

Fig. 7 — Comparison of the relative-humidity calculations obtained from the dry-bulb and wet-bulb temperature measurements of flight 51 (points) and the model estimates of relative humidity based only on shipboard observations (line)

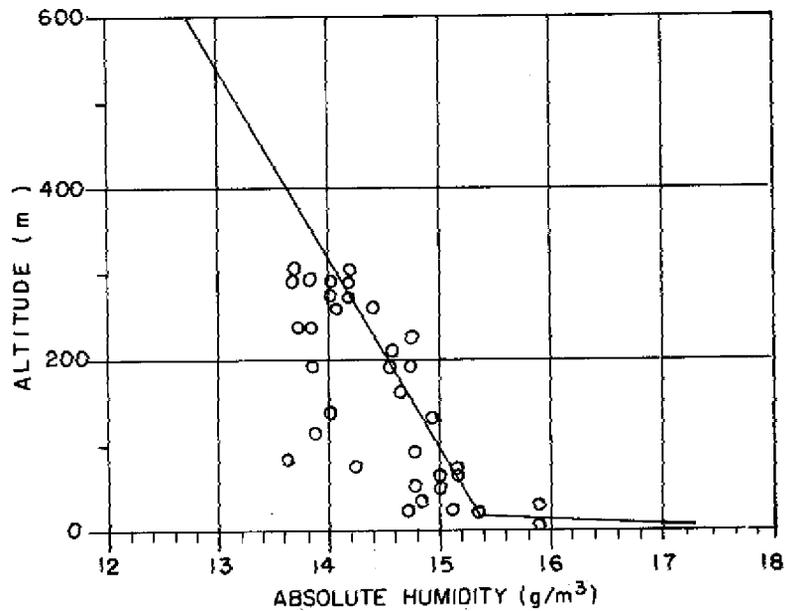


Fig. 8 — Comparison of the absolute-humidity calculations obtained from the dry-bulb and wet-bulb temperature measurements of flight 51 (points) and the model estimates of the absolute humidity based only on shipboard observations (line)

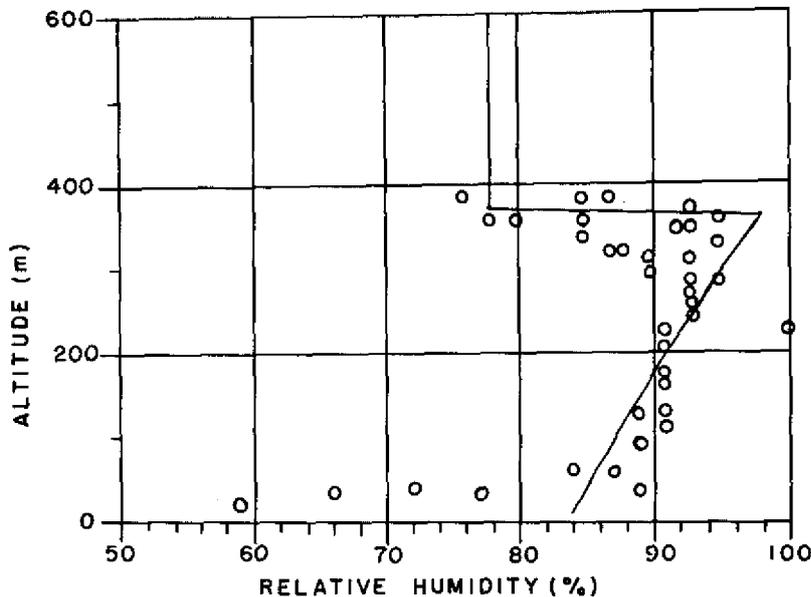


Fig. 9 — Comparison of the model estimates of a relative-humidity profile which predicted a dry upper air layer from surface measurements made at the time of flight 42 (line) and the kite-balloon measurements of relative humidity based on dry-bulb and wet-bulb temperatures (points)

Not all predictions are this reassuring however. Figure 10 is a plot of air-temperature measurement from flight 21 near the Gulf Stream. The air-temperature measurements (although the most basic and simplest of the measurements to make) seem to give absurd data. The problem results from the time required for the measurement, that is for the kite balloon to climb to altitude and then to be reeled back to deck height. Because the before and after calibration measurements of these data are accurate, the conclusion is that the atmosphere is changing rapidly. Thus during the 50 minutes required by the kite-balloon system to climb to maximum altitude and then to be reeled to the surface the assumption of a stationary atmosphere does not hold. Consequently a static model cannot duplicate these rapid changes in air-mass characteristics. It may well be that the model prediction is closer to a snapshot of the profile at one instant of time while the profile is changing from one form to the other than can be obtained from a measuring system which is slow relative to the time response of the nonstationary atmosphere.

The model predictions and the in-situ data taken over all of the flights of the 1977 EOMET cruise at the positions shown in figure 5 were compared as follows to allow the user some idea of the error bars on the model estimates. A calculation was made of the RMS error of the model predictions at 100 altitudes equally spaced below the maximum altitude obtained on the flight relative to the least-square third-order polynomial fits to the measurement data at the same 100 altitudes. This analysis shows that the worst-case air-temperature RMS error by the model was 3.7°C (for flight 3) but that the average RMS error for all flights was 0.7°C . A similar analysis shows that the model had an overall average RMS error

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20. ABSTRACT (Continue on reverse side if necessary and identify by block number)		
<p>A model is presented which estimates the vertical profiles of air temperature, relative humidity, and water-vapor density within the lowest 2 kilometers of the marine atmosphere. Inputs to the model are a subset of standard marine surface observations, obtainable from ships throughout the world ocean. The 1977 EOMET cruise of the USNS <i>Hayes</i> provided a set of kite-balloon measurements of air temperature, relative humidity, and water-vapor density together with standard marine surface observations throughout the North Atlantic Ocean and the Mediterranean Sea. The model estimates of the</p> <p style="text-align: right;">(continued)</p>		

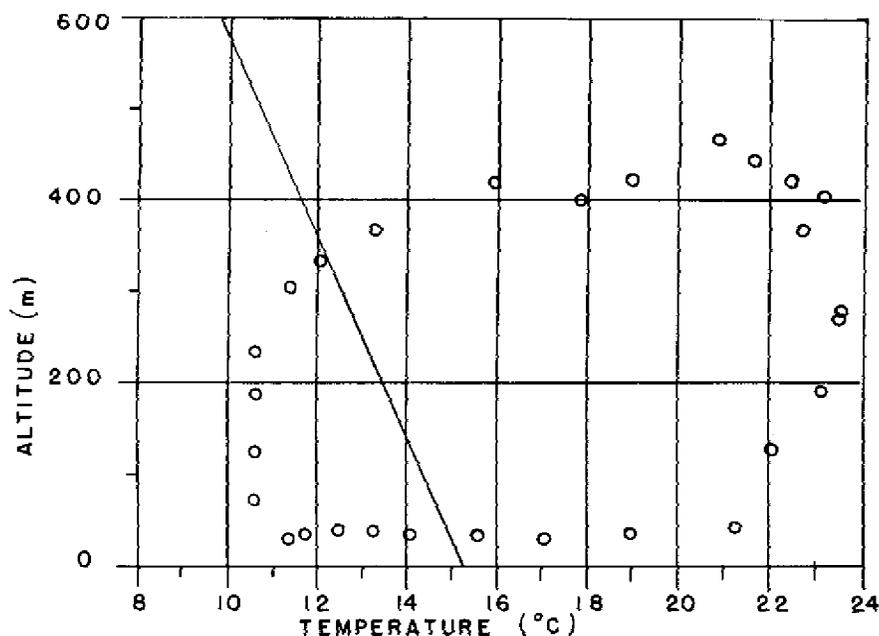


Fig. 10 — Example of an air-temperature profile (points) (obtained in a rapidly changing air mass from a sounding system that takes about 50 minutes to make a complete ascent and descent) that cannot be used to represent a stationary profile against which to test a static model (line). The data in this plot were obtained from flight 21.

in the relative-humidity estimates of about 20%. These errors however include not only inaccuracies of the model but all types of errors which might exist in this experiment, such as nonstationarity of the air mass.

The model test was also analyzed to see if the errors of the model depended on altitude. One would expect that, since the model is based on measurements at the surface, the model accuracy would decrease with altitude, so that the RMS errors between data points and model predictions would increase with altitude.

Table 1 shows the RMS differences between data points at the various measurement altitudes and the model predictions of the data at the same altitudes. These differences are averaged over all of the flights in 100-meter altitude bands. The RMS errors indeed tend to increase with altitude.

