observeredes i maj 1983 i masseforekomst i Bælthavet og det vestlige Kattegat med fiskedød i havbrug tilfølge. Årsagerne til disse masseforekomster er uklare, men det menes, at der er tale om en kombineret effekt af høje næringssalt koncentrationer og specielle hydrografiske og meteorologiske forhold.

Pearson et al.´s (1985) undersøgelser i 1984 af bundfaunaen i Kattegat viser sammenlignet med Petersen´s (1913) undersøgelser i 1911, at biomassen er signifikant faldet specielt i det nordvestlige Kattegat, at dominansforholdene mellem arterne er væsentligt ændret og at gennemsnitsvægten af de enkelte dyr er mindre. Årsagen til ændringerne kan være naturlige hydrografiske langtidsændringer, eutrofieringseffekter, øget bundtrawlfiskeri, eller at predationstrykket fra bundlevende fisk er ændret.



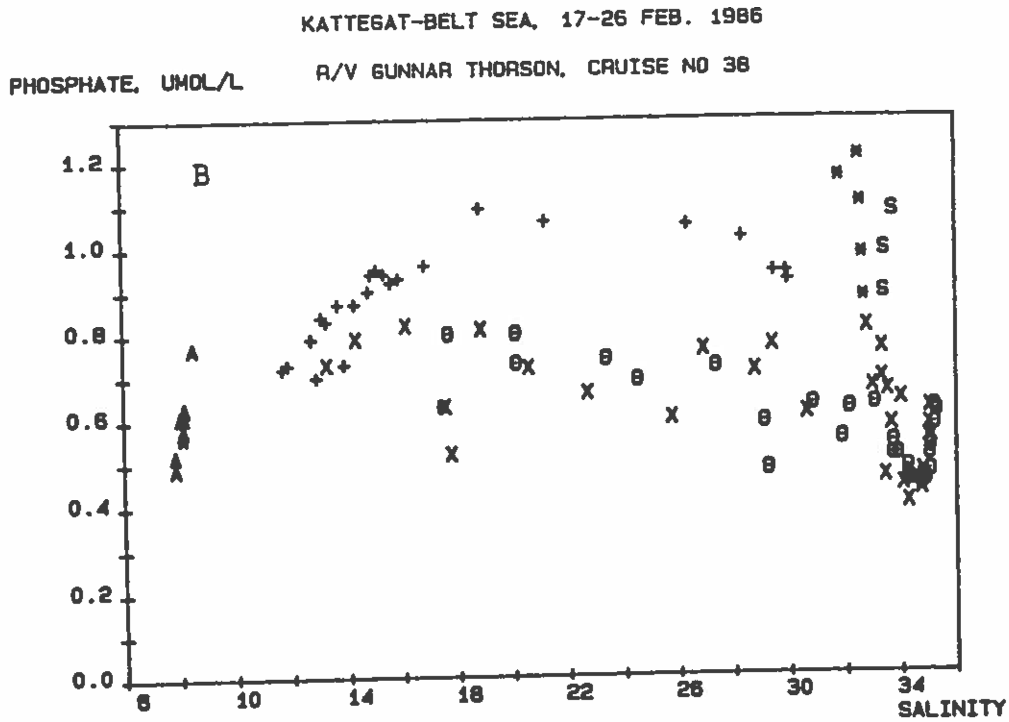


Fig. 7. Koncentrationen af a) nitrat og b) fosfat som funktion af saltholdigheden i Kattegat, Øresund, Bælthavet og Arkona Havet (vestlige Østersø) i februar 1986.

A: Arkona Havet *: Øresund +: Bælthavet x : Kattegat

O: Grænsen til Skagerrak S: Fosfat, Kattegat SE, bund.

Person peger selv på den stigende eutrofiering som hovedårsagen.

Rødspættefangsterne i Kattegat er faldet drastisk i årene 1979-81 til ca. 1/4 af niveauet i 1970'erne (Nielsen & Bagge 1985). Hvorvidt dette skyldes naturlige svingninger, eutrofieringen og indflydelse af oxygen-mangel, eller den høje infektionsprocent af parasitten Myxobolus er ikke afgjort, men flere indicier peger på, at eutrofieringen af især de kystnære områder, der er vigtige gyde og opvækstområder for mange fiskearter, har fået konsekvenser for rekrutteringen til de åbne områder. Således er mængden af fladfiskeyngel i Ålborg Bugt gået kraftigt tilbage de senere år. Det samme er kysttorskene specielt på den svenske side af Kattegat.