Table 1 — Altitude Dependence of the RMS Errors of the Model

Altitude Interval (m)	Number of Observations in Altitude Interval	RMS Error		
		Air Temperature (°C)	Relative Humidity (%)	Absolute Humidity (g/m ³)
0-100	480	1.01	9.54	1.06
100-200	296	1.21	9.71	0.94
200-300	365	1.24	9.86	1.06
300-400	341	1.34	11.93	1.40
400-500	155	2.77	20.47	2.15
over 500	81	1.65	16.36	1.68

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APPENDIX
PROGRAM LISTING

```

FUNCTION RELHUM(Z,IQ,ISHIP,TALE,VAPLD)
C*****
C 1.IDENTIFICATION NAME: RELHUM
C 2.TITLE: METEOROLOGICAL PROFILE ESTIMATES FROM SHIP OBS.
C 3.AUTHOR: STUART G. GATHMAN
C 4.ORGANIZATION: NRL
C 5.DATE: JULY 1977
C 6.ROUTINES CALLED: ABORT,ALT,PALF,VAPPR,C10,ZZERO,QZ,QSTAR,
C TBZ,THSTR,FMLNT,ZCON,SMIXR,IR,TEMP,VP,TABS,TCENT,LIMIT
C*****
DIMENSION ISHIP(12),IFOG(14),IDRIZ(17),ICLR(6)
IPW=ISHIP(1)
ILCL=ISHIP(2)
IASL=ISHIP(3)
IAT=ISHIP(4)
ICLHT=ISHIP(5)
IDPT=ISHIP(6)
ISSI=ISHIP(7)
IWS=ISHIP(8)
IVLS=ISHIP(9)
IWS=ISHIP(10)
IIVIS=ISHIP(11)
ICA=ISHIP(12)
DATA(IFOG(1),I=1,14)/10,11,12,28,40,41,42,43,44,45,
1 46,47,48,49/
DATA(IDRIZ(1),I=1,10)/14,15,16,20,24,50,51,52,53,54/
DATA(IDRIZ(1),I=11,17)/55,56,57,58,59,60,61/
DATA(ICLR(1),I=1,6)/0,1,2,3,4,5/
C
C
C CHECK FOR INDICATIONS OF FOG AND IF FOUND GO TO 100
C
IF(IVIS.LE.94) GO TO 100
IF(IIVIS.EQ.1) GO TO 100
DO 20 J=1,14
20 IF(IPW.EQ.IFOG(J)) GO TO 100
C
C
C CHECK FOR INDICATIONS OF DRIZZLE AND IF
C FOUND GO TO STATEMENT NUMBER 300
C
DO 30 J=1,14
30 IF(IPW.EQ.IDRIZ(J)) GO TO 300
C
C
C CHECK FOR CLEAR WEATHER(IE NO INDICATION OF PREC.) AND
C IF FOUND GO TO STATEMENT NUMBER 600
C
DO 40 J=1,6
40 IF(IPW.EQ.ICLR(J)) GO TO 600
C
C
C IF THE PROGRAM GETS TO THIS POINT WE MUST ABORT.
C
50 CALL ABORT(Z,IQ,IDPT,IAT,RH)
REL HUM=RH
TALE=TCENT(IAT)

```

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```
VAPLD=2.165*RELHUM*VAPPR(TAIR)/(TAIR+273.15)
CALL LIMITS(RH,VAPLD,TAIR,IQ)
RETURN
```

```
C
C
C THIS PART OF THE PROGRAM IS FOR TIMES OF FOG INDICATIONS.
```

```
100 RELHUM=100.
    TAIR=TCENT(IAT)
    VAPLD=216.5*VAPPR(TAIR)/(TAIR+273.15)
110 IQ=10
    IF(Z.GT.400.) IQ=6
    IF(Z.GT.1000.) IQ=2
    RETURN
```

```
C
C
C THIS PART OF THE PROGRAM IS FOR NON PRECIPITATION TIMES.
```

```
600 IF(ILCL.EQ.6) GO TO 700
    IF(ILCL.EQ.4) GO TO 700
    IF(ILCL.EQ.7) GO TO 700
    IF(IASD.LT.0) GO TO 800
    IF(IASD.EQ.0) GO TO 900
```

```
C
C
C THE STABLE ATMOSPHERE
```

```
777 CONTINUE
    IF(Z.LT.20.) GO TO 1000
    GO TO 1100
```

```
C
C THIS PART OF THE PROGRAM IS FOR UNSTABLE CASES.
C DETERMINE THE DEPTH OF THE SUPERADIABATIC LAYER.
```

```
800 IQ=10
    DIF=ABS(FLOAT(IASD)/10.)
    IF(DIF.GT.0.5) GO TO 850
    ZSA=10.*DIF
    GO TO 870
850 ZSA=11.92+9.69*ALOG(DIF)
```

```
C
C
C NEXT DETERMINE IF Z IS ABOVE OR BELOW THE TOP OF THE
C SUPERADIABATIC LAYER.
```

```
870 IF(Z.LE.ZSA) GO TO 2000
```

```
C
C
C FOR ZSA < Z < ZH, THETA AND Q ARE CONSTANT WRT ALTITUDE
C AND EQUAL TO THE VALUES DETERMINED AT Z=ZSA.
```

```
1500 THSA=THZ(IWS,IWS,IQ,ISS,T,IAT,ZSA)
    QSA=QZ(IWS,IWS,IQ,ISS,T,IAT,ZSA)
```

```
C
C
C FIND THE LIFTING CONDENSATION LEVEL FROM VALUES OF THETA
C AND Q DETERMINED AT THE TOP OF THE SUPERADIABATIC LAYER.
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C
      ZC=ZCON(THSA,QSA,ZSA)
C
C
C DETERMINE THE CLOUD AMOUNT TO SEE IF THERE IS A
C TRANSITION ZONE EXISTING.
C
1810 CONTINUE
      IF(ICA.LT.2) GO TO 1600
      ZH=ZC
      IF(Z.GE.ZH) GO TO 1700
1800 E=VF(QSA,Z)
      TAIR=TEMP(THSA,Z)-273.15
      ESAT=VAPPR(TAIR)
      VAPLD=216.5*E/(TAIR+273.1)
      RELHUM=100.*E/ESAT
      CALL LIMITS(RELHUM,VAPLD,TAIR,IQ)
      RETURN
C
C
C THE IN CLOUD CASE
C
1700 RELHUM=100.
      TAIR=TEMP(THSA,Z)-273.15
      VAPLD=216.5*VAPPR(TAIR)/(TAIR+273.15)
      IF(Z.LT.2000.) RETURN
      T2000=TEMP(THSA,2000) - 273.15
      Q2000=SMIX(Z,T2000)
      Q=Q2000*4E6/Z**2
      ESAT=VAPPR(TAIR)
      E=VF(Q,Z)
      RELHUM=100*E/ESAT
      VAPLD=216.5*E/TEMP(THSA,Z)
      RETURN
C
C
C THIS PART OF THE PROGRAM IS FOR A CLOUDLESS CLOUDLAYER.
C
1600 ZH=0.8*ZC
      IF(Z.GT.ZC) GO TO 1900
      IF(Z.LT.ZH) GO TO 1800
      DELZ=ZC-ZH
      TZH=TEMP(THSA,ZH)
      TZ=TZH-0.006*(Z-ZH)
      QZH=QSA
      TZC=TZH-0.006*DELZ
      QZC=0.78*SMIXR(ZC,TZC-273.15)
      THSA=TH(TZ,Z)
      QSA=QZH*(QZC-QZH)*(Z-ZH)/DELZ
      GO TO 1800
C
C
C THIS IS THETA AND R CALCULATED FOR A SUPER A LAYER
C
2000 X=THZ(11WS,IWS,IQ,ISST,IAT,Z)
      R=QZ(11WS,IWS,IQ,ISST,IDPT,Z,IAT)
      TC=TEMP(X,Z)-273.15

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```

VTEST=VP (R,Z)
VAPTST=VAPPR (TC)
RELHUM=100.*VTEST/VAPTST
TAIR=TC
VAPLD=216.5*VTEST/(TAIR+273.15)
ZO=ZZERO (IWS,IWS,10)
FL=FMLNT (IWS,IWS,ISST,IAT,10)
FCN=(Z+ZO)/FL
IF (FCN.LE.0.03) GO TO 2010
IQ=IQ-2
2010 CONTINUE
CALL LIMITS (RELHUM,VAPLD,TAIR,IQ)
RETURN

C
C
C   STABLE ATMOSPHERE, Z> SHIP MAST HEIGHT.
C
1100 IQ=9
    CLBAS=FLOAT (ICLHT)
    CTOP=CLBAS+500
    IF (Z.GE.CTOP) GO TO 1105
    IF (Z.GE.CLBAS) GO TO 1104
    THS=TH (TABS (IAT),20.)+(Z-20.)*2./(CLBAS-20.)
    QS=SMIXR (20.,TCENT (IDPT))
    GO TO 1110
1104 THS=TH (TABS (IAT),20.)*2.+(Z-CLBAS)*.006
    QS=SMIXR (20.,TCENT (IDPT))
    GO TO 1110
1105 THS=TH (TABS (IAT),20.)+5+.00001*(Z-CTOP)
    QS=SMIXR (20.,TCENT (IDPT))*CTOP*CTOP/Z**2
1110 CONTINUE
    E=VE (QS,Z)
    TAIR=TEMP (THS,Z)-273.15
    ESAT=VAPPR (TAIR)
    VAPLD=216.5*E/TEMP (THS,Z)
    RELHUM=100.*E/ESAT
    IF (RELHUM.LE.100.) RETURN
    RELHUM=100.
    VAPLD=216.5*ESAT/TEMP (THS,Z)
    RETURN

C
C
C   STABLE ATMOSPHERE,Z< SHIP MAST HEIGHT.
C
1000 THSP=TH (TABS (IAT),20.)
    THW=TH (TABS (ISST),0.)
    QSP=SMIXR (20.,TCENT (IDPT))
    QWT=.98*SMIXR (1.,TCENT (ISSI))
    THSA=THW+Z*(THSP-THW)/20.
    QSA=QWT+Z*(QSP-QWT)/20
    IQ=10
    GO TO 1000

C
C
C   NEUTRAL ATMOSPHERE CASE
C
900 IQ=10

```

Model for Estimating Meteorological Profiles From Shipboard Observations

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*Atmospheric Physics Branch
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December 26, 1978



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```

IF(Z.LT.20.) GO TO 975
THSA=TH(TABS(ISSI),0.)
QSA=SMIXR(20.,TCENT(IDPT))
950 ZC=121*(TCENT(IAT)-TCENT(IDPT))
IF(Z.LE.ZC) GO TO 1800
GO TO 1810
975 TAIB=FLOAT(1SS1)/10.
QSEA=0.98*SMIX(0.,TCENT(1SS1))
QMAST=SMIX(20.,TCENT(IDPT))
Q=(QMAST-QSEA)*Z/20.+QSEA
E=VF(Q,Z)
ESAT=VAPPR(TAIR)
RELHUM=100.*E/ESAT
VAPLD=216.5*E/(TAIR+273.15)
RETURN
C
C
C THIS PART OF THE PROGRAM TREATS LIGHT NON FREEZING PREC.
C OR DRIZZLE. RELATIVE HUMIDITY IS ASSUMED EVERYWHERE TO BE
C THAT OF THE SHIP.
C
C 100 RELHUM=100.*VAPPR(TCENT(IDPT))/VAPPR(TCENT(IAT))
C GO TO 110
C
C THIS IS AN APPROXIMATION FOR THE CLOUDLESS "CLOUD LAYER"
C AFTER MALKUS (1958)
C
C 1900 RELHUM=78.
C IQ=7
C TAIR=TEMP(THSA,Z)-273.15
C VAPLD=168.9*VAPPR(TAIR)/(TAIR+273.15)
C RETURN
C
C THIS PART OF THE PROGRAM IS FOR THE STRATUS CLOUD CASE
C
C 700 CONTINUE
C IF(Z.LT.20.) GO TO 1000
C CLBAS=FLOAT(ICLHT)
C CTOF=CLBAS+500
C IF(Z.GE.CTOF) GO TO 705
C IF(Z.GE.CLBAS) GO TO 704
C THS=TH(TABS(IAT),20.)+(Z-20.)*Z/(CLBAS-20.)
C PC=TH(TABS(IAT),20.)*2
C TC=TEMP(PC,CLBAS)-273.15
C QB=SMIXR(20.,TCENT(IDPT))
C QC=SMIXR(CLBAS,TC)
C QS=QB+(Z-20.)*(QC-QB)/(CLBAS-20.)
C GO TO 1110
C 704 THS=TH(TABS(IAT),20.)*2+.006*(Z-CLBAS)
C QS=SMIXR(Z,TEMP(THS,Z)-273.15)
C GO TO 1110
C 705 THS=TH(TABS(IAT),20.)*5+.00001*(Z-CTOF)
C QS=SMIXR(CTOF,TEMP(THS,CTOF)-273.15)*CTOF*CTOF/Z**2
C GO TO 1110
C END

```

Note: The Malkus citation on this page is Ref. 3 in the list at the end of the main text.

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```

C*****
C
C SUBROUTINES AND FUNCTIONS FOR USE WITH RELHUM FUNCTION.
C
C
C     SUBROUTINE ABOBT (Z,IQ, IDPT, IAT,RH)
C
C     THIS SUBROUTINE SIMPLY TRANSMITS THE SHIPBOARD MEASUREMENTS
C     TO OTHER ALTITUDES AND DECREASES IQ AS THE PREDICTIONS
C     BECCME LESS ACCURATE.
C
C     DEWPT=FLOAT (IDPT)/10.
C     TEMP=FLOAT (IAT) / 10.
C     RH=100.*VAPPR (DEWPT) /VAPPR (TEMP)
C     IQ=9
C     IF (Z.GT. 100.) IQ=6
C     IF (Z.GT. 500.) IQ=3
C     IF (Z.GI. 1000.) IQ=0
C     RETURN
C     END
C
C
C     FUNCTION ALT (P)
C
C     THIS FUNCTION COMPUTES ALTITUDE IN METERS FROM PRESSURE
C     IN MILLIBARS. THIS FORMULA IS A FIT TO THE N.A.C.A.
C     STANDARD LOWER ATMOSPHERE DATA: SMITHSONIAN MET.TABLES N63
C     LIST (1968).
C
C     IF (P.LE. 1013) GO TO 10
C     ALT=0
C     RETURN
C 10  IF (P.LE. 958.) GO TO 20
C     ALT=9.09*(1013.-P)
C     RETURN
C 20  ALT=7850*ALOG (1021.38/P)
C     RETURN
C     END
C
C
C     FUNCTION PALT (Z)
C
C     THIS IS THE INVERSE OF FUNCTION ALT (P) WHERE ALTITUDE,Z
C     IS IN METERS AND PALT IS THE PRESSURE IN MILLIBARS AT Z.
C
C     IF (Z.LT. 500.) GO TO 10
C     PALT=1021.38*EXP (-1.2739E-4*Z)
C     RETURN
C 10  PALT=1013.-55.*Z/500.
C     RETURN
C     END
C
C
C     FUNCTION VAPPR (T)
C
C     THIS IS AN APPROXIMATION TO THE GOFF-GRATCH TEMPERATURE,
C     SATURATED VAPOR PRESSURE OVER LIQUID WATER FORMULA.

```

The Smithsonian tables cited are Ref. 10, and the List citation is Ref. 10.

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```

C      IT WAS DESCRIBED BY RICHARDS (1971) AND WIGLEY (1974).
C      IN THIS FUNCTION T IS IN DEG. C AND VAPPR IN MB.
C
      T4=1.0-373./(T+273.)
      R1=13.3185
      R2=1.976
      R3=0.6445
      R4=0.1299
      R0=1013.25
      VAPPR=R0*EXP(R1*T4-R2*T4**2-R3*T4**3-R4*T4**4)
      RETURN
      END

C
C
C      FUNCTION C10(IWS,IWS,IQ)
C
C      THIS FUNCTION CALCULATES THE DRAG COEFFICIENT. IF WIND
C      SPEED AT 10 METERS IS MEASURED, DEACONS SUGGESTED FORM
C      IS USED: ROLL (1965), P.161. IF NO WIND IS MEASURED A
C      CONSTANT VALUE IS USED.
C
      IF(IWS.NE.-1) GO TO 10
      IQ=IQ-1
      C10=2E-3
      RETURN
10 C10=1.1E-3+0.04E-3*FLOAT(IWS)
      RETURN
      END

C
C
C      FUNCTION ZZERO(IWS,IWS,IQ)
C
C      THIS FUNCTION CALCULATES THE DYNAMIC ROUGHNESS, ZZERO
C      IN METERS AS A FUNCTION OF WIND SPEED FOLLOWING CHARNOCK
C      (1955) AND EXPRESSES FRICTION VELOCITY IN TERMS OF THE
C      DRAG COEFFICIENT AND MEASURED WIND SPEED.
C
      IF(IWS.NE.-1) GO TO 10
      IQ=IQ-2
      ZZERO=1.E-5
      RETURN
10 ZZERO=C10(IWS,IWS,IQ)*3.35E-4*FLOAT(IWS)**2
      RETURN
      END

C
C
C      FUNCTION QZ(IWS,IWS,IQ,ISSI,LDPT,Z,IAT)
C
C      THIS FUNCTION CALCULATES THE MIXING RATIO AT ALT. Z IN AN
C      ATMOSPHERE WHICH OBEYS A LOG - LINEAR RELATIONSHIP:
C      ROLI (1965), P. 237.
C
      T0=FLOAT(ISSI)/10
      Q0=0.98*SMIXR(0,T0)
      QSR=QSTAR(LDPT,Z,IWS,IWS,ISSI,IQ)
      Z0=ZZERO(IWS,IWS,IQ)
      X=(Z+Z0)/Z0

```

The Richards, Wigley, Roll, and Charnok citations are Refs. 7, 8, 1, and 9 respectively.