7. Konklusion

Sandsynligvis er belastningen af Kattegat og Bælthavet med nitrogen næringssalte fra især landbrugsområderne i de tilgrænsende lande steget betydeligt fra 1950'erne til 1980'erne. Dette har medført en stigning i eutrofieringsniveauet, idet fytoplanktonets produktion af organisk stof er steget tilsvarende, ligesom masseforekomster af planktonalger er forekommet i de senere år. Dette har igen medført et øget oxygenforbrug i bundvandet, og bunddyrdød på grund af oxygen mangel er konstateret i en række områder i 1980'erne. Foruden sådanne dramatiske hændelser er den øgede eutrofiering den sandsynlige årsag til, at bundfaunaens artssammensætning og biomasse generelt er ændret i Kattegat.

Den stigende eutrofiering ændrer således livsbetingelserne i de åbne områder for blandt andet økonomisk vigtige bundlevende fiskearter, samtidig med at den kan have forårsaget tab af vigtige gyde og opvækstområder, især i kystområderne. Dette kan være årsag til den drastiske nedgang i rødspættefiskeriet i Kattegat, og på længere sigt få alvorlige konsekvenser for erhvervsfiskeriet med væsentlig nedgang også i en række andre fiskerier på ikke-pelagiske arter som torsk, søtunge, jomfruhummere m.m.

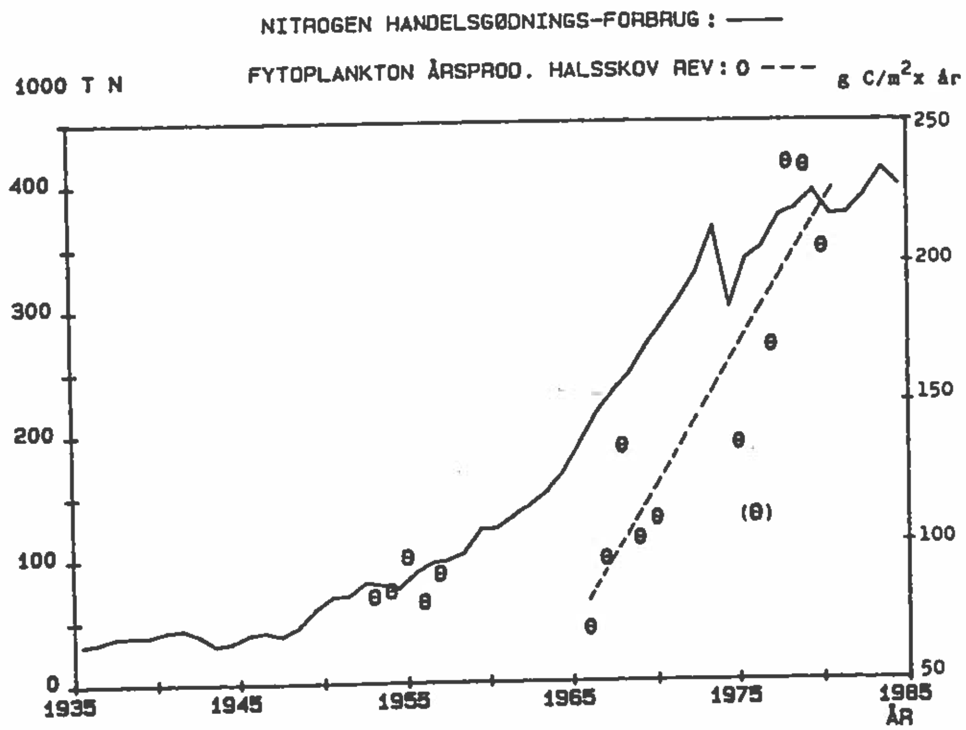


Fig. 6. Nitrogen handelsgødningsforbruget i Danmark 1935-1984 sammen med fytoplanktonets årsproduktion ved Halsskov Rev Fyrskib og station 39, Halsskov Rev, i Storebælt 1953-80, r=0,845.