20. Abstract (Continued)

meteorological parameters are compared with simultaneously obtained in-situ measurements for 56 flights. In some cases agreement is good, but when the profiles change during the 50 minutes required for a kite-balloon measurement, comparison with the static model becomes invalid.

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```

PL=FMLNT (IWS,IIWS,ISST,IAT,IQ)
QZ=Q0+QSE*(ALOG(X)+4.8*Z/PL)
RETURN
END

```

C
C

```

FUNCTION THZ (IIWS,IWS,IQ,ISST,IAT,Z)

```

C
C
C
C

```

THIS FUNCTION CALCULATES THE POTENTIAL TEMPERATURE AT
ALTITUDE Z IN AN ATMOSPHERE WHICH OBEYS A LOG - LINEAR
RELATIONSHIP: ROLL (1965) ,P.237.
T0=273.15+FLOAT (ISST)/10.
TH0=TH (T0,0.)
Z0=ZZERO (IIWS,IWS,IQ)
X=(Z+Z0)/Z0
PL=FMLNT (IWS,IIWS,ISST,IAT,IQ)
TSTAR=THSTR (IAT,Z,IIWS,IWS,ISST,IQ)
THZ=TH0+TSTAR*(ALOG(X)+4.8*Z/PL)
RETURN
END

```

C
C

```

FUNCTION QSTAR (IDPT,Z,IIWS,IWS,ISST,IQ)

```

C
C
C
C
C

```

THIS IS THE "FRICTION MIXING RATIO" USED IN QZ AND
CALCULATED USING THE APPROXIMATIONS OF THE BULK AERODYNAMIC
METHOD: ROLL (1965), P.252 ,272.

```

```

QA=SMIXR (10.,TCENT (IDPT))
Q0=.98*SMIXR (0.,TCENT (ISST))
CA=C10 (IWS,IIWS,IQ)
FK=0.38
QSTAR=SQRT (CA) * (QA-Q0) /FK
RETURN
END

```

C
C

```

FUNCTION THSTR (IAT,Z,IIWS,IWS,ISST,IQ)

```

C
C
C
C
C

```

THIS IS THE "FRICTION POTENTIAL TEMPERATURE" USED IN THE
CALCULATION OF THZ USING THE APPROXIMATION OF THE
BULK AERODYNAMIC METHOD: ROLL (1965) P.252,272.

```

```

TA=273.15+FLOAT (IAT)/10.
T0=273.15+FLOAT (ISST)/10.
CA=C10 (IWS,IIWS,IQ)
FK=0.4
THSTR=SQRT (CA) * (TH (TA,10.)-TH (T0,0)) /FK
RETURN
END

```

C
C

```

FUNCTION FMLNT (IWS,IIWS,ISST,IAT,IQ)

```

C
C
C
C

```

THIS IS AN APPROXIMATION FOR THE MIXING LENGTH USING
ROLL (1965), P. 144,252.

```

The Roll citation is Ref. 1.

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```

IF (IWS.NE.-1) GO TO 10
PMLNT=-100
IQ=IQ-2
RETURN
10 TO=273.+FLOAT(ISST)/10.
TA=273.+FLOAT(IAT)/10.
CA=C10 (IWS,IWS,IQ)
X=TO*SQRT(CA)*0.2646*FLOAT(IWS)**2
Y=9.8*.38*(TH(TO,0.)-TH(TA,10.))
PMLNT=-1.*X/Y
RETURN
END

C
C
FUNCTION ZCON(TASA,RSA,ZSA)
C
C THIS FUNCTION CALCULATES THE LIFTING CONDENSATION LEVEL
C WHERE TASA IS THE POTENTIAL TEMPERATURE IN KELVIN AT ALT
C ZSA AND RSA IS THE MIXING RATIO AT THIS LEVEL IN G/KG.
C THIS FORMULATION IS AN EMPIRICAL FIT TO DATA IN THE
C SMITHSONIAN MET. TABLES P.328.
C
TSA=TEMP (TASA,ZSA)
ESA=VP (RSA,ZSA)
FK=1./286
X=EXP (FK*ALOG (TSA))
TC=732.02-150.41*(ALOG 10 (X)-ALOG 10 (ESA))
TC=TC+7.21*(ALOG 10 (X)-ALOG 10 (ESA))**2
TC=TC+273.15
PZ=1000*EXP (-FK*ALOG (TASA/TC))
ZCON=ALT (PZ)
RETURN
END

C
C
FUNCTION SMIXR(Z,T)
C
C CALCULATES THE SATURATION MIXING RATIO IN G/KG AT HEIGHT Z
C AND TEMPERATURE T (C).
C
SMIXR=0.622*VAPPR(T)/(PALT(Z)-VAPPR(T))
RETURN
END

C
C
FUNCTION TH(T,Z)
C
C CONVERTS TEMPERATURE (K) TO POTENTIAL TEMPERATURE (K) AT
C ALTITUDE Z.
C
TH=T*EXP (0.286*ALOG (1000./PALT (Z)))
TH=TH
RETURN
END

C
C
FUNCTION TEMP(TE,Z)

```

The Smithsonian tables cited on this page are Ref. 10.

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```

C
C   CONVERTS POTENTIAL TEMPERATURE (K) AT ALTITUDE Z (M) TO
C   TEMPERATURE (K)
C
TEMP=TE/(EXP(0.286*ALOG(1000./PALT(Z))))
RETURN
END

C
C   FUNCTION VP (R,Z)
C
C   CONVERTS MIXING RATIO (G/KG) AT ALTITUDE Z (M) TO
C   VAPOR PRESSURE (MB).
C
VP=PALT(Z)*R/(-6.22+R)
RETURN
END

C
C   FUNCTION TABS(I)
C
C   CONVERTS AN INPUT TEMPERATURE INTEGER TO TEMPERATURE (K).
C
TABS=273.15+FLOAT(I)/10.
RETURN
END