8. Referencer

Edler, L. 1984. Västerhavet. I: Gödning av havsområden kring Sverige. En Kunskapsöversikt. Red: R. Rosenberg. Naturvårdsverket, PM 1808, Stockholm, pp. 74-111.

Edler, L., Ærtebjerg, G. & Granëli, E. 1982. Exceptional plankton blooms in the entrance to the Baltic Sea - the Kattegat and the Belt Sea area. ICES C.M. 1982/L:20, 6 pp.

Hagebro, C., Bang, S. & Somer, E. 1983. Nitrate load/discharge relationships and nitrate load trends in Danish rivers. Paper presented at IAMS Symposium in Hamburg, August 1983, 11 pp.

Hedeselskabet 1982. Tilstrømning til indre danske farvande 1975-1981. Det Danske Hedeselskab, Hydrometriske Undersøgelser, Slagelse, 35 pp.

HELCOM 1986. Assessment of the state of the marine environment of the Baltic Sea area 1980-1985. Baltic Marine Environment Protection Commission, Helsinki, (in press).

Jensen, T.G., Kullenberg, G. & Olsen, O.V. 1984. The Jutland Current. ICES C.M. 1984/C:6, 6 pp.

Jørgensen, V. 1979. Luftens og nedbørens kemiske sammensætning i danske landområder. 1434, Beretning fra Statens Planteavlsforsøg, pp. 633-656.

Miljøstyrelsen 1983. Opgørelse af belastningen fra land af de indre danske farvande med organsik stof, total-N og total-P. Miljøstyrelsen, København, 85 pp.

Miljøstyrelsen 1983 b. Nitrat i drikkevand og grundvand i Danmark. Miljøstyrelsen, København,

Miljøstyrelsen 1984, Iltsvind og fiskedød i 1981. Omfang og årsager. Miljøstyrelsen, København, 247 pp.

Miljøstyrelsen 1984 b. NPO-Redegørelsen. Tilførsel af kvælstof, fosfor og organisk stof til grundvand, fersk og marint overfladevand. Virkningen af denne tilførsel. Forslag til afhjælpende foranstaltninger. Miljøstyrelsen, København, 218 pp.

Nielsen, E. & Bagge, O. 1985. Preliminary investigations of 0-and I-group plaice surveys in the Kattegat in the period 1950-84. ICES C.M. 1985/G:19, 34 pp.

Oden, S. 1976. The acidity problem - An outline of concepts. Water, Air and Soil Pollution, 6, 137-166.

Pearson, T.H., Josefson, A.B. & Rosenberg, R. 1985. Petersen's benthic stations revisited. I. Is the Kattegat becoming eutrophic? J.Exp. Mar. Biol. Ecol. 92, 157-206.

Petersen, C.G.Johs. 1913. Havets bonitering. II. Om havbundens dyresamfund og om disses betydning for den marine zoogeografi. Beretn. Minist. Landbr. Fisk. Dan. Biol. Stn., Vol. 21, 1-42.

Rydberg, L. 1983. Västkustens hydrografi och närsalttransporter. Trender och klimatberoende i Östersjön och västerhavet. Röd serie 6, Oceanografiska Institutionen, Göteborgs Universitet, 29 pp.

Rydberg, L. & Sundberg, J. 1984. On the supply of nutrients to the Kattegat. Oceanografiska Institutionen, Göteborgs Unviersitet, rep. no 44, 14 pp.

Schrøder, H. 1984. Udviklingen i kvælstoftabene fra dansk landbrug og konsekvenserne for vandmiljøet. Vandkvalitetsinstituttet, Hørsholm, 186 pp.

Ærtebjerg, G. 1985. Årsager til iltsvind i Kattegat og Bælthavet. Vand & Miljø, 2, 59-62.

Ærtebjerg, G. & Borum J. 1984. Exceptional phytoplankton occurrences and related events in Danish waters 1981-83. ICES, Special meeting on the causes, dynamics and effects of exceptional marine blooms and related events, B 8, 10 pp.