C
C   FUNCTION TCENT(I)
C
C   CONVERTS AN INPUT TEMPERATURE INTEGER TO TEMPERATURE (C).
C
TCENT=FLOAT(I)/10.
RETURN
END

C
C   SUBROUTINE LIMITS(RH,V,T,IQ)
IF(RH.LE.65.) GO TO 10
IF(RH.LE.100.) RETURN
RH=100.
GO TO 20
10 RH=65.
20 IQ=IQ-4
V=2.165*RH*VAPP(T)/(T+273.15)
RETURN
END

```

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MODEL FOR ESTIMATING METEOROLOGICAL PROFILES FROM SHIPBOARD OBSERVATIONS

INTRODUCTION

Large quantities of shipboard meteorological observations made throughout the world ocean over an impressive length of time are limited in nature and accuracy. But they provide a marine climatological data base which would be too expensive to duplicate to obtain other types of meteorological data subsequently needed but not specifically included in the data base. One example of subsequent need is the statistical characterization of detailed humidity and temperature profiles within the lowest 2 kilometers above the ocean surface. The relatively small group of radiosonde data taken by various weather ships and researchers is unfortunately sparse in both a geographical and a statistical sense. There exists therefore a need to invent a model which will provide an estimated profile of humidity and temperature when given the standard set of shipboard observables. Profiles obtained with such a model will be inferior to those observed directly by the use of radiosondes and are not intended to replace them, only to supplement them in areas where sufficient profile measurements do not exist.

The chief advantage is that profiles can be generated where no profiles have been measured and the results can be believed to within a certain degree of accuracy. One area of interest in which such a model could be useful is the transmission of optical energy over various nonhorizontal paths. Along these paths humidity estimates are valuable both for its effect on molecular absorption of optical energy and in its effect on the growth of aerosols.

This report describes such a model and a method and data to test this model. The test method can also be used on future models to ascertain their usefulness.

THE EMPIRICAL MODEL

The model divides the standard shipboard weather observations into essentially four cases:

- fog,
- light precipitation,
- heavy precipitation, and
- no precipitation.

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Characterization by the model of all but the last of these items is trivial in that present values on the ship are estimated to exist at higher altitudes. In the last case, however, more elaborate predictive methods are used.

An integer is delivered to the main calling program which gives a rough estimate of the quality of the particular value being calculated. When many assumptions are made, the quality integer (IQ) becomes low or even negative. Values of 10 are considered reasonable estimates.

The stability of the marine boundary layer is an important input to the model. The measure of this parameter that is available to the model is the air/sea temperature difference.

In the no-precipitation case the following distinct classes of profiles are modeled:

- Unstable atmosphere with cloud cover greater than 25%,
- Unstable atmosphere with cloud cover less than 25%,
- Neutral atmosphere with cloud cover greater than 25%,
- Neutral atmosphere with cloud cover less than 25%,
- Stable atmosphere with a stratus deck, and
- Stable atmosphere with no stratus deck.

The particular class of model chosen depends on the sign of the air/sea temperature difference, which determines the stability, and the level and/or type of cloud cover reported. The specific characterization of each of the profiles from any of these classes depends also on the specific values of the other shipboard observables.

In the modeling, potential temperature and mixing ratio are chosen as the particular meteorological pair of variables to describe the vertical thermal and vapor-loading characteristics of the atmosphere because of their insensitivity to adiabatic processes. One limiting feature of this model is that it is designed to describe only the layer of the ocean atmosphere at or below 2 kilometers; thus attempts to use the model outside of this limitation may result in severe errors.

The output of the model is relative humidity, absolute humidity, and air temperature (all quantities which are used directly in propagation calculations). In making these estimates, many conversions are required between various meteorological variables. Thus a number of conversion functions and subroutines are required by the model to convert easily from one set of parameters to another. These subroutines and functions (shown in the listing of the program in the Appendix) use for the most part standard meteorological relationships.

Unstable Atmosphere With Cloud Cover Less Than 25%

In modeling the class of profiles identified as unstable atmosphere with cloud cover less than 25% four levels are considered. Starting from the layer closest to the sea, they are

Since the area is clear (total cloud amount is less than 25%), then a transition layer is considered to start at an altitude of about 80% of the lifting condensation level. The transition layer starts at the top of the homogeneous layer. In the homogeneous layer both potential temperature and the mixing ratio at all altitudes remain constant at the values obtained for the top of the superadiabatic layer.

According to measurements by Malkus [3], the relative humidity in the subsiding cloudy layer is 78%. Therefore in the transition layer the model slightly increases the potential temperature linearly and adjusts the mixing ratio so that 78% relative humidity is achieved at the altitude calculated as the lifting condensation level. Throughout the less cloudy layer the model sets the potential temperature to that calculated at the lifting condensation level and correspondingly decreases the mixing ratio to keep the relative humidity constant at 78%.

Unstable Atmosphere With Cloud Cover Greater Than 25%

Figure 2 is a diagram of the same unstable atmosphere but with clouds. For altitudes below the transition layer the model is identical to that for the cloudless case. The model for the cloud case, however, keeps both the potential temperature and the mixing ratio constant with respect to altitude throughout the transition zone. Above the lifting condensation level when the relative humidity is 100%, the potential temperature remains constant and the mixing ratio is reduced just enough to keep the relative humidity at 100%.

The cloud layer extends from the lifting condensation level to about 2 kilometers where the moist marine layer is usually capped by an inversion which divides dry warm air on top from the cool moist air of the marine layer. Above the 2-km level, temperature decreases at the standard lapse rate, and moisture is allowed to decrease rapidly toward the zero level.

Neutral Atmosphere

The neutral atmosphere, profiled in Fig. 3, is similar to the unstable atmosphere in both the cloudy and noncloudy conditions except that no superadiabatic layer exists. The potential temperature of the atmosphere is considered adiabatic from the sea surface up to the lifting condensation level. The potential temperature measured at the ship is used as the potential temperature throughout the lower atmosphere from the sea surface to the lifting condensation level in the cloudy case and up to the bottom of the transition zone in the subsiding cloudless case.

The mixing ratio is interpolated linearly between values corresponding to relative humidity of 98% at the sea surface and that measured at the assumed mast height of 10 meters. Above 10 meters and below the lifting condensation level the mixing ratio is set constant to that measured on the ship.

- The superadiabatic layer,
- The homogeneous layer,
- The transitional layer, and
- The cloudless cloud layer.

The extent of each of these layers varies according to the individual input parameters. The unstable model is based on the description of the moist marine layer in the trade-wind zones sketched by Roll [1].

Figure 1 is a conceptual representation of profiles of the potential temperature θ and the water-vapor mixing ratio r for this class of atmosphere. The first layer above the ocean surface is called the superadiabatic bottom layer. In this layer the potential temperature and mixing ratio obey a log-linear formulation, with the defining parameters being related to the shipboard observations. The height of this layer above the sea surface is calculated by an empirical formula fitted to a series of measurements made by Brocks [2]:

$$Z_{sa} = 11.92 + 9.69 \ln(|T_{air} - T_{sea}|), \quad 5 < T_{air} - T_{sea},$$

$$= T_{air} - T_{sea}, \quad 0 < T_{air} - T_{sea} \leq 5,$$

where Z_{sa} is in meters and T_{air} and T_{sea} are the air temperature and sea-surface temperature in degrees Celsius.

The meteorological parameters of θ and r at any altitude Z within the superadiabatic layer are computed from

$$\theta(Z) = \theta(0) + \theta_* \left[\ln \frac{Z + Z_0}{Z_0} + 4.8 \frac{Z}{L} \right]$$

and

$$r(Z) = r(0) + r_* \left[\ln \frac{Z + Z_0}{Z_0} + 4.8 \frac{Z}{L} \right].$$

In these formulas $\theta(0)$ and $r(0)$ are the values of potential temperature and mixing ratio computed from measurements of the temperature of the sea surface. The variables θ_* and r_* are computed from

$$\theta_* = \frac{\sqrt{C_{10}}}{0.38} [\theta(10) - \theta(0)]$$

and

$$r_* = \frac{\sqrt{C_{10}}}{0.38} [r(10) - r(0)],$$

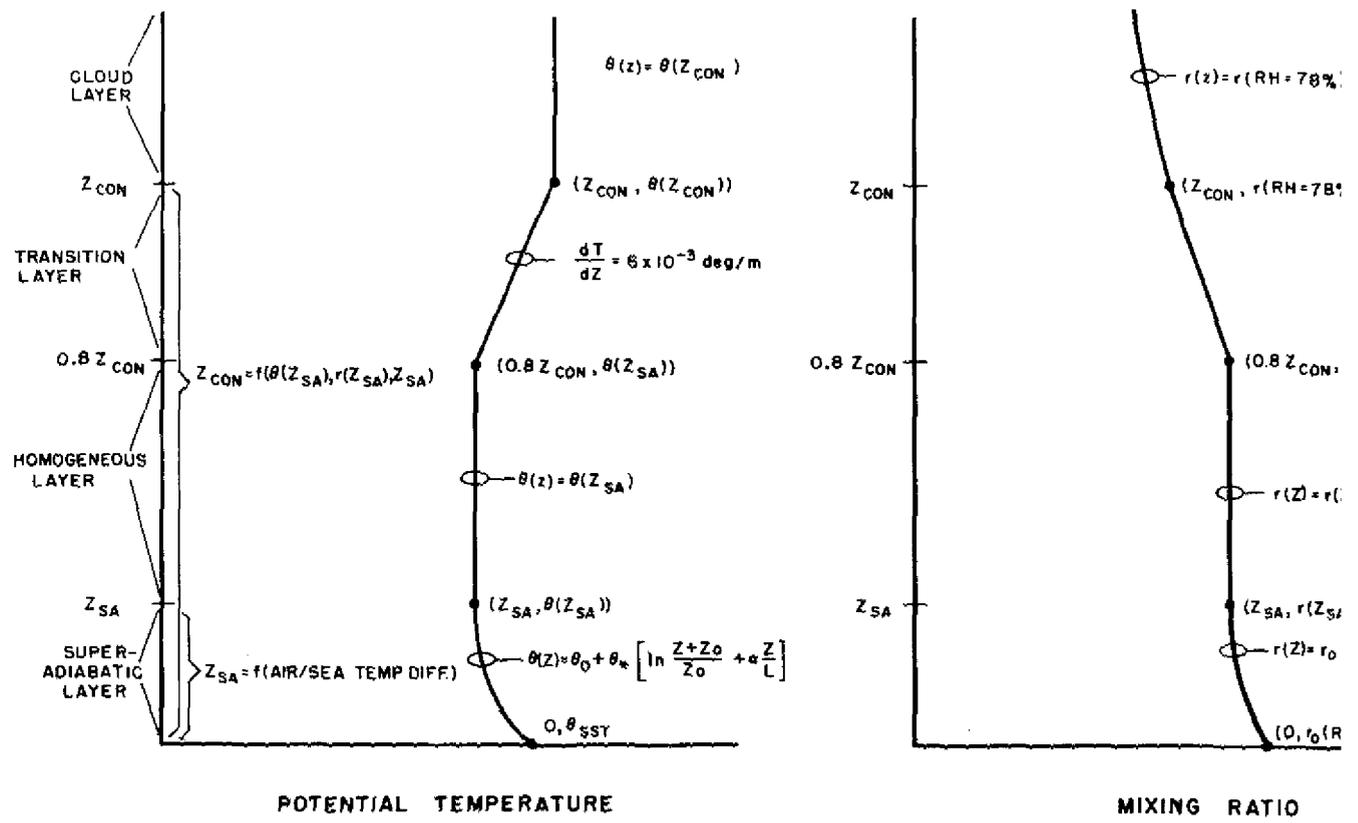


Fig. 1 — Profiles of the characteristic potential temperature θ and the water-vapor mixing ratio r for the unstable atmosphere less than 25%. Key points on these curves are set by appropriate shipboard observations. (Z_{CON} is the lifting condensat the sea-surface potential temperature.)

where

$$C_{10} = [1.1 + 0.04 \mu(10)] \times 10^{-3},$$

with $\theta(10)$, $r(10)$, and $u(10)$ being the values of potential temperature, mixing ratio, and real wind speed obtained at the nominal mast height of 10 meters. Values of Z_0 are obtained as follows: If the

real wind speed is measured, then

$$Z_0 = 3.3 \times 10^{-4} C_{10} u(10)^2,$$

and if the real wind speed is not measured, then

$$Z_0 = 10^{-5}.$$

Finally the values of the so-called stability length L are calculated from

$$L = \frac{-T_0 \sqrt{C_{10}} u(10)^2}{9.8 \times 0.38 [\theta_A(0) - \theta_A(10)]},$$

where the sea-surface temperature T_0 as well as potential temperatures $\theta_A(0)$ and $\theta_A(10)$ are in absolute temperature units. If values of the wind speed are not measured, L is set at -100 meters. When the relative humidity is to be estimated at an altitude less than Z_{sa} , then it is calculated by use of θ and r from these equations. If Z is greater than Z_{sa} , the values of potential temperature θ and mixing ratio r at Z_{sa} are used to estimate these same quantities at higher levels.

The height of the lifting condensation level Z_{con} must next be calculated from measurements available on the ship. The method used in this model is based on the solution of the equation for adiabatic lifting:

$$e/T^{1/0.286} = e_{sc}/T_c^{1/0.286},$$

where T is the temperature calculated at the top of the superadiabatic bottom layer, e is the vapor pressure calculated at the same point, T_c is the temperature at the condensation level, and e_{sc} is the saturation vapor pressure at the temperature of the condensation level. The potential temperature at the top of the superadiabatic layer is known, because both the temperature and the altitude are known. Also known is that the potential temperature remains constant with respect to altitude changes for adiabatic processes. Therefore at the condensation level the pressure p_c can be calculated from

$$\theta = T_c (1000/p_c)^{0.286}.$$

The model then uses the NACA standard low-level atmosphere to convert from pressure to altitude.

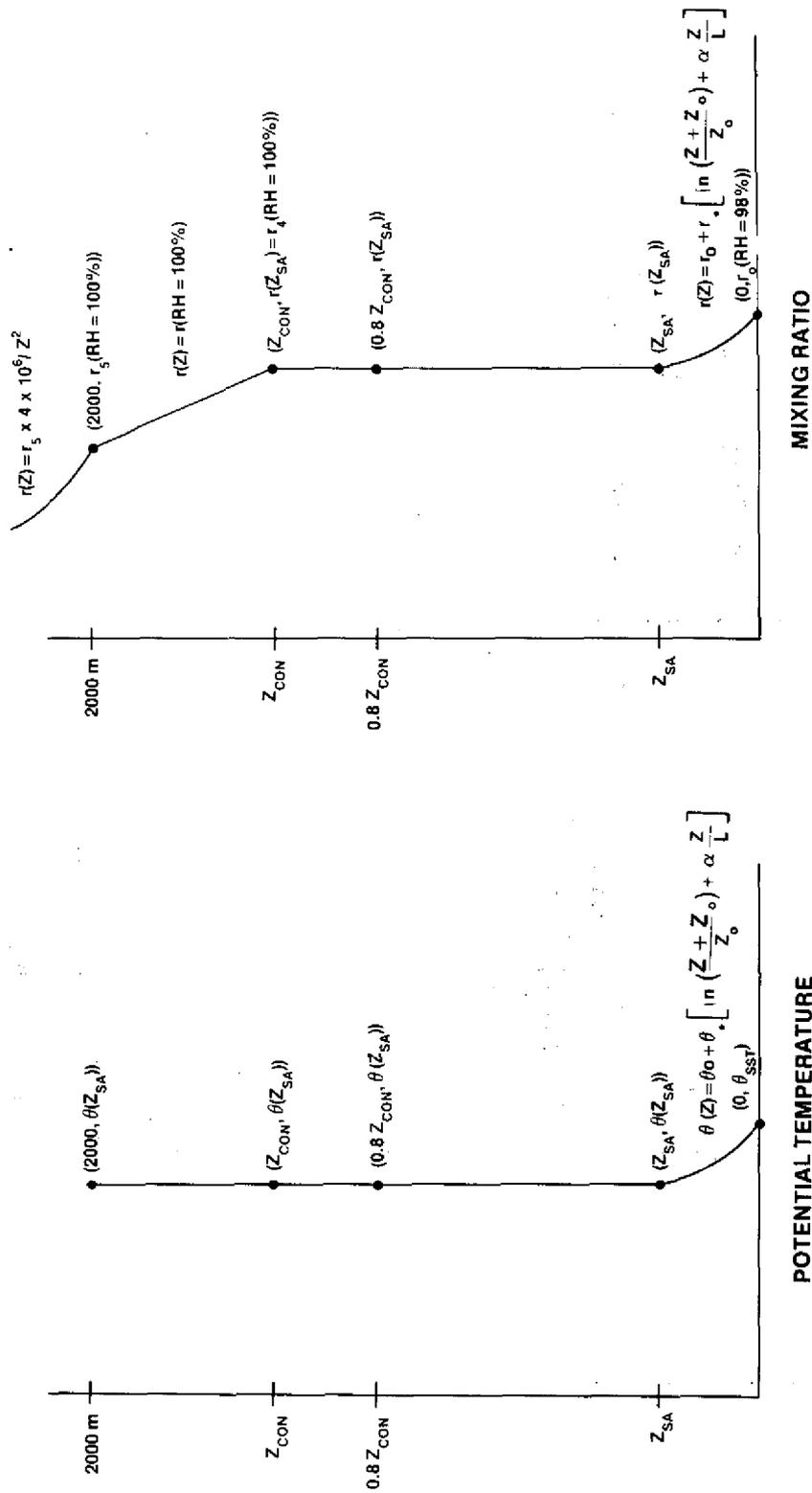


Fig. 2 — Profiles of the characteristic potential temperature and the water-vapor mixing ratio for an unstable atmosphere with cloud cover greater than or equal to 25%

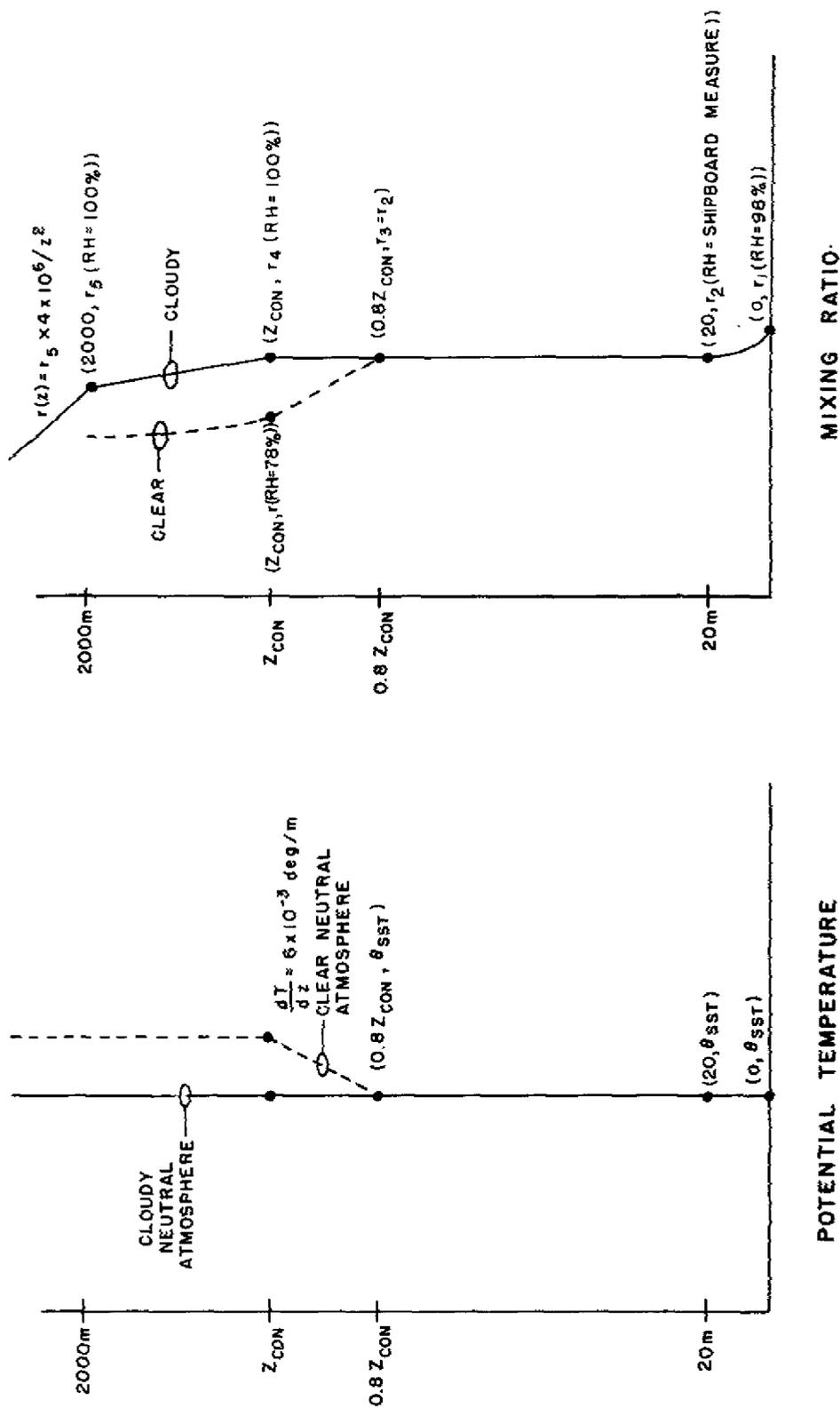


Fig. 3 — Profiles of the characteristic potential temperature and the water-vapor mixing ratio for a neutral atmosphere. The dashed lines above $0.8 Z_{CON}$ apply when the cloud cover is less than 25%; the solid lines above $0.8 Z_{CON}$ apply when the cloud cover is 25% or greater.

Stable Atmosphere

The basis for the model estimates of the stable atmosphere is a series of tethered balloon flights made aboard the USNS *Hayes* during the 1975 fog cruise [4]. If a stratus deck is observed, the model constructs the following atmospheric thermal and water-vapor structure from four additional shipboard-measurable inputs: the sea surface temperature, the dew-point temperature and air temperature at 20 meters, and an estimate of the cloud-base altitude.

The potential-temperature structure is estimated in the following way and as shown in Fig. 4. The atmosphere is divided into four layers separated by four reference levels: the sea surface, 20 meters (the height of shipboard measurements), the cloud-base level (an input to the model), and the cloud top (assumed to be 500 meters above the cloud base). In the lowest layer a linear interpolation is assumed between the potential temperatures of the sea surface itself and the air at ship height. In the second layer a linear interpolation is made between the potential temperature at the 20-meter level to that at the cloud base level, which is assumed to be that of the ship plus 2°C. Over the next 500 meters an increase of 3°C of potential temperature is assumed. Finally above the cloud layer the potential temperature is allowed to increase at a small rate.

Similarly the mixing ratio is constructed in a piecewise linear fashion in the same four levels. In the lowest layer a vapor pressure of 98% of the saturation vapor pressure at the sea-surface temperature is assumed at the sea surface. The mixing ratio is then linearly interpolated between the mixing ratio r_1 calculated at the sea surface and the mixing ratio r_2 calculated at the mast height. When a stratus cloud is known to exist, then saturation is known to be achieved at the cloud base and therefore the mixing ratios r_3 and r_4 which are necessary for saturation are assumed both at the cloud top and at the cloud base. Linear interpolations are then made between these various fixed points. Above the cloud top the mixing ratio is allowed to drop off at a $1/Z^2$ rate.

The stable-atmosphere no-stratus-cloud profiles are identical to the stable-atmosphere stratus-cloud profiles except that the mixing-ratio profile is assumed constant from the ship-mast height to 500 meters above the calculated lifting condensation level.

IMPLEMENTATION OF THE MODEL

The model is in the form of a Fortran function: RELHUM (Z, IQ, ISHIP, TAIR, VAPLD). The outputs of the model are the relative humidity in percent which becomes the functional value itself, the air temperature represented by parameter TAIR, and the water-vapor load of the atmosphere represented by the variable VAPLD. The integer output variable IQ is a quality index which is roughly related to the trustworthiness of a particular prediction as described above. The model requires only two inputs: the altitude Z (meters) at which the meteorological values are desired to be known and an integer array ISHIP which contains 12 integers from the standard shipboard weather reports.

The elements of array ISHIP are defined as follows:

- ISHIP (1) is an integer between 0 and 99 (coded in accordance with Ref. 5) which describes the present weather situation;

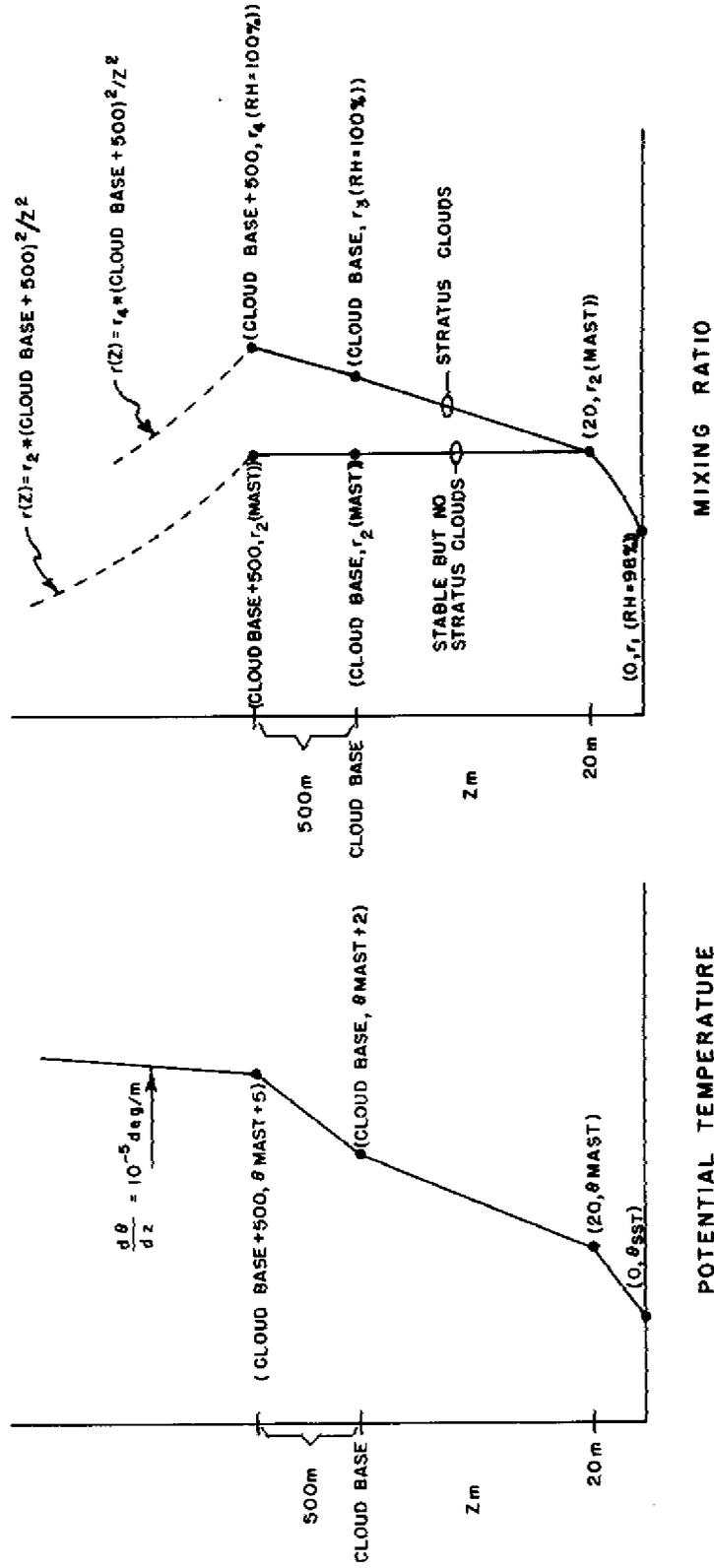


Fig. 4 — Profiles of the characteristic potential temperature and the water-vapor mixing ratio for stable atmospheres with a stratus-cloud deck and with no stratus-cloud deck

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- ISHIP (2) is a parameter between 0 and 10 which describes the lower cloud types (ILCL);
- ISHIP (3) is a signed integer between -999 and +999 which is 10 times the air/sea temperature difference;
- ISHIP (4) is a signed integer between -999 and +999 which is 10 times the air temperature (IAT);
- ISHIP (5) is a noncoded integer which is the estimated height of the cloud base of the lowest cloud layer in meters (ICLHT).
- ISHIP (6) is a signed integer between -999 and +999 which is 10 times the dew-point temperature at deck level (IDPT);
- ISHIP (7) is a signed integer between -999 and +999 which is 10 times the measured sea-surface temperature (ISST);
- ISHIP (8) is an integer between 0 and 99 which is the wind speed in knots;
- ISHIP (9) is a integer between 90 and 99 (coded in accordance with Ref. 5) which represents the visibility;
- ISHIP (10) is an integer which if equal to 0 indicates that wind speed was measured;
- ISHIP (11) is an indicator of the horizontal visibility at ship level, with 0 indicating that visibility was measured and 1 indicating fog;
- ISHIP (12) is a integer between 0 and 9 (coded in accordance with Ref. 5) which indicates the total cloud amount (ICA).

TEST OF THE MODEL

A test of the accuracy of this model within the marine boundary layer has been devised by comparing the model-produced estimates with measured profiles of temperature and water vapor. Field data suitable for this test was obtained from the EOMET cruise of the USNS *Hayes* during May and June 1977. During this cruise kite-balloon soundings were made, producing many observations within the marine boundary layer. Comparisons between the measured parameters of air temperature, relative humidity, and vapor density and those same parameters calculated from the model which used the appropriate standard ship-board measurements taken on the ship itself provide a useful test of any model of a boundary-layer profile.

The locations of the soundings are shown in Fig. 5 superimposed on the ship track, which covered portions of the North Atlantic and the Mediterranean Sea. Although the experiment provided over 1700 valid parameter sets obtained with the NRL boundary-layer sonde [6], the data are limited in duration at any one location and therefore do not form a climatologically complete set of measurements. It is nevertheless an excellent data base on

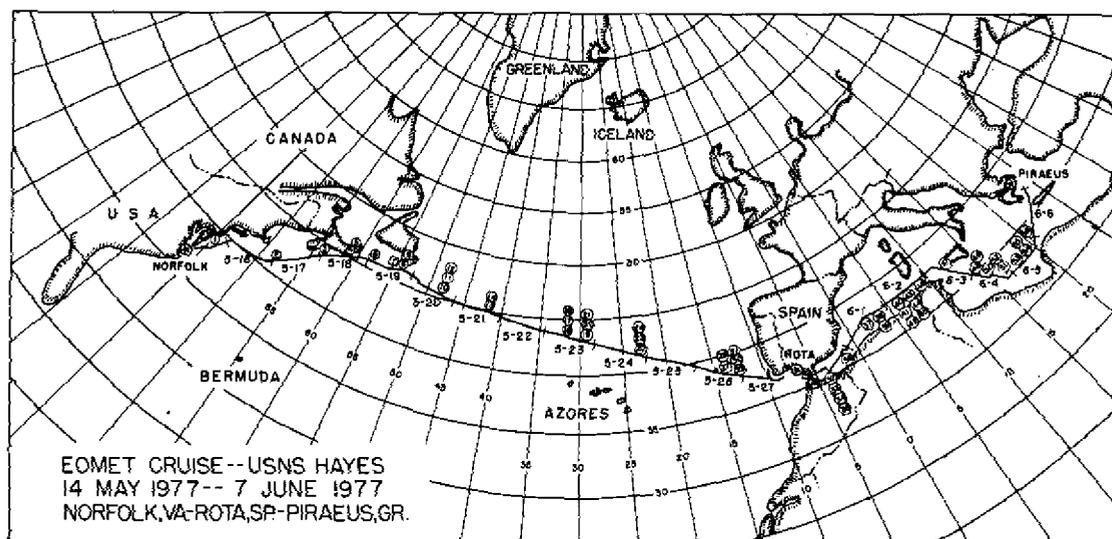


Fig. 5 — Cruise plot of the EOMET cruise of the USNS *Hayes* during May and June 1977. The circles represent positions during the track where kite-balloon profile measurements of air temperature, relative humidity, and water-vapor density were made.

which to test the model, because it does provide a wide variety of stability classes. The temperature measurements themselves are produced by dry and wet thermisters supported on the tether line to the kite balloon and ventilated by the relative wind. A solid-state pressure altimeter provides altitude information. Data are collected from these three channels and transmitted to the ship, where they are plotted in real time for the convenience of operators and digitally recorded.

The sensitivity of the system under real operational conditions allows both wet-bulb and dry-bulb temperatures to be measured to within an error band of $\pm 0.15^{\circ}\text{C}$. The absolute accuracy of the radiosonde is checked before and after every kite-balloon flight by flying the measurement package at mast height and comparing these telemetered measurements with the shipboard measurements at mast height. Thus the calibration standard for the airborne units is the shipboard Cambridge model 100SM air and dewpoint temperature instrument. The altimeter is checked before and after each flight by observations of the pressure at the mast height, which eliminates the time variations of the atmospheric pressure from the altimeter. The operational accuracy of the altimeter is ± 14 meters at the surface but decreases with altitude. One advantage of the tethered kite balloon is that the same instrument package is used many times and the calibrations of the devices can be checked before and after each flight to make sure that changes have not occurred in the launch process or during the flight itself.

Part of the problem in obtaining an evaluation of the performance of the model is to include in the evaluation some of the nonhomogeneous characteristics of the atmosphere that might be present during a particular test. Sometimes at sea continuously recorded shipboard observations diverge significantly even during the time it takes to make a profile, which is the time for the kite balloon to climb to its maximum altitude and to return to deck

level. Consequently plots of all shipboard meteorological observations were made 6 hours prior to a flight and 6 hours after landing of the apparatus to aid in determining the stability of the shipboard observations. The first part of the test was to prepare from data for the middle of each flight a list of shipboard observables such as would be entered into any ship weather report. For every experimental observation of dry-bulb temperature, wet-bulb temperature, and altitude the model also would estimate air temperature, relative humidity, and vapor density at the same altitude, using as its input the list of shipboard observables. Differences between measured quantities and the estimated quantities represent information on the accuracy of the model.

Comparisons between the model predictions and measured data have been made for all of the flights flown during the 1977 EOMET cruise. In some cases the predicted profiles of air temperature, relative humidity, and water-vapor density were close to those measured. Figures 6, 7, and 8 are examples of good correlation. They were obtained from flight 51, which took place in the Mediterranean Sea on June 4, 1978. Each point in the figures refers to a sampling of the data transmitted continuously to the shipboard receiver and digitally recorded every minute throughout the flight. The scatter in the kite-balloon data is partly due to the instrumental uncertainties in both altitude and temperature.

Figure 9 is a plot of a particularly impressive and suprising prediction of a dry layer over the Mediterranean Sea which was both measured by the boundary-layer sonde and predicted by the model. Here both the model estimates and the measurements show that a sharply defined dry layer exists above 350 meters in altitude.

